ESTIMATING GROUNDWATER RECHARGE USING THE OKLAHOMA MESONET

By

BRIANA SALLEE

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Thesis Approved:

Dr. Tyson Ochsner

Thesis Adviser

Dr. Brian Carter

Dr. Garey Fox

Name: BRIANA SALLEE

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ABSTRACT:

Groundwater supplies a large portion of water used in Oklahoma, and accurate and current information regarding groundwater recharge rates is essential for long-term, sustainable groundwater management. The Oklahoma Mesonet has provided soil moisture data at 120 monitoring stations for nearly two decades. Using these data in conjunction with site-specific soil hydraulic properties, we have estimated potential groundwater recharge, hereafter called drainage, at a depth of 60 cm for 78 Mesonet sites. Our working hypothesis is that these drainage rates are greater than or equal to the amount of groundwater recharge at each location. Calculated mean drainage rates from 1996-2012 ranged from 4 mm yr^{-1} at Hinton to 275 mm yr^{-1} at Bristow, with a state-wide median of 61 mm yr^{-1}. To clarify the relationship between these site-specific drainage rates and actual, independently-estimated groundwater recharge rates, three techniques were used. For site-specific recharge estimates, the unsaturated zone chloride mass balance method (CMB_{uz}) was applied to soil cores taken at eight Mesonet sites in western Oklahoma. Secondly, HYDRUS 1-D was used to model the effects of root water uptake below 60 cm at these eight sites. Third, to compare aquifer-wide Mesonet drainage rates to independently-estimated regional recharge rates, the saturated zone chloride mass balance method (CMB_{sz}) was applied to groundwater chloride data. CMB_{uz} analysis estimated recharge rates ranging from 0.12 mm yr^{-1} at Boise City to 2.5 mm yr^{-1} at Arnett, values much lower than Mesonet-based drainage rates. Modeled water flux at the 60-cm depth was generally higher than Mesonet-based drainage, ranging from 7.5 mm yr^{-1} at Goodwell to 145 mm yr^{-1} at Fort Cobb. Simulations also showed that a significant portion of water passing the 60-cm depth may be lost to root water uptake before reaching the 3-m depth. CMB_{sz} calculations yielded recharge rates ranging from 4.8 to 25mm yr^{-1} for five aquifers in western and central Oklahoma, giving values similar to or less than median Mesonet-based drainage rates. Overall, Mesonet-based drainage estimates seem to provide a reasonable upper limit for recharge rates at a regional scale, but may substantially overestimate recharge at sites with deep soil in western Oklahoma.

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CHAPTER I

INTRODUCTION

Access to accurate, up-to-date information regarding Oklahoma's groundwater resources is necessary in order for water resource managers, both governmental and private, to effectively plan for the future, as well as to reduce the number of groundwater-related conflicts in the state. In 2007, groundwater use in Oklahoma was approximately 800,000 acre-feet, with 73% of all irrigation water in the state coming from groundwater sources (Oklahoma Water Resources Board, 2015). One of the earliest instances of groundwater conflict in the state was the 1936 case of *Canada v. City of Shawnee*, where the City of Shawnee pumped groundwater from the Garber-Wellington aquifer (Figure 1) at such a high rate that groundwater beneath certain citizens' property was also used (Ashley and Smith, 1999). This case established the rule of "reasonable use," meaning that a land owner has a right to use as much groundwater as he pleases from his own land, even if it adversely affects surrounding citizens, so long as he has a reasonable use for that water.

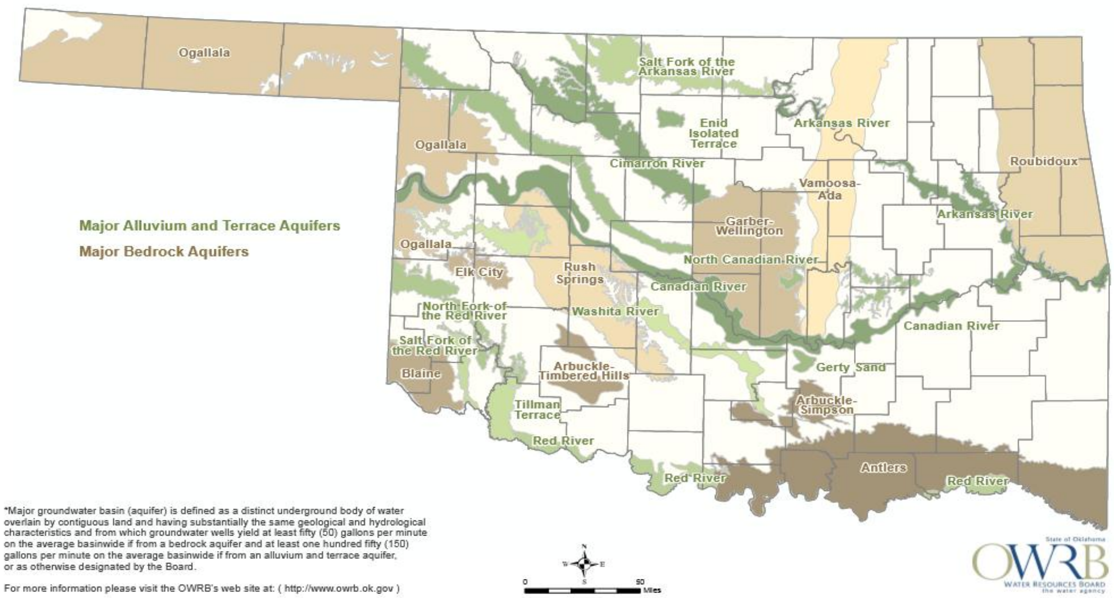

Figure 1. Major and minor bedrock aquifers of Oklahoma (Oklahoma Water Resources Board).

In 1979, the *Oklahoma Water Resources Board et al. v. Texas County Irrigation and Water Resources Association, Inc.* case was considered by the Oklahoma Supreme Court. In this case, the Texas County district court had previously granted Mobil Oil Company the right to groundwater use in the Ogallala aquifer (Figure 1), but the Oklahoma Supreme Court found that though water use for oil production could be considered a "beneficial use," landowners and agricultural producers were also in need of the groundwater (Ashley and Smith, 1999).

More recently, a study was conducted on the Arbuckle-Simpson aquifer (Figure 1) to determine the suitability of the aquifer for increased withdrawals. In 2002, the Central Oklahoma Water Authority (COWA) sought to transport water from the Arbuckle-Simpson to the center of the state. At the same time, residents of the cities of Ada and Sulphur, who depend on water from the aquifer for their drinking water, sought to protect it from outside use. It was decided that there was not enough information available to make an informed decision regarding the effects of a large increase in withdrawal of water from the aquifer, and the resulting study was an attempt to provide such information (Christenson et al., 2011). These conflicts over scarce groundwater resources are only a few of the cases that have occurred in Oklahoma, and groundwater conflicts continue today as landowners, agriculturalists, industry, and municipalities continue to struggle for their right to use this increasingly limited resource.

Increased use of groundwater resources in the state of Oklahoma has led to declining aquifer levels across the state (Figure 2), which in turn has resulted in a growing number of conflicts over groundwater withdrawal rights (Oklahoma Water Resources Board, 2012). To reduce the number of conflicts related to groundwater use, and to understand the sustainability of our state's groundwater resources, it is necessary to have accurate groundwater recharge information. Groundwater recharge is the portion of water introduced to an area, either by rainfall or by irrigation, which percolates through the unsaturated (vadose) zone and reaches the water table. Recharge rates can be highly variable and are dependent upon factors such as topography, soil type, soil texture, and climatic factors (Nolan et al., 2007).

While the depletion of groundwater reservoirs is an obvious problem to those who study and utilize these resources, the solutions to this challenge are less obvious and require the utilization of novel techniques and approaches. Currently, there is no groundwater recharge monitoring network in place in Oklahoma, and the most recently published map of state-wide recharge rates was produced in 1983 using stream baseflow data from the 1970's (Oklahoma Water Survey, 2014; Pettyjohn et al., 1983). The

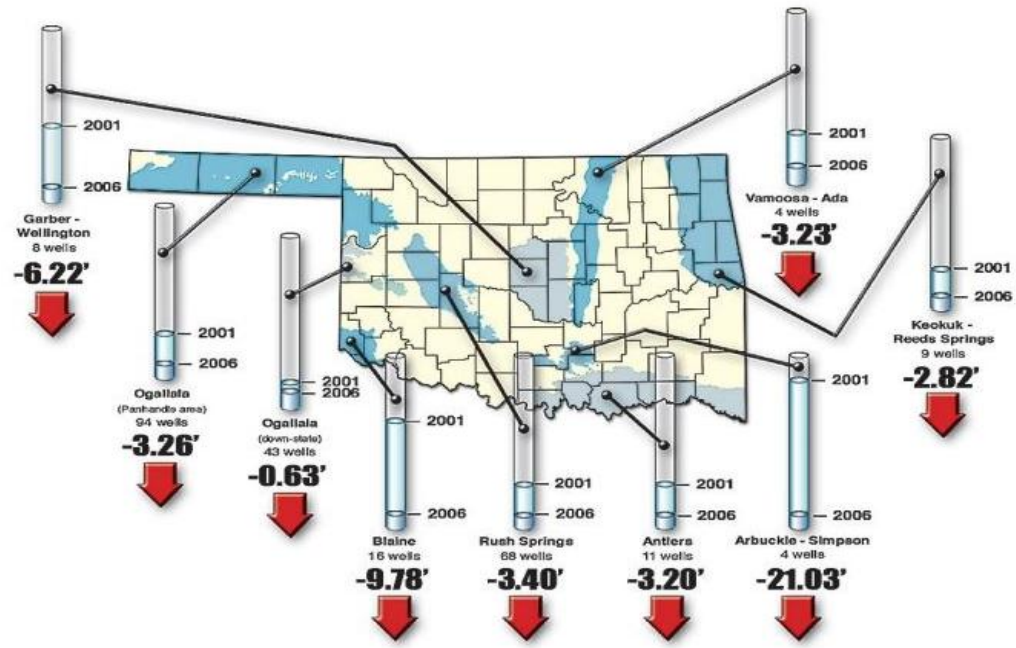

Figure 2. Decline in groundwater wells from 2001-2006, which also indicate declines in aquifer levels (Oklahoma Water Resources Board).

Oklahoma Mesonet (Figure 3) has provided data, including soil moisture data, at over 100 hydro-meteorological monitoring sites throughout the state since 1994 (soil moisture sensors were added in 1996) (McPherson et al., 2007). Soil hydraulic properties were recently determined at the 5, 25, 45, 60, and 75 cm depths for every Mesonet site by Scott et al. (2013). When soil moisture data are accompanied by these soil hydraulic property data, drainage rates from soil profiles can be calculated using a unit-gradient assumption (Nolan et al., 2007). This assumption has been widely applied in multiple research studies (Chong et al., 1981; Gardner, 1964; Sisson, 1987). Nimmo et al. (2003) stated that under this unit-gradient assumption, long-term average recharge rates can be indicated even by a single hydraulic conductivity measurement under certain conditions. In addition to site-specific recharge rates, the spatial distribution of recharge may be estimated given an adequate number of reliable recharge rates over an area (Nolan et al., 2007).

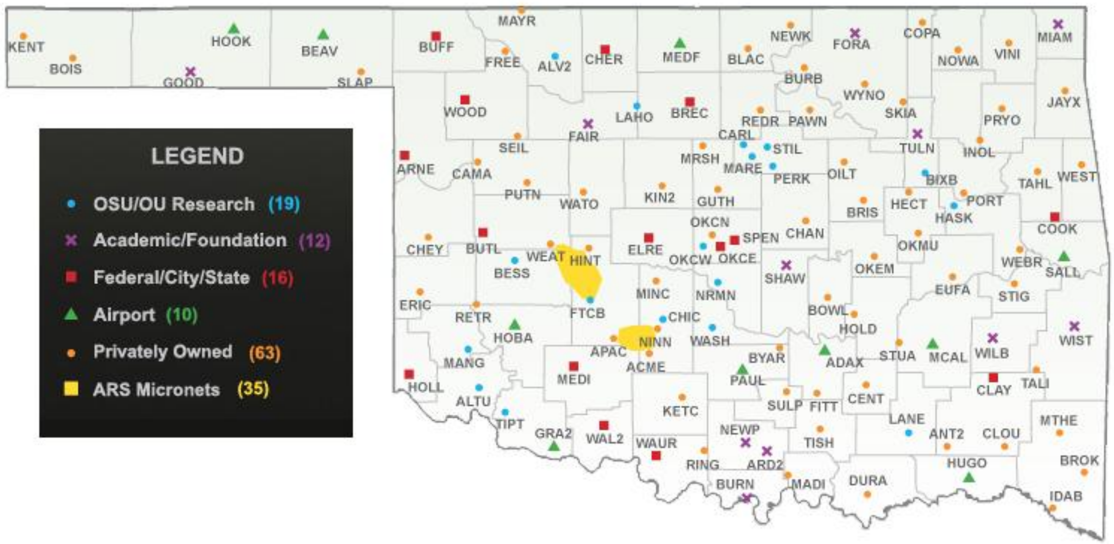

Figure 3. Oklahoma Mesonet sites labelled with four-character abbreviated name. From mesonet.org website.

Because we have nearly two decades of daily soil moisture measurements that may be used to derive unsaturated hydraulic conductivity estimates, these Mesonet-based drainage rates may provide a reliable indicator of the amount of water potentially returned to groundwater aquifers beneath the Mesonet sites. The term ‘drainage’ in this thesis refers to the amount of water percolating below the 60 cm soil depth as found by calculations using Mesonet soil moisture data. This water will eventually reach any underlying unconfined aquifer and become recharge if not taken up by roots below that depth or redirected by an impermeable layer. These assumptions may be inapplicable in some areas throughout the state, and may lead to an overestimation of recharge for certain sites. Thus, it is important to exercise caution in considering drainage rates to be equal to recharge rates at every Mesonet site and to take into account factors such as possible impermeable layers, perched water tables, and whether or not the aquifer underlying a site is confined or unconfined. Drainage rates at a Mesonet site with a confined underlying aquifer are likely not indicative of the recharge that actually occurs in that aquifer. However, these calculated drainage rates may provide land owners and policy makers helpful information related to long-term and yearly fluctuations in groundwater recharge rates throughout the state.

The overall objective of this study is to evaluate the relationship between Mesonet-based drainage rates and independent estimates of groundwater recharge across the state of Oklahoma. First, Mesonet-based drainage rates will be compared to previously published recharge rates. Additionally, Mesonet drainage rates will be compared to recharge rates estimated by the unsaturated zone chloride mass balance (CMB_{uz}) method, which was applied to soil cores taken at eight focus Mesonet sites chosen based on their deep unsaturated zones. Soil water samples from these cores were extracted and analyzed for chloride content. These chloride concentrations are indirectly proportional to recharge rates (i.e., high chloride concentrations indicate low recharge rates). HYDRUS 1-D was used to simulated water flow and root water uptake at these eight Mesonet focus sites in order to determine the effects of root water uptake beneath the 60 cm depth. The chloride mass balance method was also applied in the saturated zone (CMB_{sz}) in order to determine aquifer-scale recharge rates. Chloride concentrations in groundwater, like those in soil pore water, are indirectly proportional to recharge rates (Scanlon et al., 2010).

CHAPTER II

LITERATURE REVIEW

Significant prior research has been conducted in Oklahoma regarding the state's groundwater resources, and an understanding of previous research findings is essential when trying to develop an even greater understanding of the sustainability of our groundwater resources. Previous studies of Oklahoma's aquifers give a wide range of recharge rates, but spatial and temporal variability in soil and climatic factors makes consistent and accurate measurements of recharge rates difficult to collect. Reported recharge rates throughout the state of Oklahoma are highly variable, ranging from 0.8 to 333 mm yr^{-1} (Table 1). A significant amount of this variation is likely explained by the spatial variations in precipitation and soil type that occur throughout the state. The methods by which previous recharge estimates have been made also vary, with different information, data, and assumptions being used for each study. Additionally, there are generally only three to four studies that have been done on each aquifer within the state, and many of these studies were done in the 1960-1980s. The low number of studies completed on these aquifers, the generally small amount of data per study, the small temporal scale of most studies, and the amount of time between studies are all factors which lead to uncertainty regarding the accuracy of recharge estimates. Additionally, because of decadal-scale climate variability, many of the recharge rates found in past studies could not be considered applicable today, as variability in climatic factors has a strong effect on recharge rates (Nolan et al., 2007).

The following is a summary of previously published recharge rates for the three largest Oklahoma aquifers: the Ogallala, Rush Springs, and Garber-Wellington aquifers (Figure 1). A summary is also included for the Arbuckle-Simpson aquifer, the only sole source aquifer in Oklahoma. Additional information regarding previously published recharge rates for these and other Oklahoma aquifers is given in Table 1.

Ogallala Aquifer

The Ogallala aquifer, also called the High Plains Aquifer, spans a large area of the Midwestern United States, underlying Nebraska, Colorado, Kansas, Oklahoma, Texas, Wyoming, and South Dakota (Guru and Horne, 2000). This aquifer underlies the majority of the Oklahoma panhandle (Figure 1) and is the state's largest aquifer in terms of storage, with an in-state storage of over 90 million acre feet (Oklahoma Water Resources Board, 2015). The aquifer consists of Tertiary-Age alluvial sediments deposited by streams from the Rocky Mountains (Nativ and Smith, 1987). Saturated thickness in the Ogallala ranges from zero to nearly 430 ft (0 to 131 m) and the depth to water ranges from shallow (less than 10 ft) to deep (over 300 ft). Recharge to the aquifer is mainly from precipitation, with local high recharge rates occurring under seasonal playa lakes (Hart et al., 1976; Osterkamp and Wood, 1987). Discharge from the Ogallala ranges from

500 to 1000 gallons per minute (gpm), with certain areas reaching discharge rates up to 2,000 gpm (Oklahoma Water Resources Board, 2012).

The majority of water withdrawn from the Ogallala is used for agriculture, specifically irrigation by center pivot, with Texas County accounting for over half of all water withdrawals. In the years 1996 and 1997, 389,000 acre feet of water were withdrawn from the aquifer, while only 175,000 acre feet (45% of the amount withdrawn) were replenished by recharge (Oklahoma Water Resources Board, 2012). Groundwater level declines due to excessive pumping have been reported by many, and it is well understood that the high pumping rates common in the panhandle are in excess of what is capable of being recharged by precipitation (Guru and Horne, 2000; Gutentag et al., 1984; McGuire, 2011). In the 2012 Oklahoma Comprehensive Water Plan, the Oklahoma Water Resources Board (OWRB) stated that a continued decrease in the level of groundwater reservoirs in the majority of the state panhandle is likely if pumping rates for agricultural production do not decrease. However, this scenario was based on current groundwater use rates and may be changed if water use from this portion of the aquifer is reduced. Several studies conducted on the Oklahoma portion of the Ogallala have shown that a decrease in water extraction may lead to an increase of groundwater levels in some areas (Guru and Horne, 2000; Yang et al., 2010).

Due to the importance of groundwater from the Ogallala for agriculture and land owners, many studies have been conducted to determine the rate of recharge to the aquifer. Hart et al. (1976) used a water budget method using groundwater monitoring well data to estimate recharge and found that recharge to the Ogallala ranged from 6.4 to 25 mm yr^{-1}. Similar recharge estimates have been found in other studies regarding recharge to the Ogallala. Recharge rates ranging from 5.1 to 56 mm yr^{-1} were found by Morton (1980), who used a digital model to predict aquifer storage. Pettyjohn et al. (1983) published a state-wide recharge map that shows a recharge rate of 2.5 mm yr^{-1} for most of the Oklahoma panhandle, and a rate of 5.1 mm yr^{-1} in the portion of the Ogallala east and southeast of the panhandle. This study used a stream baseflow method to estimate recharge. This method may underestimate recharge due to the lack of consideration of evapotranspiration and possible upstream pumping from the alluvial aquifer (Scanlon et al., 2002). Recharge for the Ogallala was estimated using the MODFLOW model by Luckey and Becker (1999), who estimated that recharge ranged from 1.5 to 23 mm yr^{-1}. The OWRB published a Water Supply Availability Report in 2011 that gave a recharge range of 2.5 to 56 mm yr^{-1} (Oklahoma Water Resources Board, 2011). The most recently published recharge value for the Oklahoma portion of the Ogallala was by the Oklahoma Water Resources Board, who reported a recharge rate of 13 to 23 mm yr^{-1} in the 2012 Oklahoma Comprehensive Water Plan.

Rush Springs Aquifer

The Rush Springs aquifer, located in west-central Oklahoma (Figure 1), underlies 2400 mi^2 of land and consists of two geologic formations: the Rush Springs and Marlow formations (Runkle et al., 1997). The Rush Springs formation is made of poorly cemented Permian-age sediments, while the underlying Marlow formation consists of interbedded sandstones, siltstones, mudstones, gypsum-anhydrite, and dolomite. The Marlow formation acts as a confining unit and does not allow downward flow of water. (Oklahoma Water Resources Board, 2012). A portion of the western side of the aquifer is capped by the Chief Eagle formation, which has a very low hydraulic conductivity and likely restricts recharge in that area. The aquifer tends to produce water that has a high total dissolved solids (TDS) content. Where water quality allows, water from the aquifer is used for irrigation, domestic, and municipal water supplies, with the largest amount of water being withdrawn from the central portion of the aquifer in Caddo County. In some areas, use of groundwater from the aquifer is limited by concentrations of sulfate, nitrate, and arsenic dissolved from the parent material that exceed drinking water standards. The Rush Springs aquifer has a saturated thickness ranging from less than 200 ft in the southern portion of the formation to nearly 350 ft in the northern portion (Oklahoma Water Resources Board, 2015). Estimated groundwater withdrawal from the aquifer was estimated in 1990 to be approximately 61,272 acre feet per year, 78% of which was used for irrigation (Becker and Runkle, 1998).

Recharge to the aquifer occurs primarily from precipitation, though small portions of recharge occur due to gains from losing streams and from irrigation water applied to the land surface (Tanaka and Davis, 1963). It was estimated by Tanaka and Davis (1963) that recharge to the Rush Springs aquifer averaged 61 mm yr^{-1} for the years 1953-1956. This rate was determined by analysis of well hydrographs for four monitoring wells across the aquifer. Runkle et al. (1997) published recharge values ranging from 2.3 to 81 mm yr^{-1} that were later used as input values for their groundwater vulnerability simulation of water flow in the Rush Springs. According to Pettyjohn et al.'s (1983) state-wide recharge map, mean annual recharge for the Rush Springs ranged from approximately 5.1 mm in the northwest portion of the aquifer to approximately 25 mm in the southernmost portion. The OWRB published a recharge range of 5.1 to 89 mm yr^{-1} for the Rush Springs in 2011, while an average recharge rate of 46 mm yr^{-1} was given for the Rush Springs aquifer in the 2012 Oklahoma Comprehensive Water Plan (Oklahoma Water Resources Board, 2011; Oklahoma Water Resources Board, 2012).

Garber-Wellington

The Garber-Wellington aquifer (also known as the Central Oklahoma aquifer), located in central Oklahoma, underlies approximately 3,000 mi^2 of the state (Figure 1). Water from this aquifer is used for public, agricultural, industrial, and commercial water supplies. Except for Oklahoma City itself, every community surrounding the state capitol depends upon water from the Garber-Wellington for part of its water supply, and in some cases, all of its water supply. The aquifer is composed of Quaternary-age alluvium and terrace deposits, as well as the Permian-age Garber Sandstone, Wellington Formation, and Chase, Council Grove, and Admiral groups. Mean groundwater use from 1967-2008 was estimated to be over 37,000 acre feet, 63% of which was used for public water supplies (Mashburn et al., 2014).

Mashburn et al. (2014) compared well water levels from 1986-1987 with measurements taken in 2009, and observed a decrease in median depth-to-water measurements of 1.1 m. This same study used three different methods to estimate groundwater recharge to the Garber-Wellington. The first approach utilized the Soil-Water Balance code to estimate recharge, and was chosen because it allows for the determination of the spatial distribution of recharge over an area. This code incorporates climatic, soil, and land use data to estimate evapotranspiration, runoff, infiltration rates, and runoff. Because this method estimates potential recharge and not actual recharge, a scaling factor of 0.4 was applied to the results of the simulation during calibration. After this scaling factor was applied to the results, this method produced recharge rates ranging from 19 to 86 mm yr^{-1}, with a mean annual recharge rate of 47 mm yr^{-1}.

The second recharge estimation method applied by Mashburn et al. (2014) was the Rorabaugh method, a recession-curve displacement method (Rorabaugh, 1964). However, this method estimates recharge of the subsurface drainage basin, which was a different size than the surface drainage basin in this case. The Rorabaugh method was applied to streamflow measurements made in three locations: the Deep Fork River near Arcadia, OK, the Deep Fork near Warwick, OK, and at the Little River near Tecumseh, OK. Data for the site near Arcadia was collected from 1969-1983, with an estimated mean annual recharge rate of 104 mm yr^{-1}. Data collected near Warwick was collected only for the years 1984 and 1985 (before Arcadia Lake Dam was built), with an estimated mean annual recharge rate of 89 mm yr^{-1}. Data from the third site at Tecumseh was from the years 1943-1961, and gave a mean annual recharge rate of 33 mm yr^{-1}.

The third method of recharge estimation used by Mashburn et al. (2014) was estimation from baseflow data, with analysis being done on two sets of data: the first from the years 1987-1989 and the second from 2009 data. Data from 1987 to 1989 yielded recharge estimates ranging from 4.8 mm yr^{-1} to 81 mm yr^{-1}, while data from 2009 gave recharge rates ranging from 5.1 to 51 mm yr^{-1}.

Parkhurst et al. (1996) published recharge rates for the Garber-Wellington aquifer ranging from 4.8 to 104 mm yr^{-1}, with a median recharge rate of 41 mm yr^{-1}. This study used stream baseflow data from the years 1987-1989 to estimate recharge. Pettyjohn and Miller (1982) estimated from baseflow data that the recharge rate of the aquifer ranges from 0.8 to 211 mm yr^{-1}. A Water Supply Availability Report published by the OWRB in

2011 gave a recharge range of 28 to 41 mm yr^{-1}. The 2012 Oklahoma Comprehensive Water Plan gave a recharge value of 41 mm yr^{-1} for the aquifer.

Arbuckle-Simpson

The Arbuckle-Simpson aquifer is a karstic system located in south central Oklahoma (Figure 1), and is the only designated sole source aquifer in the state, meaning that water from the aquifer is the only source of water for a particular city. In this case, there are a number of cities which receives their water supply from the aquifer: Ada, Sulphur, Mill Creek, and Roff. The aquifer is composed of clastic and carbonate rocks and covers an area of approximately 500 mi^2. The karstic nature and geological fractures in this area allow for rapid transmission and high storage capacity for groundwater. Byrd's Mill spring, which flows from the aquifer, has been the primary water source for the nearby city of Ada, Oklahoma since 1911 (Savoka and Bergman, 1994). Generally, groundwater produced from the aquifer is chemically suitable for any use, with dissolved solids content below 400 ppm (Christenson et al., 2009). While spring flow generally meets municipal needs, three wells have been installed to allow the city to pump water during times of low precipitation. A hydrologic study was conducted in 1992, and the study showed that even high levels of pumping from the aquifer over several days had little effect on the saturated thickness of the aquifer (Savoka and Bergman, 1994). A second study conducted in 2005, which projected future groundwater levels in the aquifer based on current use rates, found similar results (Kumar, 2005).

Recharge to the Arbuckle-Simpson is generally from precipitation, which the area receives at an average rate of 1041 mm yr^{-1}, but can be from losing streams in some places (Christenson et al., 2009; Savoca and Bergman, 1994). Recharge was estimated by

Fairchild et al. (1990), using the baseflow method, to average 119 mm yr^{-1}. A more recent study using a water balance equation estimated mean annual recharge for the years 1994-2006 to be 188 mm at Connerville, a city located above the Arbuckle-Simpson (Vieux and Moreno, 2008). Additionally, this study gave a range of recharge rates from 56 to 333 mm yr^{-1}. The MODFLOW model was used to estimate recharge by Christenson et al. (2011), who found an average annual recharge range of 66 to 295 mm yr^{-1}, with the mean recharge rate being 142 mm yr^{-1}. The recharge rate published by the OWRB in their 2011 Water Supply Availability report was 119 mm yr^{-1}, while the recharge rate published in the OWRB's 2012 Comprehensive Water Plan for the Arbuckle-Simpson was 142 mm yr^{-1}.

A greater level of knowledge regarding recharge rates to key aquifers is essential if we are to sustainably manage Oklahoma's groundwater resources. The relationship between soil moisture-based drainage estimates and groundwater recharge rates is not currently known, nor has this relationship been studied in the past. An understanding of this relationship may allow for soil moisture data to be used to estimate the amount of groundwater recharge that is occurring at a location. Similarly, the use of long-term soil moisture data may be able to provide an accurate estimate of long-term average recharge, as well as provide previously unknown information regarding changes in recharge rates over time. The ability to know and study fluctuations in groundwater recharge rates using soil moisture data over long time periods, as opposed to recharge rates found by short-term studies, could provide invaluable information to water planners seeking to increase the sustainability of our state's groundwater resources.

Table 1. Summary of previously published recharge rates for the aquifers of Oklahoma. Aquifer abbreviations include AS (Arbuckle-Simpson), ATH (Arbuckle-Timbered Hills), EIT (Enid Isolated Terrace), GW (Garber-Wellington), NCR (North Canadian River), TT (Tillman Terrace), and VA (Vamoosa-Ada). Shading is used to differentiate between aquifers.

			Data Year		Recharge ($mm\ yr^{-1}$)			
Year	**Author(s)**	**Aquifer**	**First**	**Last**	**Min**	**Max**	**Mean**	**Recharge method**
2012	OWRB	Antlers			7.6	43		Computer Model
1990	Fairchild et al.	AS	1969	1979	-	-	119	Baseflow
2011	Christenson et al.	AS	2004	2008	66	305	142	MODFLOW model
2012	OWRB	AS	-	-	-	-	142	-
2008	Vieux & Moreno	AS	1994	2006	56	333	188	Water Balance
2012	OWRB	AS	-	-	-	-	119	Baseflow
2012	OWRB	ATH	-	-	7.6	15	-	Percent of Precipitation
2012	OWRB	Arkansas R.	-	-	-	-	127	-
2012	OWRB	Blaine	-	-	-	-	38	-
1986	Dugan & Peckenpaugh	Boone	-	-	-	-	254	-
1994	Imes & Emmett	Boone	-	-	267	305	-	Finite Difference Model
2012	OWRB	Boone	-	-	-	-	267	-
2012	OWRB	Canadian R.	-	-	-	-	51	-
2012	OWRB	Cimarron R.	-	-	-	-	58	-
2012	OWRB	Elk City	-	-	-	-	71	-
2011	OWRB	Elk City	-	-	51	102	-	
2012	OWRB	EIT	-	-	-	-	58	-
2014	Mashburn et al.	GW	1987	2009	19[1]	86[1]	46[1]	Soil-Water balance
1996	Parkhurst et al.	GW	1987	1989	4.8	81	-	Baseflow
2014	Mashburn et al.	GW	-	2009	5.1	51	-	Baseflow

2014	Mashburn et al.	GW	1969	1983	-	-	104	Rorabough Method
2014	Mashburn et al.	GW	1984	1985	-	-	89	Rorabough Method
2014	Mashburn et al.	GW	1943	1961	-	-	33	Rorabough Method
2014	Mashburn et al.	GW	-	1987	-	-	86	Soil-Water balance
1993	Parkhurst et al.	GW	1987	1989	4.8	103	41[2]	Baseflow
1982	Pettyjohn & Miller	GW	1973	1979	0.8	211	53	Baseflow
2012	OWRB	GW	-	-	-	-	41	-
2011	OWRB	GW	-	-	28	41	-	Baseflow
2012	OWRB	Gerty Sand	-	-	-	-	25	-
2012	OWRB	NCR	-	-	25	127	-	-
1989	Havens	NCR	-	-	43	178	84	FDM
1999	Luckey & Becker	Ogalalla	1961	1990	1.5	23	4.6	MODFLOW model
1976	Hart et al.	Ogalalla	1938	1969	6.4	25	6.4-13	Water-budget
1980	Morton	Ogalalla	-	-	5.1	56	-	Computer Model
2012	OWRB	Ogalalla	-	-	13	23	-	-
2011	OWRB	Ogallala	-	-	2.5	56	-	Computer Model
2012	OWRB	Roubidoux	-	-	-	-	64	-
2011	OWRB	Roubidoux	-	-	64	305	-	Computer Model & Percent of Precipitation
2012	OWRB	Rush Springs	-	-	-	-	46	-
1998	Becker & Runkle	Rush Springs	1989	1991	5.1	51	46	Baseflow
1963	Tanaka & Davis	Rush Springs	1953	1956	38	89	61	Well hydrograph
1983	Pettyjohn et al.	Rush Springs	1970	1979	>5.1	25	-	Baseflow
2011	OWRB	Rush Springs	-	-	5.1	89	-	Baseflow
2012	OWRB	TT	-	-	-	-	74	-
1978	Al-Sumait	TT	-	-	-	-	74	-
2012	OWRB	VA	-	-	18	36	-	-
2011	OWRB	VA	-	-	13	39	-	Baseflow

[1]Recharge values after scaling modelled values by a factor of 0.4. [2]Median recharge value, not mean.

CHAPTER III

METHODOLOGY

Soil Moisture-Based Drainage Estimates

Drainage rates were calculated using daily soil moisture data at a depth of 60 cm obtained from 78 Oklahoma Mesonet sites for the years 1996 to 2012, excluding site-years with 30 or more consecutive missing data points (Figure 4). Heat dissipation sensors (CS 229, Campbell Scientific, Logan, Utah) are installed at each of these Mesonet sites at depths of 5, 25, and 60 cm. These small sensors have a cylinder-shaped body 60 mm in length, with a ceramic matrix that makes up 32 mm of the total sensor length. This ceramic matrix, which has water-absorbing qualities similar to a silt loam soil, surrounds a hypodermic needle that contains a thermocouple (Illston et al., 2008). Also located inside the ceramic matrix is a resistance heater. The thermocouple measures and records the temperature within the sensor's porous ceramic matrix before and after a 21-s heat pulse is generated by the heater inside the needle. Data are output as the temperature difference between the initial and final thermocouple readings. These temperature differences are normalized to account for sensor-to-sensor variability, and the daily mean normalized temperature rise data were converted to matric potential using the equation given by Illston et al. (2008):

$$\Psi_m = -c \exp(a \Delta T_{ref}) \tag{1}$$

where Ψ_m is matric potential (kPa), c and a are calibration constants (0.717 kPa and 1.788°C^{-1}, respectively), and ΔT_{ref} is the reference (i.e., normalized) temperature differential recorded by the sensors (°C). These matric potential values were then converted to soil volumetric water content using the van Genuchten equation (van Genuchten, 1980):

$$\theta = \theta_r + (\theta_s - \theta_r)\left[1 + (-\alpha \Psi_m)^n\right]^{-m} \tag{2}$$

where θ is the volumetric water content, θ_s is the volumetric water content at saturation, θ_r is the residual water content, α and n are shape parameters estimated from the soil water retention curve, and m = 1- 1/n. Values for θ_r, θ_s, α, and n for each Mesonet site and each sensor depth were determined by Scott et al. (2013) using the best available sub-model within the Rosetta pedotransfer function (Schaap et al., 2001) (Tables 2 and 3).

Table 2. Rosetta input data at the 60-cm depth for eight Mesonet focus sites. Data include **s**and, silt, and clay percentages, bulk density, water contents at -33 and -1500 kPa, residual water content, and water content at saturation (Scott et al., 2013).

Site	**Sand**	**Silt**	**Clay**	**ρ_b**	**θ_{33}**	**θ_{1500}**	**θ_r**	**θ_s**
	%	%	%	g cm^{-3}	cm^3 cm^{-3}			
Arnett	39	31	31	1.43	0.205	0.111	0.055	0.406
Boise City	10	43	47	1.59	0.358	0.265	0.081	0.42
Fort Cobb	77	4	19	1.73	0.159	0.076	0.036	0.343
Freedom	18	51	32	1.31	0.214	0.137	0.078	0.431
Goodwell	39	28	33	1.53	0.214	0.144	0.078	0.388
Hooker	11	47	42	1.50	0.327	0.216	0.08	0.431
Slapout	78	4	18	1.75	0.137	0.06	0.033	0.333
Woodward	39	36	25	1.30	0.158	0.079	0.042	0.421

Table 3. Rosetta output parameters used to calculate Mesonet drainage rates at the 60-cm depth for the eight Mesonet focus sites (Scott et al., 2013)**.**

Site	**Alpha**	**N**	**K_s**	**K_0**	**L**
	kPa^{-1}	No units	cm d^{-1}	cm d^{-1}	No units
Arnett	0.332	1.36	36.2	15.6	-1.34
Boise City	0.254	1.17	2.7	5.3	-4.04
Fort Cobb	0.360	1.35	70.1	17.0	-1.45
Freedom	0.447	1.40	18.0	20.9	-1.56
Goodwell	0.042	1.364	1.5	1.2	-1.80
Hooker	0.203	1.24	6.8	5.1	-1.97
Slapout	0.401	1.37	93.4	20.5	-1.42
Woodward	0.479	1.43	39.6	36.8	-1.19

Daily volumetric water content data at the 60 cm depth were used to determine hydraulic conductivity using

$$K(S_e) = K_0 S_e^L \{1 - [1 - S_e^{n/(n-1)}]^{1-1/n}\}^2 \tag{3}$$

where effective saturation, S_e, is calculated by $S_e = (\theta-\theta_r)/(\theta_s-\theta_r)$. K_0 is a matching point conductivity (cm d^{-1}) and L is an empirical coefficient (Schaap et al., 2001). Values for K_0 and L for each Mesonet site and sensor depth were also determined by Scott et al. (2013) using Rosetta. Assuming unit-gradient conditions (i.e. gravity-driven flow), the resulting daily hydraulic conductivity was set equal to the drainage rate, or potential recharge, for that day. This unit-gradient assumption has been applied in multiple previous studies using Darcian flow methods to estimate water flux in the soil (Chong et al., 1981; Gardner, 1964; Sisson, 1987; Nolan et al., 2006).

Unsaturated zone chloride mass balance (CMB_{uz}) recharge

Independent recharge estimates were obtained for the following eight Mesonet sites in western Oklahoma: Arnett, Boise City, Fort Cobb, Freedom, Goodwell, Hooker, Slapout, and Woodward (Figure 3; site descriptions begin on page 23). These focus sites were chosen based on their location (above the Ogallala and Rush Springs aquifers), and based on the suitability of the surrounding soil for coring. Soil cores were collected in the field using either a Geoprobe (Geoprobe Systems, Salina, KS) or Giddings probe (Giddings Machine Company, Inc., Windsor, CO), with sampling depths ranging from 1 m to 8.2 m. Cores were cut into segments in the field and bagged, or collected inside a plastic core liner, to prevent evaporation. After being returned to the lab, subsamples were taken every 0.5 m or every 0.2 m from the cores collected, depending on the type of core. These subsamples were oven dried for 24 hours at 105°C to determine gravimetric water content. Subsamples (25 g) were then mixed with 50 mL of reverse osmosis water, shaken for 4 hours on a reciprocal shaker, and filtered using 0.45 µm filters (HAWP04700, Merck Millipore). Leachate was sent to Oklahoma State University's

Soil, Water, and Forage Analytical Laboratory for chloride analysis by flow injection auto-analyzer (QuikChem 8000, Lachat Instruments, Loveland, CO). The reverse osmosis water used to perform the extractions had a mean chloride concentration of 3.84 mg L^{-1}. In order to minimize the effect of this contamination, chloride results were corrected by subtracting the mean concentration of chloride in the reverse osmosis water used from CMB_{uz} results. Recharge rates (R_{CMB}) were then calculated by the mass balance equation:

$$P \; x \; Cl_P = \; R_{CMB} \; x \; Cl_{uz} \; ; \quad R_{CMB} = \; \frac{P \text{ x } Cl_P}{Cl_{uz}} \tag{4}$$

where P is precipitation, and Cl_P and Cl_{uz} are chloride concentrations in precipitation and soil water in the unsaturated zone, respectively (Scanlon et al., 2010). This steady-state equation assumes that chloride deposited by precipitation is removed from the unsaturated zone only by drainage and is thus able to be used as an indicator of percolation throughout the soil profile. Therefore, high chloride concentrations in soil water indicate areas of low drainage (and recharge), while low concentrations of chloride in soil water indicate high drainage rates.

Mean annual precipitation at each site was calculated using Mesonet rainfall records from January 1994-May 2014. Chloride concentrations in precipitation were obtained from the National Atmospheric Deposition Program, and chloride concentrations were doubled to account for dry fallout (National Atmospheric Deposition Program, 2014; Scanlon et al., 2010).

CMB_{uz} site descriptions:

Arnett: This site is located in western Ellis County, Oklahoma on an Enterprise very fine sandy loam soil (coarse-silty, mixed, superactive, thermic Typic Haplustepts) with a 3 to 5% slope. This soil is classified as being well drained, non-saline, and having a water table deeper than 80 in (>200 cm) (NRCS Web Soil Survey). The site receives an average of 558 mm yr^{-1} of rainfall, with mean annual maximum and minimum temperatures of 22° and 7.8° C, respectively.

Boise City: This site is located in central Cimarron County, Oklahoma. The dominant soil series is Sherm clay loam (fine, mixed, superactive, mesic Torrertic Paleustolls) with 0 to 1% slopes. This soil is classified as being well drained, non-saline, and having a water table deeper than 80 in (>200 cm) (NRCS Web Soil Survey). The site receives an average of 387 mm yr^{-1} of rainfall, with mean annual maximum and minimum temperatures of 21° and 4.2° C, respectively.

Fort Cobb: This site is located in Caddo County, Oklahoma. The dominant soil series is Binger fine sandy loam (fine-loamy, mixed, active, thermic Udic Rhodustalfs) with 1 to 3% slopes, comprised of loamy residuum weathered from sandstone. This soil is classified as being well drained, and has a water table depth greater than 80 in (>200 cm) (NRCS Web Soil Survey). The site receives an average of 710 mm yr^{-1} of rainfall, with mean annual maximum and minimum temperatures of 23° and 9.6° C, respectively.

Goodwell: This site is located in southern Texas County, Oklahoma. The dominant soil series is Ulysses clay loam (fine-silty, mixed, superactive, mesic Aridic Haplustolls), which is derived from calcareous loess. This soil has a maximum calcium carbonate content of 15%, is classified as being well drained, and has a water table depth greater than 80 in (>200 cm) (NRCS Web Soil Survey). The site receives an average of 414 mm yr^{-1} of rainfall, with mean annual maximum and minimum temperatures of 22° and 5.7° C, respectively.

Hooker: This site is located in north Texas County, Oklahoma. The dominant soil series is Gruver clay loam (fine, mixed, superactive, mesic Aridic Paleustolls), which is derived from calcareous loamy and clay aeolian deposits. This soil has a maximum carbonate content of 35%, is classified as being well drained, and has a water table depth greater than 80 in (>200 cm) (NRCS Web Soil Survey). The site receives an average of 432 mm yr^{-1} of rainfall, with mean annual maximum and minimum temperatures of 22° and 5.7° C, respectively. This Mesonet station is located in a drainage ditch (Figure 4), where pooling of water during and after rain events may lead to unrealistically high and non-representative calculated drainage rates.

Figure 4. Panoramic photo of the Hooker Mesonet site, which is located near a drainage ditch. Photo from mesonet.org website.

Slapout: This site is located in south Beaver County, Oklahoma. The dominant soil series is a Mansic-Mobeetie complex (fine-loamy, mixed, superactive, thermic Aridic Calciustolls), which is derived from calcareous loamy alluvium. This soil has a maximum carbonate content of 40%, is classified as being well drained, and has a water table depth greater than 80 in (>200 cm) (NRCS Web Soil Survey). The site receives an average of 521 mm yr^{-1} of rainfall, with mean annual maximum and minimum temperatures of 22° and 7.0° C, respectively. Due to a thick caliche (carbonate) layer adjacent to the Mesonet site, only one sample was taken near the site and is hereafter called the 'first' core sample taken at Slapout. The second core sample taken at this location was taken approximately 150 m from the Mesonet station at a higher elevation. This core is referred to as the 'second' core for this site.

Woodward: This site is located in south Woodward County, Oklahoma. The dominant soil series is an eroded Carey silt loam (fine-silty, mixed, superactive, thermic Typic Argiustolls), which is comprised of sandy and silty residuum weathered from sandstone and siltstone. This soil is non-saline, has a maximum carbonate content of 20%, a maximum sodium absorption ratio (SAR) of 4.0, is classified as being well drained, and has a water table depth greater than 80 in (>200 cm) (NRCS Web Soil Survey). The site receives an average of 618 mm yr^{-1} of rainfall, with mean annual maximum and minimum temperatures of 22° and 8.5° C, respectively.

Modeled recharge rates

Modelled recharge rates were obtained for these same eight sites using HYDRUS 1-D (PC-Progress, Prague, Czech Republic), which solves the Richards equation for saturated-unsaturated water flow. The main processes evaluated in the model were water flow and root water uptake in the unsaturated zone. Upper boundary conditions were defined as atmospheric conditions with surface runoff, meaning that precipitation rates in excess of soil hydraulic conductivity did not increase the pressure head at the upper boundary. Upper boundary condition inputs included daily precipitation values for the years 1994-2014, potential evaporation, and potential transpiration values.

The FAO Penman-Monteith method (also referred to as the FAO-56 method) was used to estimate potential evapotranspiration at each site. The FAO-56 method uses daily maximum and minimum air temperature, maximum and minimum relative humidity, precipitation, wind speed, incoming solar radiation, and average vapor pressure deficit data to estimate potential evapotranspiration (Allen et al., 1998). All climatic input data were collected from the Oklahoma Mesonet for the years 1994-2014. The HYDRUS model requires the partitioning of potential evapotranspiration input data into its potential evaporation and potential transpiration components, which can be done in a number of ways. However, each of these methods relies on a quantitative measurement of the leaf area index (LAI) at the studied site. Because information regarding the time-varying leaf area index at the Mesonet sites is unavailable, sensitivity analysis was done to determine the effect of varying evaporation and transpiration partitioning within the model. This was done for the Arnett site by considering evaporation and transpiration to be varying percentages of the total estimated potential evapotranspiration (e.g., Run 1: evaporation = 20% of ET_o, transpiration = 80% of $ET_{o;}$ Run 2: evaporation = 25% of ET_o, transpiration = 75% of ET_o, etc.). The HYDRUS model was then run and the resulting flux rates at 60 cm and 3 m were recorded and compared. This sensitivity analysis was conducted only for the simulation of the Arnett site because it is likely that other sites would show similar results to the same type of analysis.

Additional upper boundary inputs included a factor termed hCritA, which is the minimum allowed pressure head at the soil surface. This value only comes into effect when the pressure head at the soil surface is less than the defined hCritA value. When the pressure head at the surface drops below the hCritA value, evaporation is decreased from the given potential evaporation rate. An hCritA value of -100,000 cm was used in our simulations, with the exception of less than 30 days at the Fort Cobb site, where the hCritA value was reduced to -10,000 cm in order to allow for the convergence of the model. This value was within the range of values suggested in the HYDRUS 1-D documentation.

The van Genuchten-Mualem single porosity model was used to model water flow throughout the soil profile, with an air entry suction value of -2 cm. Soil profiles at each site were characterized by either 4 or 5 soil layers to a depth of 3 m. This depth was chosen because it is the same depth used for soil core collected, and because it is unlikely that the root zone extends to 3 m. Soil layer depths, residual water content (θ_r), saturated water content (θ_s), saturated hydraulic conductivity (K_s), α, n, and L values for each site and depth were taken from Scott et al. (2013). Lower boundary conditions were defined as free drainage at 3 m.

Root water uptake was simulated using the model of Feddes et al. (1978). Critical pressure head values in the water stress response function were determined by values for pasture from Wesseling (1991), with the exception of the wilting point, which was adjusted to -1500 kPa. Root density values were estimated using values published by Sun et al. (1997), who used detailed tracings of grass roots from the central grasslands of North America to estimate densities among various types of root systems. Because these root density values were found using images of plant roots and not actual soil profiles, root densities were reported as cm root/cm soil (in depth). For our purposes, the assumption that reported root densities are uniform in three-dimensional space was made. This approach was considered appropriate because HYDRUS normalizes the potential root water uptake function before each simulation.

Sensitivity analysis was conducted for the simulation at the Arnett site to determine the effects of different rooting depths and density distributions on water flux to the 3-m depth. This sensitivity analysis was done using root distributions and maximum rooting depths for grasses given by Sun et al. (1997). To determine the effect of shallow root systems on water uptake, the HYDRUS model was run using the root density distribution and maximum rooting depth of shallow-rooted grasses species. To determine the effects of deeper rooting systems on water uptake, the model was run using the root density distribution of medium-rooted grasses.

Because rooting types were not known for the simulated sites, the mean root densities of shallow and medium-rooted grasses were used for modeling at all sites (Sun et al., 2007). The maximum rooting depth at all sites was considered to be 163 cm, which is the mean maximum rooting depth for grasses given by Sun et al. (1997). This approach is reasonable given that in most grasslands the greatest root densities are seen at depths of less than 20 cm (Sun et al., 1997). Additionally, a global analysis of root zone distributions done by Jackson et al. (1996) shows that 83% of all roots in grasses are located within the top 30 cm of soil.

HYDRUS outputs included, along with many other factors, estimations of pressure head, water content, and water flow at 10 observation nodes placed at 30-cm intervals throughout each profile with the exception of the 270-cm depth. This depth was excluded because HYDRUS only allows the use of 10 observation nodes. Water flow estimates were imported into MATLAB, where analysis was done to determine mean annual flux, total annual flux, and long-term average flux at each depth where an observation node was located. The first two years of simulation, 1998 and 1999, were considered a "warm up" period and thus were excluded from these calculations.

Regional-scale recharge rates

Regional recharge rates were estimated by applying the saturated zone chloride mass balance (CMB_{sz}) method to groundwater chloride data from the Oklahoma Water Resources Board's Groundwater Monitoring and Assessment Program (GMAP). For the 2013 portion of this program, samples were collected from 156 groundwater monitoring wells located across five Oklahoma aquifers: Ogallala, Rush Springs, Gerty Sand, Elk City, and Canadian River. Samples were analyzed to determine the concentration of certain chemicals present, with a detection limit of 10 mg L^{-1}. Concentrations below detection limit were reported as being zero, but for our purposes we assumed an average concentration of 5 mg L^{-1} for all samples below this detection limit. Reported chloride concentrations in these samples were used to calculate recharge rates, again following methods by Scanlon et al. (2010), according to equation 4, except that the chloride concentration in groundwater (Cl_{gw}), rather than the chloride concentration from soil, is the term in the denominator. Because chloride may also be derived by upward migration from underlying saline aquifers, $Cl:SO_4$ ratios greater than one were used to distinguish chlorine from precipitation from chlorine derived from upward flow and were excluded from subsequent analysis (Scanlon et al., 2010).

CHAPTER IV

RESULTS & DISCUSSION

Soil Moisture Based Drainage Estimates

Calculated mean drainage rates across the state ranged from 4 mm yr^{-1} at the Hinton, OK Mesonet site to 275 mm yr^{-1} at Bristow (Figure 5). The state-wide median Mesonet-based drainage rate was found to be 61 mm yr^{-1}, which is approximately 7% of the median state-wide rainfall. A similar percentage was found by Kim and Jackson (2011), who observed that 8% of rainfall became recharge under grassland systems in their global analysis. Generally, Mesonet drainage rates decrease as you move from the southeastern portion of Oklahoma to the panhandle, following the precipitation gradient of the state (Figure 5). This state-wide drainage map agrees somewhat with a similar map published by Pettyjohn et al. (1983), in that drainage follows the pattern described above. However, there are some differences between the two maps. For instance, our calculated drainage rates in the panhandle range from 7 to 30 mm yr^{-1} and are slightly higher than recharge rate of 2.5 mm yr^{-1} estimated for the panhandle by Pettyjohn et al (1983). Additionally, a lack of data in the southeastern part of the state makes a comparison between the two maps difficult for that area.

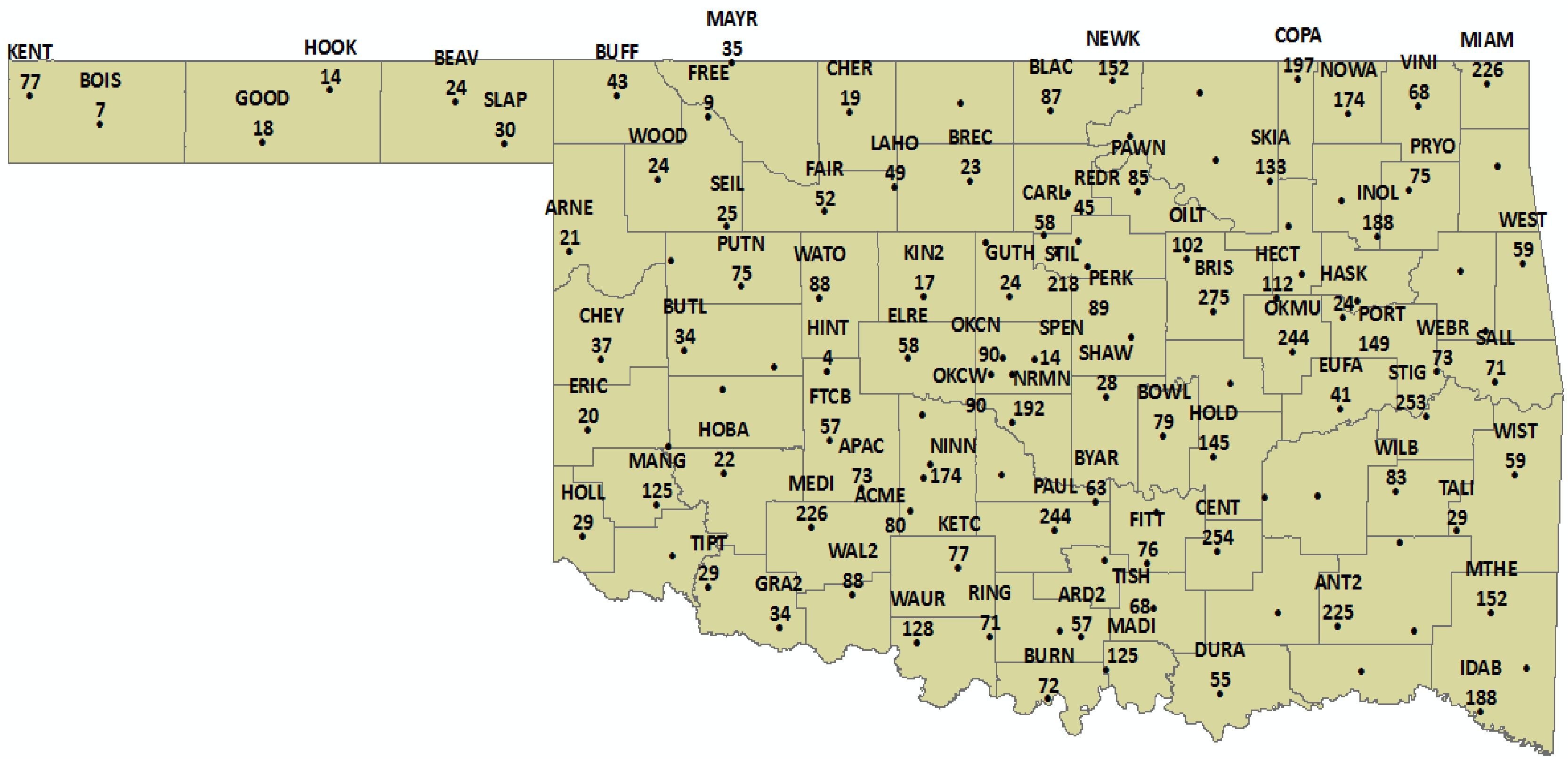

Figure 5. Calculated mean annual drainage rates (mm yr^{-1}) and four-character site ID codes at Mesonet sites across Oklahoma for the years 1996-2012. For sites where no information is given, the corresponding Mesonet station did not have adequate soil moisture data to yield a reliable drainage rate.

Quality control was done using a box plot in order to determine drainage rate outliers (Figure 6), and resulted in the exclusion of the Washington, Bixby, Lane, and Foraker sites from subsequent analyses. These sites had the four highest calculated drainage rates, which were 654, 383, 372, and 356 mm yr^{-1}, respectively.

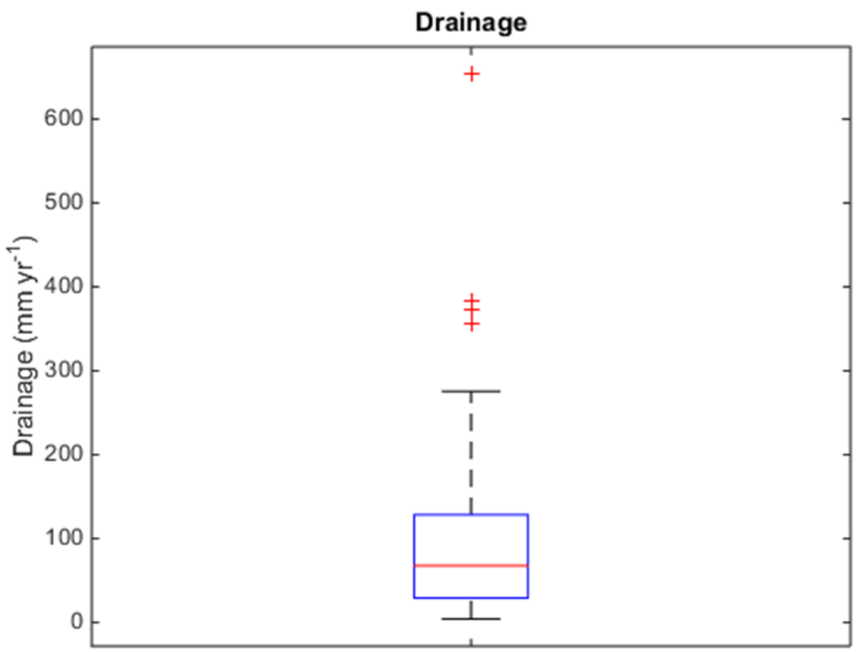

Figure 6. Boxplot used to determine outliers among mean calculated drainage rates for the years 1996-2012. The line in the center of the box is the median drainage rate, the lower and upper edges of the box are the 25th and 75th percentiles, and the whiskers extend to the most extreme data points not considered outliers. The four red crosses correspond to the four highest calculated drainage rates, which are considered outliers. These occurred at the Washington, Bixby, Lane, and Foraker sites.

A reasonable explanation was found for the high drainage rates each of these sites. At the Washington site, shallow bedrock (59-72 in) may prevent the downward flow of water, leading to non-representative high soil moisture values at that site (NRCS Web Soil Survey). This explanation is supported somewhat by soil property data reported by Scott et al. (2013) for this site, which shows a high bulk density of 1.95 g cm^3 at the 55 cm sensor depth. At Bixby it was determined that soil from nearby agricultural fields has blown onto the Mesonet site, adding up to 10 cm of loose soil to the soil surface and thus creating incorrect soil moisture sensor reading depths. Additionally, it is thought that irrigation water from those same agricultural fields may reach the Mesonet site, also creating unrealistic soil moisture values. The Lane site is located on eroded soils, and Scott et al. (2013) reported both an increase in bulk density and a sharp decrease in the saturated hydraulic conductivity at the 70-cm depth, indicating that there may be a limiting layer at that depth which restricts downward water movement. At Foraker, shallow bedrock (20-40 in) likely creates a shallow saturated zone, leading to high soil moisture values and drainage rates (NRCS Web Soil Survey).

Even after removing statistical outliers, some sites, such as the Mangum, Hinton, Norman, Bristow, Stigler, and Paul's Valley stations, have a calculated drainage rate that is noticeably higher than the surrounding sites. At the Mangum site, very sandy soils (>85% sand) and a shallow, possibly perched water table (20-40 in) may lead to overestimations of drainage (NRCS Web Soil Survey; Scott et al., 2013). At Hinton, despite having sandy soils (>75% sand), the soil is characterized as having a high runoff class and a very low ability to store water in the profile (NRCS Web Soil Survey; Scott et al., 2013). At the Norman site, bulk density increases at the 55 cm sensor depth, resulting in a significantly lower saturated hydraulic conductivity than the soil above that depth (Scott et al., 2013). This could lead to water flow being restricted at the 55 cm depth and an overestimation of drainage. The reason for high drainage rates at the Bristow site is difficult to pinpoint. Irrigation may be occurring near the site, but here are not any obvious problems that would cause an incorrect drainage rate calculation. The Stigler Mesonet site is located very close to the floodplain of a creek. The location of the site may lead to non-representative high soil moisture readings at certain times, thus leading to an overestimation of drainage. The Paul's Valley site is located near an area with several homes and paved roads, and is bounded on the other side by a small creek. It may be possible that the soil moisture sensors at this site are being influenced by increased runoff from these roads and by ephemeral stream flow. Additional quality control measures likely need to be taken to ensure reliable, representative soil moisture readings at all sites across the state. For example, the Hooker site, though it does not seem to have an unusually high mean drainage rate, is located in a drainage ditch (Figure 5). It is probable that this site is not the only one that is being impacted by poor site location leading to non-representative drainage rates, and a greater knowledge of the sites affected by poor station placement could be helpful in determining which sites' drainage rates are realistic.

The median Mesonet-based drainage rate from the eight focus sites was found to be 19 mm yr^{-1} (Table 4). The percentage of long-term mean precipitation that becomes drainage ranged from 1 to 8% at these eight Mesonet sites, with a mean value of 4% (Table 4). This mean value is lower than the estimate of 8% made by Kim and Jackson (2011), but is not unreasonable. Wood and Sanford (1995) estimated recharge to be only 2% of precipitation in the Southern High Plains of Texas and New Mexico, both of which are very near the sites located in the Oklahoma Panhandle. These sites are all located in the western portion of the state, where high temperatures coupled with high wind speeds may increase evaporation from the soil as well as transpiration by plants, reducing the amount of precipitation that is able to percolate to the deeper soil depths. It is worth noting, however, that six of the seven sites overlying the Ogallala aquifer (Arnett, Boise City, Goodwell, Hooker, Goodwell, and Woodward) have Mesonet-based drainage rates that fall within the 5 to 56 mm yr^{-1} range of recharge rates given by previous studies (Table 4). The Freedom site, located above the Cimarron River alluvial aquifer, has a calculated drainage rate of 9.3 mm yr^{-1}, which is significantly lower than the 58 mm yr^{-1} recharge rate given for that aquifer in the 2012 Oklahoma Comprehensive Water Plan. The 57 mm yr^{-1} drainage rate at the Fort Cobb site compares fairly well with previous recharge estimates that range from 5 to 264 mm yr^{-1} for the Rush Springs aquifer.

Table 4. Site name, mean annual precipitation, mean annual Mesonet drainage rate, and the ratio of drainage to precipitation.

Mesonet Site	Precipitation	Drainage	D/P
	mm yr^{-1}	mm yr^{-1}	
Arnett	561	21	0.04
Boise City	386	7.3	0.02
Fort Cobb	712	57	0.08
Freedom	655	9.3	0.01
Goodwell	410	18	0.04
Hooker	436	14	0.03
Slapout	530	30	0.06
Woodward	630	24	0.04

Yearly calculated drainage rates vary quite a bit by year, and generally follow the pattern of annual precipitation at some sites, such as at the Woodward site (Figure 7). This figure shows the variability that may be seen in drainage by year, and shows that drainage may not fluctuate according to changes in yearly precipitation alone. Regression analysis for this site showed no observable correlation between yearly precipitation and yearly drainage rates. This is likely because there are a number of other factors that influence drainage rates, such a temperature and plant cover type and amount. There is a large decrease in drainage from the year 1999 to the year 2000, despite an increase in annual precipitation. The reason for this is not clear, and this drastic decrease in drainage seems to mark the beginning of a trend that lasts over a decade. It is possible that water that was previously reaching the site (perhaps as runoff) was redirected beginning in the year 2000, or that an increase or change in vegetation led to a decrease in drainage. It is interesting to note the increase from nearly zero drainage in 2006 (the year with the lowest annual precipitation and drainage rates of the 2000-2010 decade) to approximately 35 mm yr^{-1} of drainage in 2007 (the year with the highest annual drainage rate and second highest annual precipitation level of the 2000-2010 decade).

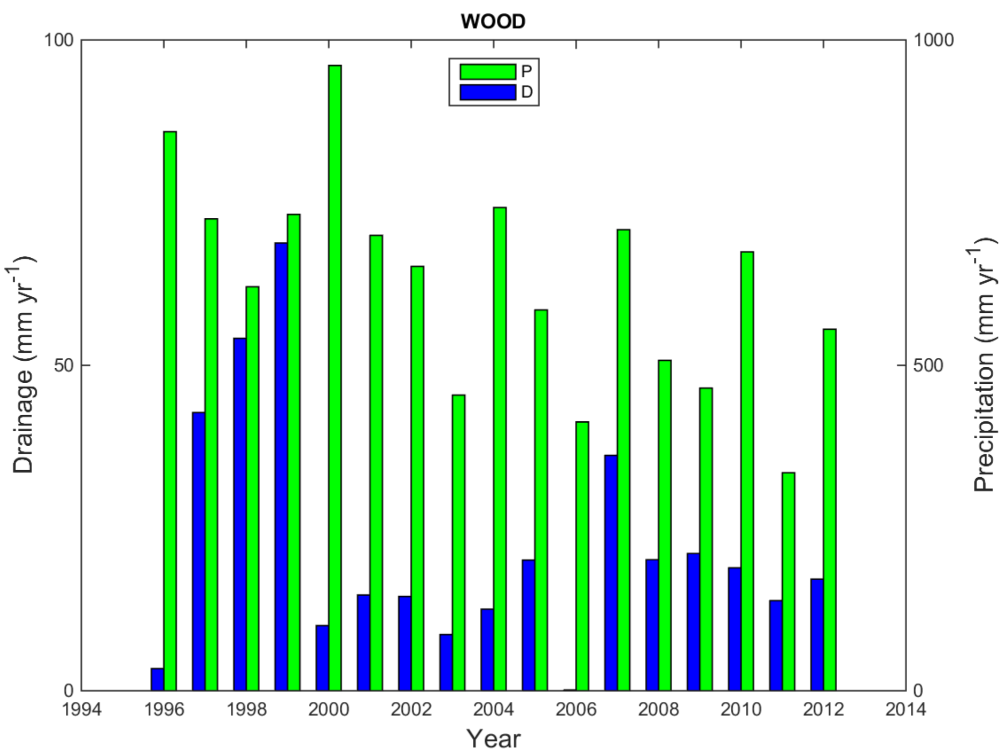

Figure 7. Total precipitation and drainage at the 60-cm depth by year from 1996-2012 at the Woodward Mesonet site.

Unsaturated zone chloride mass balance (CMB_{uz}) recharge

Measured mean chloride concentrations found by CMB_{uz} analysis of soil cores taken at the eight focus sites ranged from 39 mg L^{-1} at the Goodwell site to 1529 mg L^{-1} at the Freedom site (Table 5). These chloride concentrations led to estimated recharge rates ranging from 0.12 mm yr^{-1} at Boise City to 2.5 mm yr^{-1} at the Arnett site. CMB_{uz} analysis of soil cores taken at the eight focus sites gave a median recharge rate of 0.3 mm yr^{-1}, a significantly lower value than the 19 mm yr^{-1} median estimate made from Mesonet soil moisture data. The reasons for this discrepancy between Mesonet drainage rates and CMB_{uz} recharge values are not clear. One possibility is that the unit-gradient assumption made in our analysis of Mesonet drainage rates does not reflect actual conditions at these sites. Another possibility is that chloride inputs not reflected in the NADP data could have affected chloride concentrations in the soil, leading to an underestimation of recharge from the CMB_{uz} analyses. A third possibility is that significant root water uptake may occur beneath the 60 cm depth such that the Mesonet-based drainage rates overestimate recharge.

Previous studies have shown chloride concentrations with a similar range of chloride concentrations. Scanlon et al. (2010) reported chloride concentrations ranging from 6.9 to 1600 mg L^{-1} for soils under rain-fed systems in the Texas High Plains. However, this paper reports recharge rates only for medium-course grained rain-fed systems, whose chloride concentrations have a much smaller range of 6.9 to 83 mg L^{-1}. The resulting recharge rates range from 2.8 to 36 mm yr^{-1}, with a median recharge rate of 9.3 mm yr^{-1}. This value is two orders of magnitude higher than the median CMB_{uz}-estimated recharge rate (0.3 mm yr^{-1}) for our focus sites. One reason for this difference may be the difference in the concentration of chloride in precipitation between studies, which is often considered to be the factor with the greatest level uncertainty (Wood and Sanford, 1995). Scanlon et al. (2010) report precipitation chloride concentrations of 0.22 to 0.26 mg L^{-1}, which would lead to slightly higher recharge rates than the 0.10 to 0.19 mg L^{-1} concentrations used in our study.

Table 5. Site name, mean annual precipitation, mean chloride concentrations in precipitation, depth-weighted mean chloride concentration beneath the root zone from soil cores, and CMB_{uz} recharge estimates for the eight Mesonet focus sites.

Mesonet Site	Precipitation	Cl_P	Chloride	Recharge
	mm yr^{-1}	mg L^{-1}	mg L^{-1}	mm yr^{-1}
Arnett	561	0.19	81	2.5
Boise City	386	0.1	624	0.12
Fort Cobb	712	0.19	961	0.27
Freedom	655	0.19	1529	0.16
Goodwell	410	0.1	39	2.0
Hooker	436	0.1	687	0.13
Slapout	530	0.1	516/76	0.20/1.4
Woodward	630	0.19	127	1.8

A study done by McMahon et al. (2003) in Southwestern Kansas reported chloride concentrations from one soil sample ranging from 13 to 253 mg L^{-1}. These chloride concentrations are comparable to many of the mean chloride concentrations found at our eight focus sites. The mean precipitation reported in this previous study was 452 mm yr^{-1}, which is also comparable to the rates of precipitation at many of our study sites. Calculated recharge rates for this study were found to range from 2.4 to 10 mm yr^{-1}, with a mean value of 5.2 mm yr^{-1}. These numbers are higher than our estimated recharge rates. This is likely because, again, the estimated chloride concentrations in precipitation were higher for the McMahon et al. (2003) study than for our sites. It may be possible that low reported chloride concentrations in precipitation in Oklahoma are leading to recharge rates lower than previous studies have observed in other regions.

The chloride profile for the Goodwell site (Figure 8) shows a chloride peak near the 1-m depth. This figure shows how chloride is typically inversely related to the amount of water at that point in the soil profile (i.e., high chloride concentration, low water content and low recharge). Chloride concentrations beneath the 1.5-m depth stabilized near a concentration of approximately 40 mg L^{-1}, which suggests that chloride beneath that depth is not being collected, but rather is being flushed through the profile by moving soil water. Also, this profile demonstrates how concentrations in soil pore water are capable of varying considerably by depth.

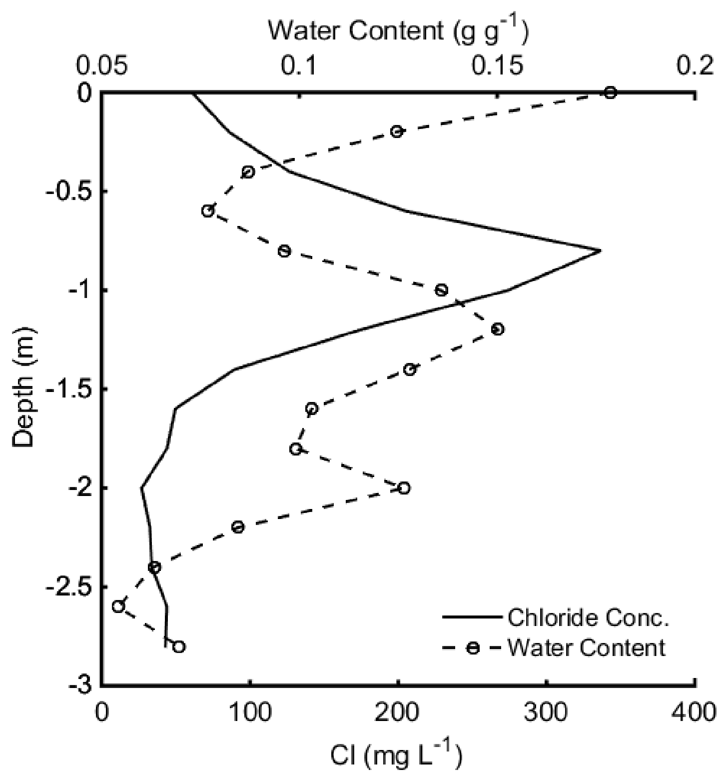

Figure 8. Chloride concentration and gravimetric water content versus depth for the soil cores collected at Goodwell, Oklahoma.

Modeled recharge rates

The results of the root density and root distribution sensitivity analysis done in HYDRUS 1-D showed that the amount of water the reaches the 3-m depth did not vary between root densities and maximum rooting depths (Table 5). However, there was a significant effect on the amount of water that passes the 60 cm depth. Shallow root systems showed a conductivity of 41 mm yr^{-1} at the 60 cm depth, while medium root systems allowed 68 mm yr^{-1} at the same depth. These results are the opposite of what one would expect to see. Normally, a deeper root system would lead to a decrease in the amount of water reaching deeper soil depths. However, because the root density of shallow-rooted plants is highest near the surface of the soil, where the greatest density of roots occurs at the 30- 60 cm depth for medium-rooted plants, it may be that high rates of root water uptake near the surface in shallow-rooted plants are restricting water flow to deeper depths. These results show that the different root densities and depths simulated have little effect on the amount of water reaching the 3-m depth, but do significantly affect the amount of water that flows past the 60 cm depth.

Table 6. Results of rooting depth and root density sensitivity analysis done in HYDRUS for the Arnett Mesonet site. Plant root types, root densities by depth, and maximum rooting depths (RD) from Sun et al. (2007), and the resulting HYDRUS flux rates at 60 and 300 cm.

Root type	0-30 cm	30-60 cm	60-100 cm	100+ cm	Max. RD	Flux	Flux
	cm cm^{-1}				cm	mm yr^{-1}	
Shallow	43	24	12	5	128	41	0.20
Medium	25	34	32	13	207	68	0.20
Average	34	29	22	9	163	54	0.20

The results of potential evapotranspiration partitioning sensitivity analysis show that varying the amounts of potential evapotranspiration partitioned to the evaporation and transpiration components has a moderate effect on the amount of water that reaches the 60 cm depth, but no effect on the amount of water that reaches the 3-m depth (Table 6). Increasing the evaporative component led to lower flux rates at the 60 cm depth, but was not found to affect the amount of water that percolates to the 3-m depth. Because we believe that evaporation at the modelled sites may be relatively high and that the sparse vegetation of the panhandle limits the amount of transpiration that may occur in this region, 50% of potential evaporation was allotted to both the evaporation and transpiration components.

Table 7. Results of potential evapotranspiration partitioning sensitivity analysis done in HYDRUS for the Arnett Mesonet site. Evaporation and transpiration (fractions), the resulting flux rates at 60 cm and 300 cm, and the amount of water from the 60 cm depth that reaches the 3-m depth.

E	**T**	**Flux 60 cm**	**Flux 300 cm**	**300:60**
		mm yr^{-1}	mm yr^{-1}	
0.2	0.8	68.0	0.2	0.002
0.25	0.75	64.7	0.2	0.002
0.4	0.6	57.5	0.2	0.003
0.5	0.5	54.4	0.2	0.004
0.6	0.4	52.5	0.2	0.005
0.75	0.25	51.4	0.4	0.008
0.8	0.2	50.8	0.6	0.012

The Mesonet-based drainage estimates from the 60 cm soil moisture sensor depth were evaluated using HYDRUS 1-D (Table 7). Water fluxes at the 60 cm depth were estimated by the model to range from 7.5 mm yr^{-1} at Goodwell to 145 mm yr^{-1} at Fort

Cobb. For all but two sites, Hooker and Goodwell, the HYDRUS-estimated flux at 60 cm was higher than Mesonet-based drainage estimates at the 60 cm depth. Interestingly, water contents beneath the root zone from soil samples taken in the field are generally higher than those simulated by HYDRUS, which would indicate that the model-generated water flow and root uptake values may not reflect actual conditions at these sites (Table 8). However, these HYDRUS results compare fairly well with previous studies conducted on the aquifers underlying the sites. All but the Freedom and Fort Cobb sites lie above the Ogallala aquifer, for which previous studies have estimated a recharge range of 5 to 56 mm yr^{-1} (Table 1). Of the seven sites located over the Ogallala, only the Slapout site has a HYDRUS-estimated recharge rate that falls outside this range, with a 60-cm flux rate of 90 mm yr^{-1}.

HYDRUS-estimated flux rates at the 3-m depth ranged from 0.1 mm yr^{-1} at Goodwell to 21 mm yr^{-1} at Fort Cobb, and are substantially lower than both HYDRUS-estimated and Mesonet-based drainage rates at all sites. These flux rates at the 3-m depth are similar to those found by Wang et al. (2009), who used HYDRUS 1-D to simulate recharge on sandy soils in Nebraska. This study modeled soil profiles to a depth of 5 m and used a unit gradient assumption. Recharge rates for sand and loamy sand were estimated to range from 0.06 to 32.8 cm yr^{-1} and 0.02 to 25.7 cm yr^{-1}, respectively. These results show the high level of variability of recharge estimates, even within a single study and location. Additionally, these results give evidence that our HYDRUS-estimated 3-m flux rates for Oklahoma, which range from 0.1 to 21 mm yr^{-1}, are likely acceptable. Though our HYDRUS-estimated flux rates, with the exception of the Fort Cobb site, are on the low end of the range found by Wang et al. (2009), this is to be expected when it is considered that the Nebraska study site received an average of 57.6 inches of rain per year, while the simulated sites in Oklahoma receive a median value of only 21.5 in yr^{-1}.

Table 8. Mean annual Mesonet-based drainage at 60 cm and mean annual flux values at 60 and 300 cm found using HYDRUS 1-D.

Mesonet Site	**Mesonet Drainage 60 cm**	**HYDRUS flux 60 cm**	**HYDRUS flux 300 cm**
	mm yr^{-1}	mm yr^{-1}	mm yr^{-1}
Arnett	21	54	0.2
Boise City	7.3	15	0.3
Fort Cobb	57	145	21
Freedom	9.3	64	0.8
Goodwell	18	7.5	0.1
Hooker	14	9.9	1.3
Slapout	30	90	0.5
Woodward	24	55	0.7

Table 9. Mean volumetric water content beneath the root zone (beneath 163 cm) for soil cores taken in the field and for HYDRUS simulations. Actual water content beneath the root zone is not available for the Woodward site because a sample was not able to be collected beneath the root zone.

Site	**θ_v actual**	**θ_v simulated**
	cm^3 cm^{-3}	cm^3 cm^{-3}
Arnett	0.15	0.11
Boise City	0.14	0.14
Fort Cobb	0.12	0.08
Freedom	0.20	0.13
Goodwell	0.13	0.09
Hooker	0.22	0.33
Slapout	0.03/0.33	0.10
Woodward	NA	0.09

Similar results were also found by Keese et al. (2005), who used UNSAT-H (a model very similar to HYDRUS) to estimate recharge in Texas. This model, like HYDRUS, solves the Richards equation for variable saturated soils and applies the Feddes et al. (1978) model for plant water uptake. Simulations were done using 30-year climatic data from the years 1961-1990. The resulting recharge estimates for vegetated, variable-textured soils were found to range from 0.2 to 117.7 mm yr^{-1}, or 1 to 10% of annual precipitation. These results compare favorably with our modelling results, especially with respect to the percentage of annual precipitation that becomes recharge. Again, our estimates fall toward the low end of the range given by Keese et al. (2005), but this is likely because areas with precipitation of up to 1783 mm yr^{-1} were considered in their study, where our greatest modelled mean annual precipitation rate was only 712 mm yr^{-1}.

Overall, our modelling results suggest that root water uptake may be significant below 60 cm for these sites and that Mesonet drainage rates may overestimate recharge for these and similar sites in western Oklahoma. However, these sites have deep soils and dry climates and these findings may not apply to shallower soils or wetter locations. While Mesonet-based drainage rates and HYDRUS-estimated water flux at the 60-cm depth do not agree, HYDRUS-estimated water flux at the 3-m depth actually compares fairly well with values given by previous studies.

Regional-scale recharge rates

Aquifer-scale median precipitation and drainage rates were found for aquifers with three or more Mesonet sites located above them (Table 9) and were compared to previously published recharge rates for those aquifers. This includes the Boone, Arkansas

River alluvial aquifer, Garber-Wellington, Rush Springs, Antlers, and Ogallala aquifers. Median annual drainage to these aquifers ranged from 21 mm yr^{-1} for the Ogallala to 235 mm yr^{-1} for the Boone aquifer. These results generally follow the precipitation gradient of the state, with decreasing drainage rates as you move from the east to the west. With the exception of the Arkansas River alluvial aquifer, for which only one previous recharge study was found, median annual calculated drainage rates fall within the range of recharge rates given by previous studies. This relationship between drainage rates and results of previous studies suggests that Mesonet drainage estimates may be capable of providing reliable recharge estimates at a regional scale.

Table 10. Summary of median precipitation and annual drainage at 60 cm for Mesonet sites above selected Oklahoma aquifers from 1996 through 2012. For comparison, prior published estimates of groundwater recharge for these aquifers are also shown.

Aquifer	Sites	Precipitation	Drainage	Recharge	No. Sources
		mm yr^{-1}	mm yr^{-1}	mm yr^{-1}	
Boone	3	1076	235	2.3-254	4
Arkansas River	5	1006	171	127	1
Garber-Wellington	3	893	121	7.6-203	4
Rush Springs	5	735	74	4.9-99	4
Antlers	4	936	70	8.1-152	4
Ogallala	8	497	21	1.5-54	4

Analysis of groundwater data from the OWRB's GMAP program by the CMB_{sz} method yielded aquifer-scale recharge rates for five Oklahoma reservoirs (Table 10). A total of 54 groundwater samples from the GMAP program had chloride concentrations below the detection limit of 10 mg L^{-1}, while 19 samples had sulfate concentrations beneath this limit. These samples were assumed to have chloride and sulfate concentrations of 5 mg L^{-1}, which would also create an upper limit on recharge rates. For the Ogallala aquifer, this assumed concentration results in the highest recharge rate possible being 18.9 mm yr^{-1}. The removal of samples with a $Cl:SO_4$ ratio greater than one led to the exclusion of 32 of the 155 groundwater samples taken.

Recharge rates from this analysis ranged from 4.8 mm yr^{-1} for the Canadian River alluvial aquifer to 25.4 mm yr^{-1} for both the Elk City and Gerty Sand aquifers. Because groundwater samples are representative of a large area and not specific to a certain location, recharge estimates found by the CMB_{sz} method are only comparable to Mesonet-based drainage rates summarized by aquifer (Table 10). The only aquifers that were tested under the GMAP program which also have an adequate number of Mesonet sites located above them to produce a reliable aquifer- scale median drainage rate are the Ogallala and Rush Springs aquifers. Aquifer-scale median Mesonet estimates of drainage and recharge estimated by CMB_{sz} show a high level agreement for the Ogallala aquifer, with rates of 21 and 17.8 mm yr^{-1}, respectively. However, the median Mesonet-based drainage rate and CMB_{sz} -estimated recharge rate for the Rush Springs aquifer vary by an order of magnitude, with rates of 74 mm yr^{-1} and 7.6 mm yr^{-1}, respectively. The cause of this discrepancy between the CMB_{sz} and Mesonet-based estimates for the Rush Springs aquifer is unclear. One reason could be that groundwater samples used in the CMB_{sz} calculations reflect chloride concentrations over only one year of sampling, while drainage estimates calculated using Mesonet data are given as the median of mean drainage rates for sites above the aquifer for the years 1996-2012. Recharge estimates from this analysis of groundwater chloride seem to be intermediate between the Mesonet-based drainage rates, which are slightly higher, and the CMB_{uz} recharge rates, which are much lower.

Table 11. Mean annual precipitation, number of samples, mean groundwater chloride concentrations, and recharge estimates calculated by the CMB_{sz} method for select Oklahoma aquifers sampled in the 2013 portion of the GMAP program.

Aquifer	Precipitation	No. Samples	GW chloride	Recharge
	mm yr^{-1}		ppm	mm yr^{-1}
Gerty Sand	894	5	11	25.4
Canadian River	770	34	52	4.8
Rush Springs	714	64	31	7.6
Elk City	683	13	9	25.4
Ogallala NW	587	39	11	17.8

A previous study that applied the CMB_{sz} method to groundwater in Texas and New Mexico gave groundwater chloride concentrations that ranged from 20 to 1110 mg L^{-1} and resulting recharge rates that averaged 11 mm yr^{-1} (Wood and Sanford, 1995). However, chloride concentrations in precipitation for this study ranged from 0.18 to 3.58 mg L^{-1}, with an average concentration of nearly 8 mg L^{-1}, which would lead to higher recharge rates than the concentrations below 1 mg L^{-1} found for our study area, given that precipitation levels were comparable between the studies.

CHAPTER V

CONCLUSION

This study represents the first known use of data from a meso-scale soil moisture monitoring network for estimation of groundwater recharge. Our preliminary results show that these soil moisture-based drainage estimates are greater than or equal to recharge estimates found by independent methods in central and western Oklahoma. Modeling water flux and root water uptake at depth has provided some useful information regarding this relationship, with Mesonet drainage rates falling between HYDRUS 1-D simulated fluxes at 60 cm and 300 cm in most cases. Recharge estimation using the CMB_{sz} method provided an aquifer-scale recharge value close to the median drainage values for the Mesonet sites in the Ogallala aquifer.

As with all recharge estimation methods, there are several weaknesses inherent to the drainage estimation method used here. Because the Mesonet system was originally created for weather monitoring, the suitability of site locations for soil moisture data collection was not taken into account. This led to the placement of stations in areas where soil moisture measurements may be adversely affected by nearby streams, ditches, irrigation systems, or simply by the natural soil and geological features of the area. It is also possible that measurement errors may lead to inaccurate soil moisture readings, or prevent readings from being taken at all. Additionally, the unit-gradient assumption applied in this method may not be realistic at all sites at all times. There is a need to develop additional quality control procedures to reduce the occurrence of anomalous drainage estimates. Despite these uncertainties, Mesonet-based drainage rates seem to provide reasonable drainage rates at most sites.

Using soil moisture data to estimate drainage has a number of benefits that are unique to this method. The first is that the monitoring system necessary for such measurements is already in place, so there is no start-up cost. A second strength is that this method provides the ability to estimating drainage throughout the entire state, which hasn't been done since the 1980s (Pettyjohn et al, 1983). Another strength is that the Mesonet system has been recording the necessary soil moisture data for nearly twenty years, allowing for the calculation of drainage rates at a long-term scale that is difficult for shorter research studies to encompass. Possibly the greatest strength of this method is that it allows for the calculation of drainage rates at a yearly (or even daily) time step, as opposed to other studies which merely estimate an average recharge rate for the duration of the study. This yearly drainage data may be useful for scientists, policy makers, or the general public in the context of drought and water management. An increase in the general understanding of how drainage rates vary according to climatic factors could lead to more informed decision making by land owners, policy makers, and citizens, each who value the sustainability of the limited groundwater resources of our state.

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APPENDICES

Table A1. Mesonet site ID, mean annual drainage rate, mean annual precipitation, and the ratio of drainage to precipitation for each Mesonet site for which a drainage rate could be calculated.

Site	Drainage	Precipitation	D/P
	mm yr^{-1}	mm yr^{-1}	%
'ACME'	80	781	0.10
'ALV2'	13	1030	0.01
'ANT2'	225	1139	0.20
'APAC'	73	771	0.09
'ARD2'	57	839	0.07
'ARNE'	21	561	0.04
'BEAV'	24	483	0.05
'BIXB'	383	991	0.39
'BLAC'	87	925	0.09
'BOIS'	7	386	0.02
'BOWL'	79	968	0.08
'BREC'	23	802	0.03
'BRIS'	275	951	0.29
'BUFF'	43	555	0.08
'BURN'	72	854	0.08
'BUTL'	34	699	0.05

'BYAR'	63	904	0.07
'CARL'	58	811	0.07
'CENT'	254	989	0.26
'CHER'	19	755	0.03
'CHEY'	37	712	0.05
'COPA'	197	947	0.21
'DURA'	55	1035	0.05
'ELRE'	58	819	0.07
'ERIC'	20	620	0.03
'EUFA'	41	1060	0.04
'FAIR'	52	720	0.07
'FITT'	76	910	0.08
'FORA'	356	958	0.37
'FREE'	9	655	0.01
'FTCB'	57	712	0.08
'GOOD'	18	410	0.04
'GRA2'	34	680	0.05
'HASK'	24	1050	0.02
'HECT'	112	990	0.11
'HINT'	4	763	0.01
'HOBA'	22	738	0.03
'HOLD'	145	815	0.18
'HOLL'	29	616	0.05
'HOOK'	14	436	0.03
'INOL'	188	1029	0.18
'KETC'	77	857	0.09
'KIN2'	17	705	0.02
'LAHO'	49	762	0.06
'LANE'	372	1077	0.35
'MANG'	125	671	0.19
'MARE'	8	887	0.01
'MAYR'	35	677	0.05
'MIAM'	226	1081	0.21
'NEWK'	152	945	0.16
'NOWA'	174	1017	0.17
'NRMN'	192	815	0.24
'OILT'	102	983	0.10
'OKCE'	68	971	0.07

'OKCW'	90	913	0.10
'PAUL'	244	894	0.27
'PAWN'	85	949	0.09
'PERK'	89	873	0.10
'PORT'	149	1024	0.15
'PUTN'	75	666	0.11
'REDR'	45	896	0.05
'SALL'	71	1097	0.06
'SEIL'	25	699	0.04
'SHAW'	28	860	0.03
'SKIA'	133	1011	0.13
'SLAP'	30	530	0.06
'SPEN'	14	882	0.02
'STIG'	253	1065	0.24
'STIL'	218	874	0.25
'TIPT'	29	631	0.05
'VINI'	68	1082	0.06
'WASH'	654	855	0.76
'WATO'	88	755	0.12
'WAUR'	128	797	0.16
'WEBR'	73	995	0.07
'WILB'	83	1102	0.08
'WIST'	59	1118	0.05
'WOOD'	24	630	0.04

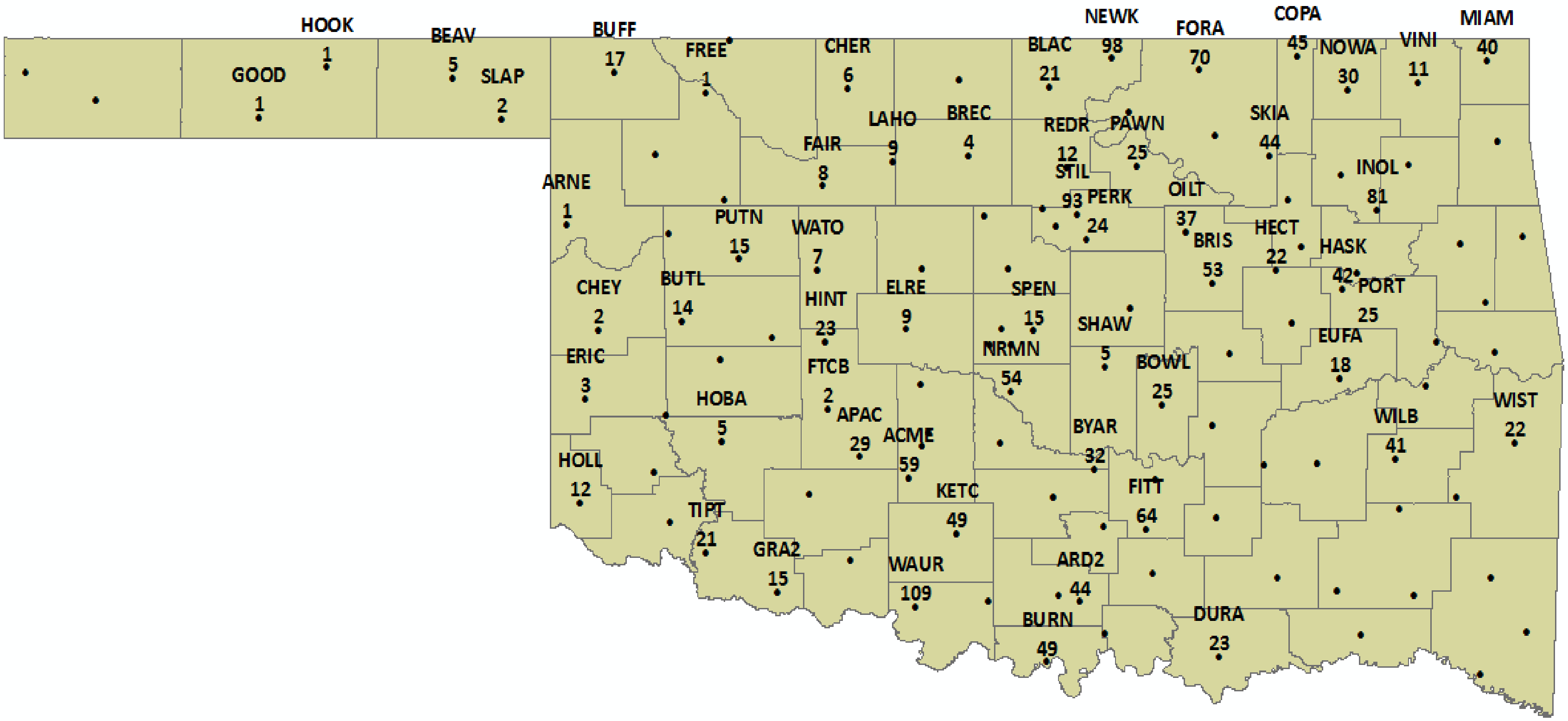

Figure A1. Mean state-wide drainage rates for the year 2006.

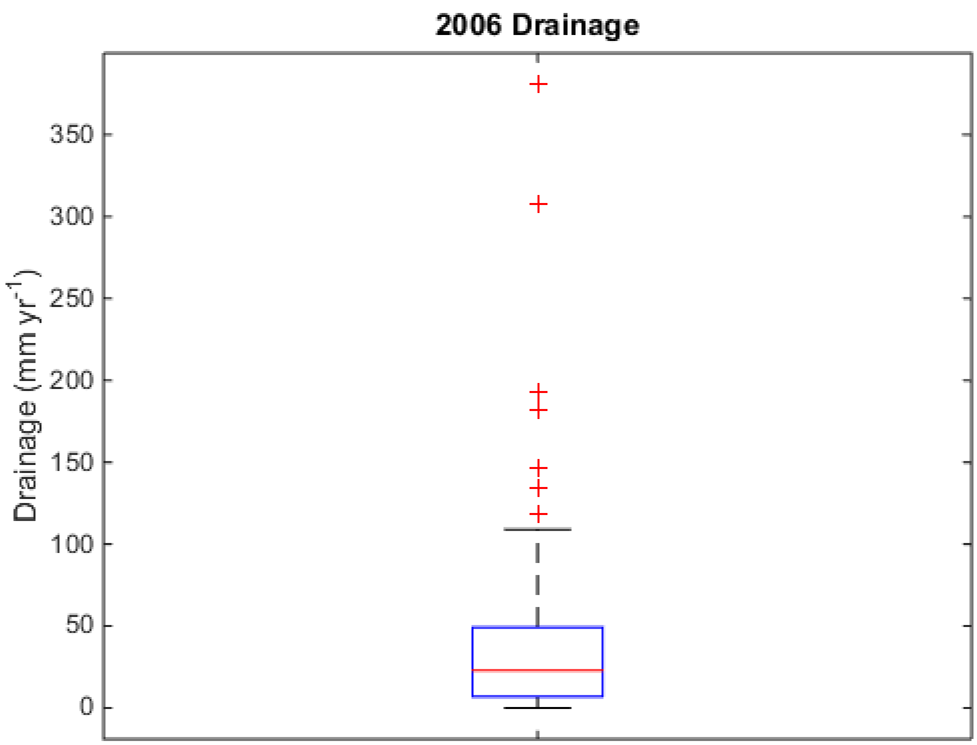

Figure A2. Boxplot of calculated drainage rates for the year 2006. The line in the center of the box is the median drainage rate, the lower and upper edges of the box are the 25th and 75th percentiles, and the whiskers extend to the most extreme data points not considered outliers. The red crosses correspond to the four highest calculated drainage rates, which are considered outliers. These sites included the Sallisaw, Washington, Lane, Paul's Valley, Centrahoma, Mangum, and Stigler Mesonet sites.

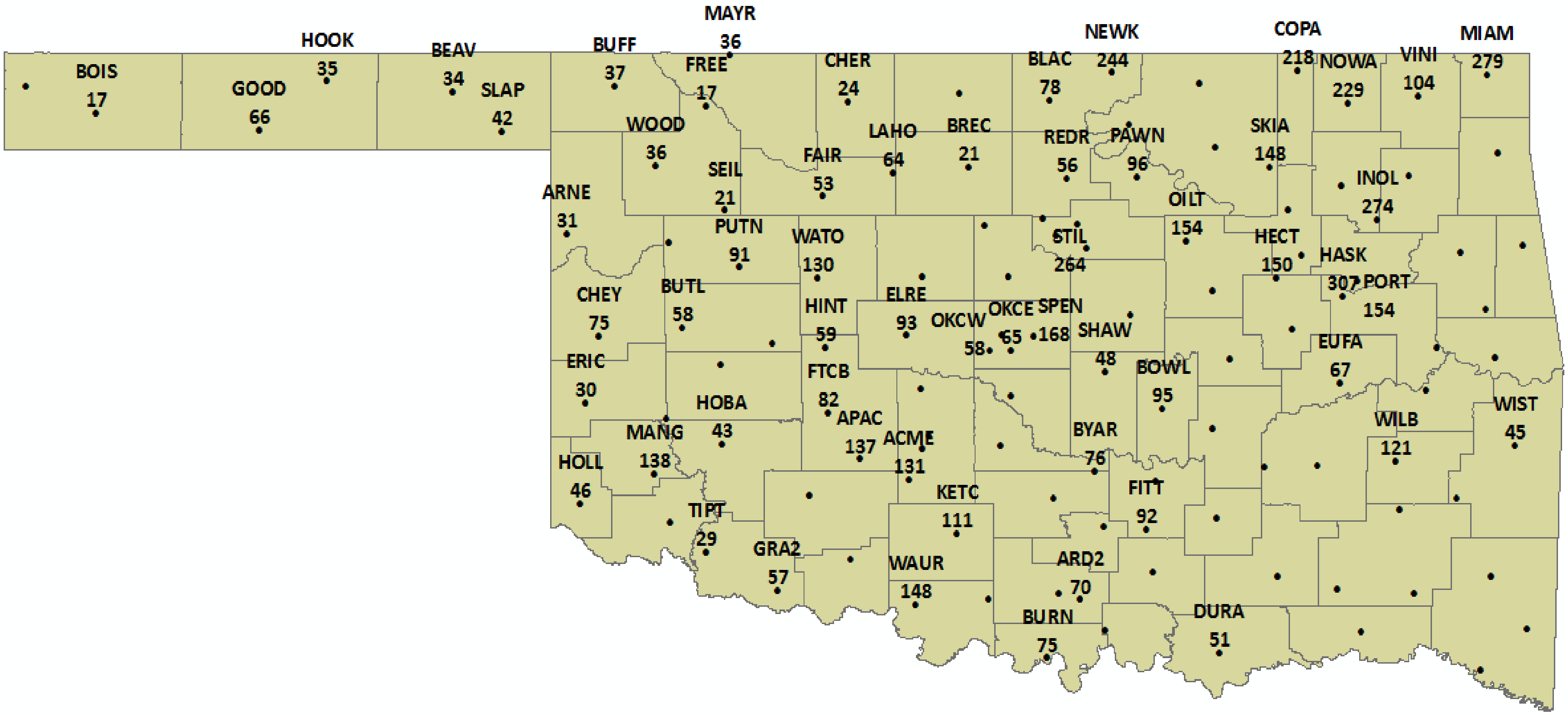

Figure A3. Mean state-wide drainage rates for the year 2007.

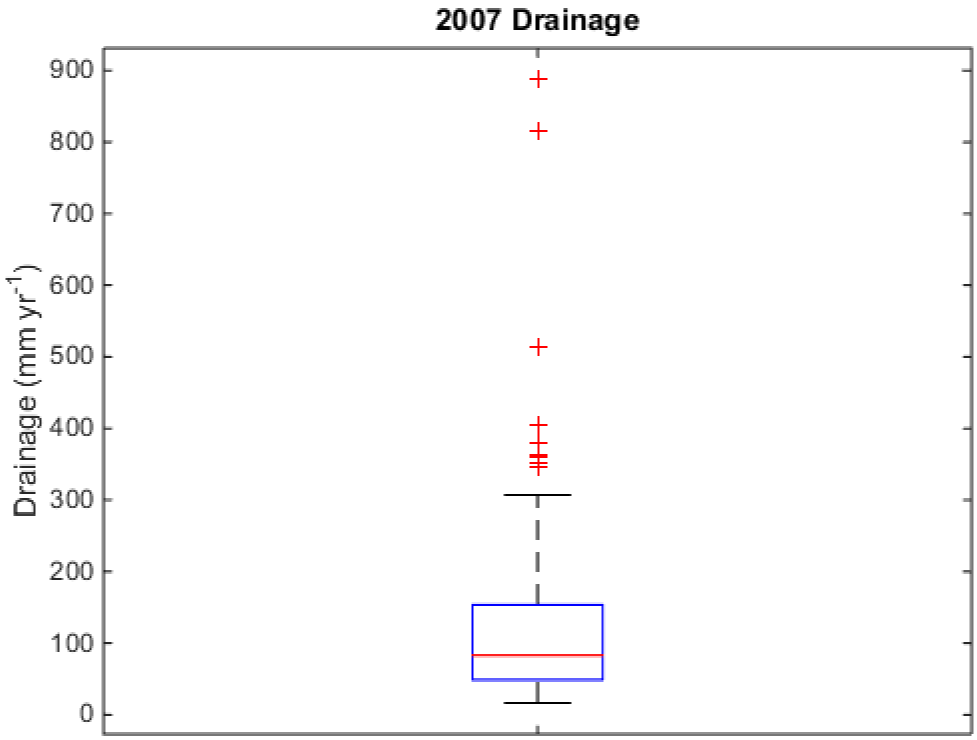

Figure A4. Boxplot of calculated drainage rates for the year 2007. The line in the center of the box is the median drainage rate, the lower and upper edges of the box are the 25th and 75th percentiles, and the whiskers extend to the most extreme data points not considered outliers. The red crosses correspond to the four highest calculated drainage rates, which are considered outliers. These sites included the Sallisaw, Washington, Lane, Bristow, Foraker, Centrahoma, Paul's Valley, Stigler, and Norman Mesonet sites.

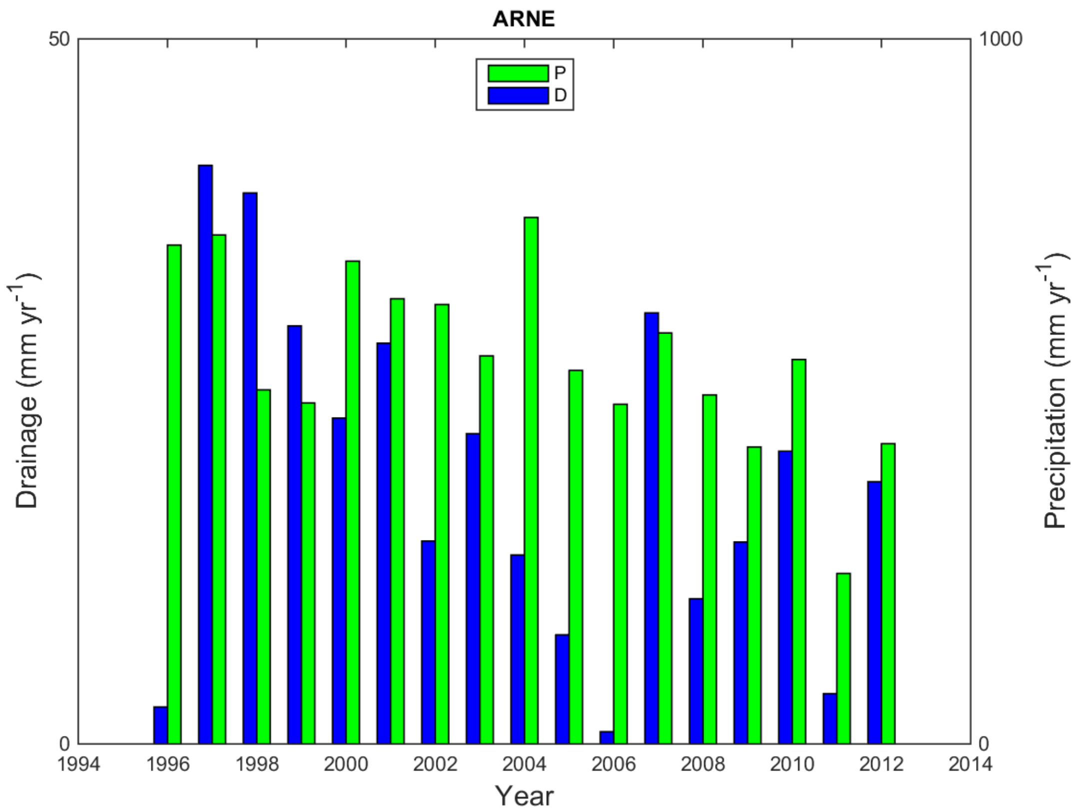

Figure A5. Total precipitation and drainage at the 60 cm depth by year from 1996-2012 at the Arnett Mesonet site.

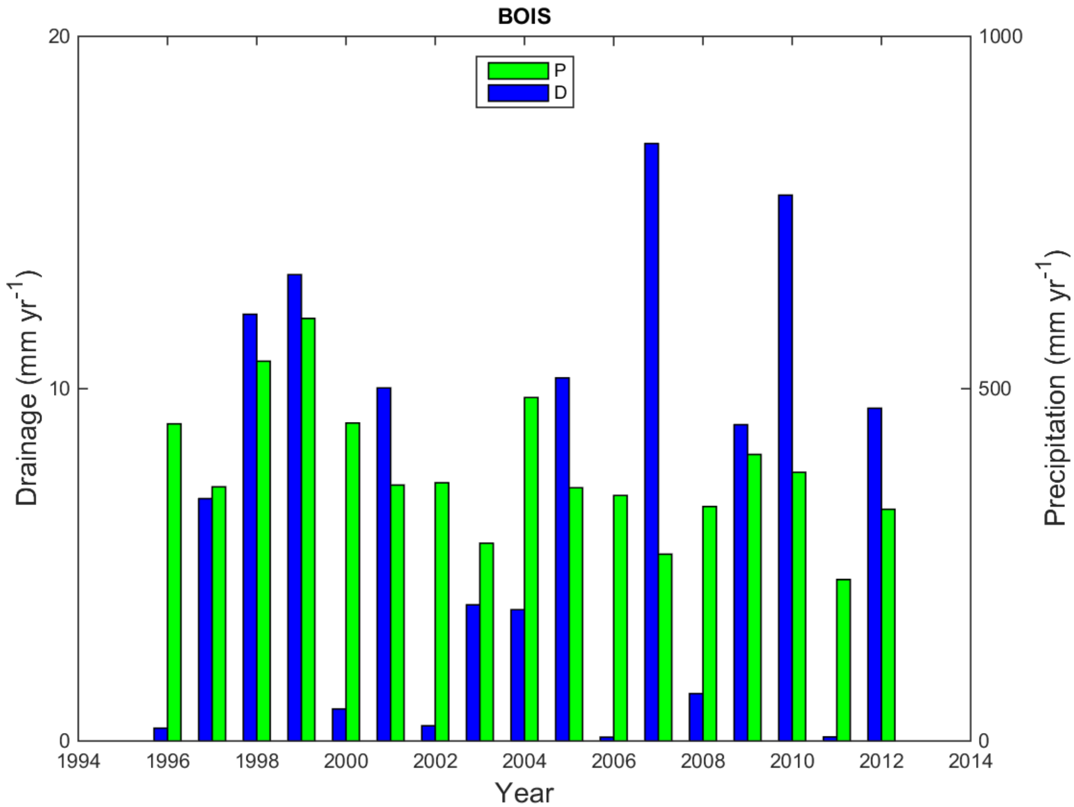

Figure A6. Total precipitation and drainage at the 60 cm depth by year from 1996-2012 at the Boise City Mesonet site.

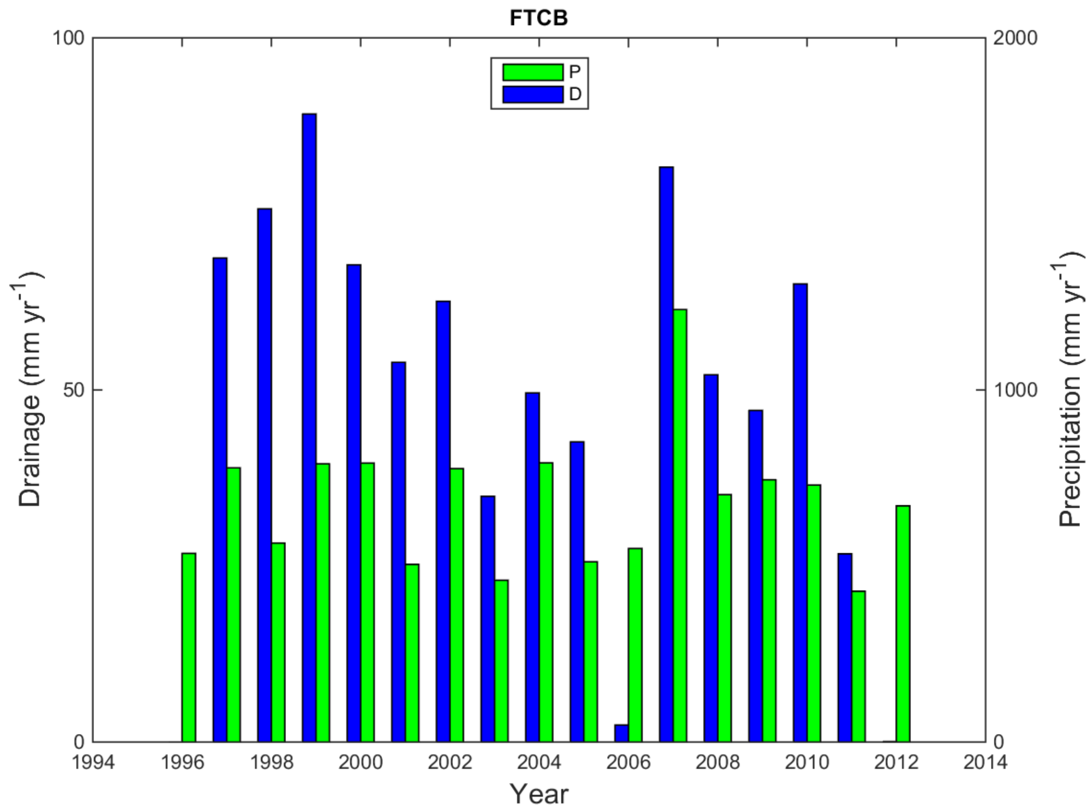

Figure A7. Total precipitation and drainage at the 60 cm depth by year from 1996-2012 at the Fort Cobb Mesonet site.

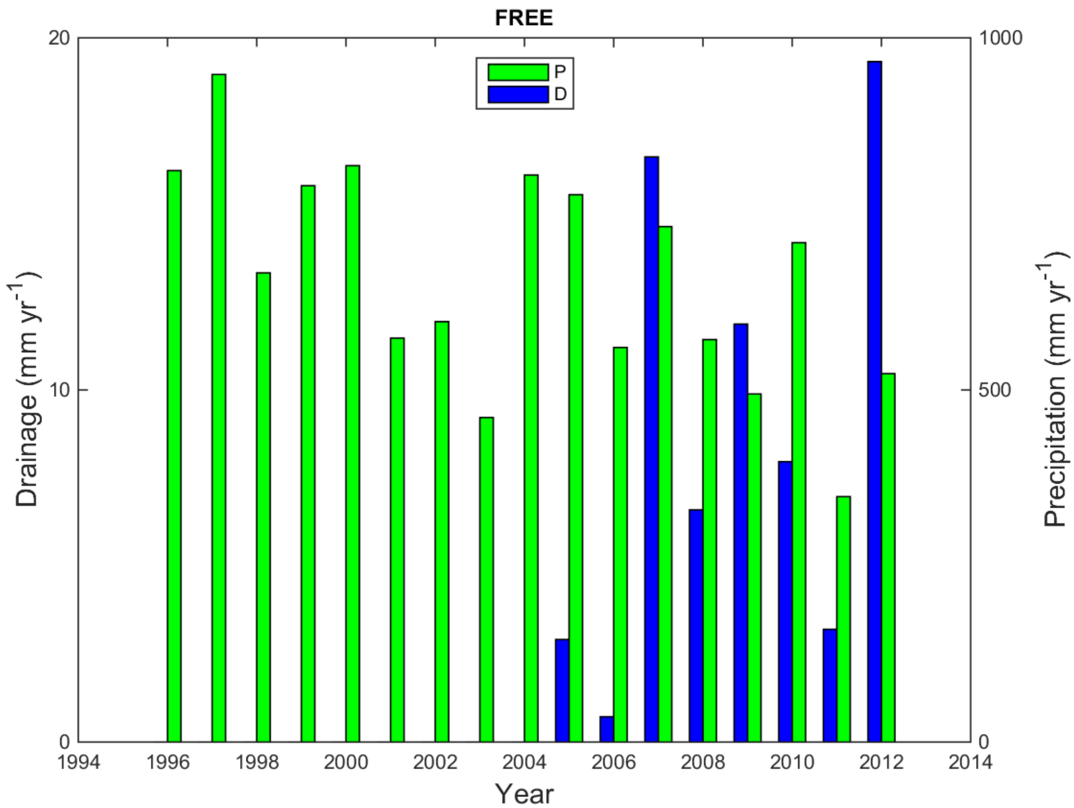

Figure A8. Total precipitation and drainage at the 60 cm depth by year from 1996-2012 at the Freedom Mesonet site.

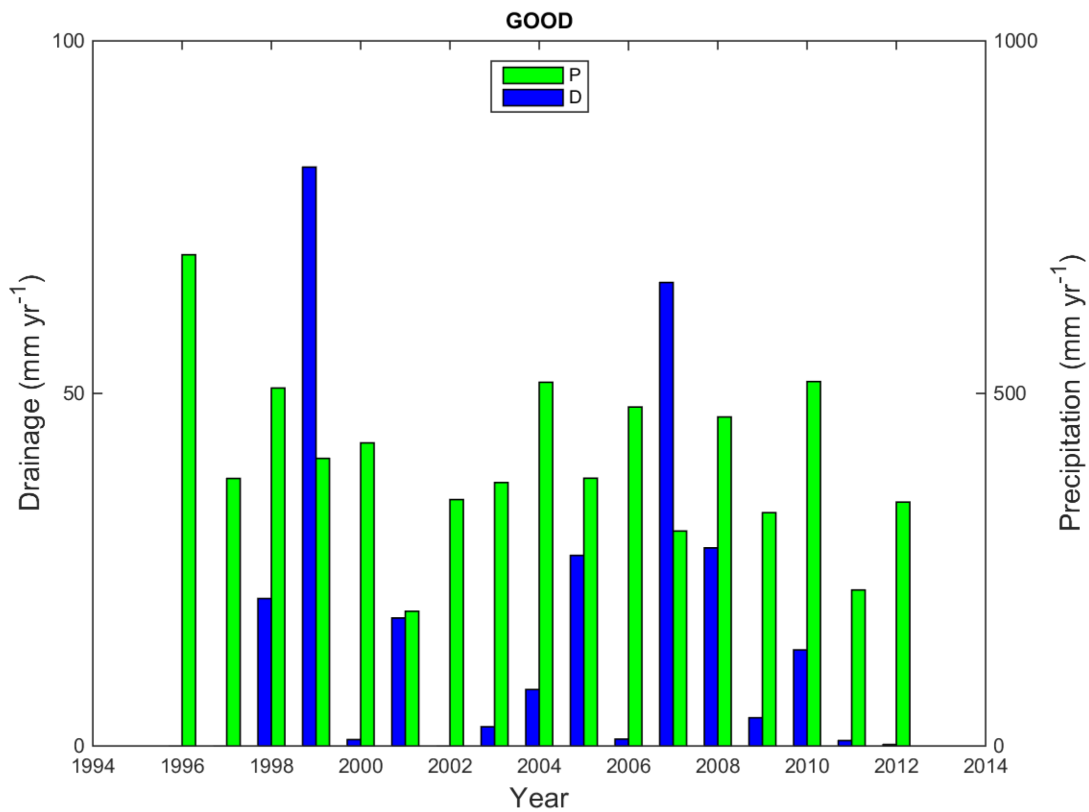

Figure A9. Total precipitation and drainage at the 60 cm depth by year from 1996-2012 at the Goodwell Mesonet site.

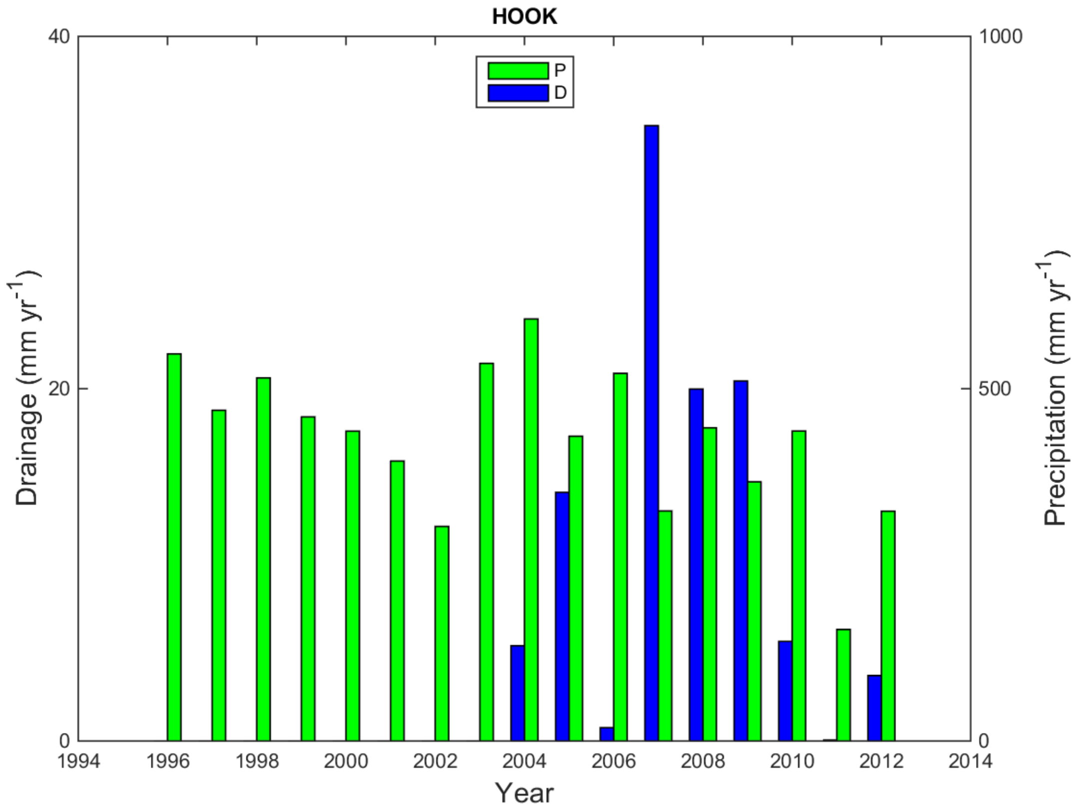

Figure A10. Total precipitation and drainage at the 60 cm depth by year from 1996-2012 at the Hooker Mesonet site.

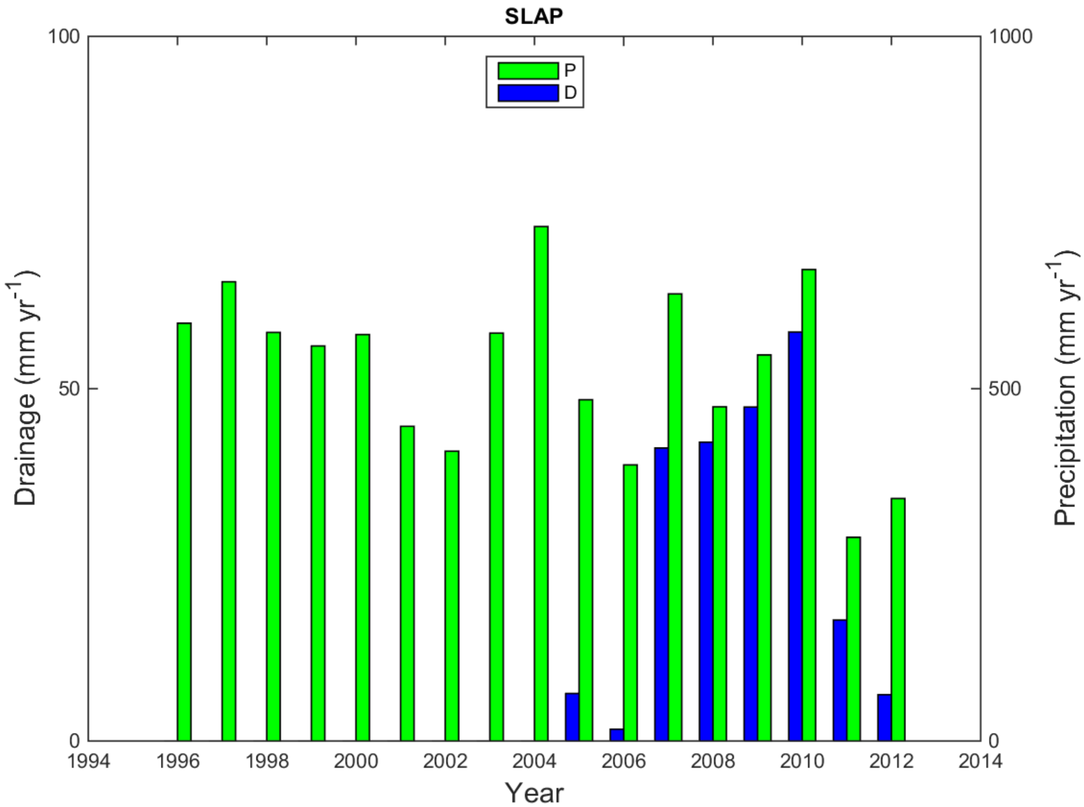

Figure A11. Total precipitation and drainage at the 60 cm depth by year from 1996-2012 at the Slapout Mesonet site.

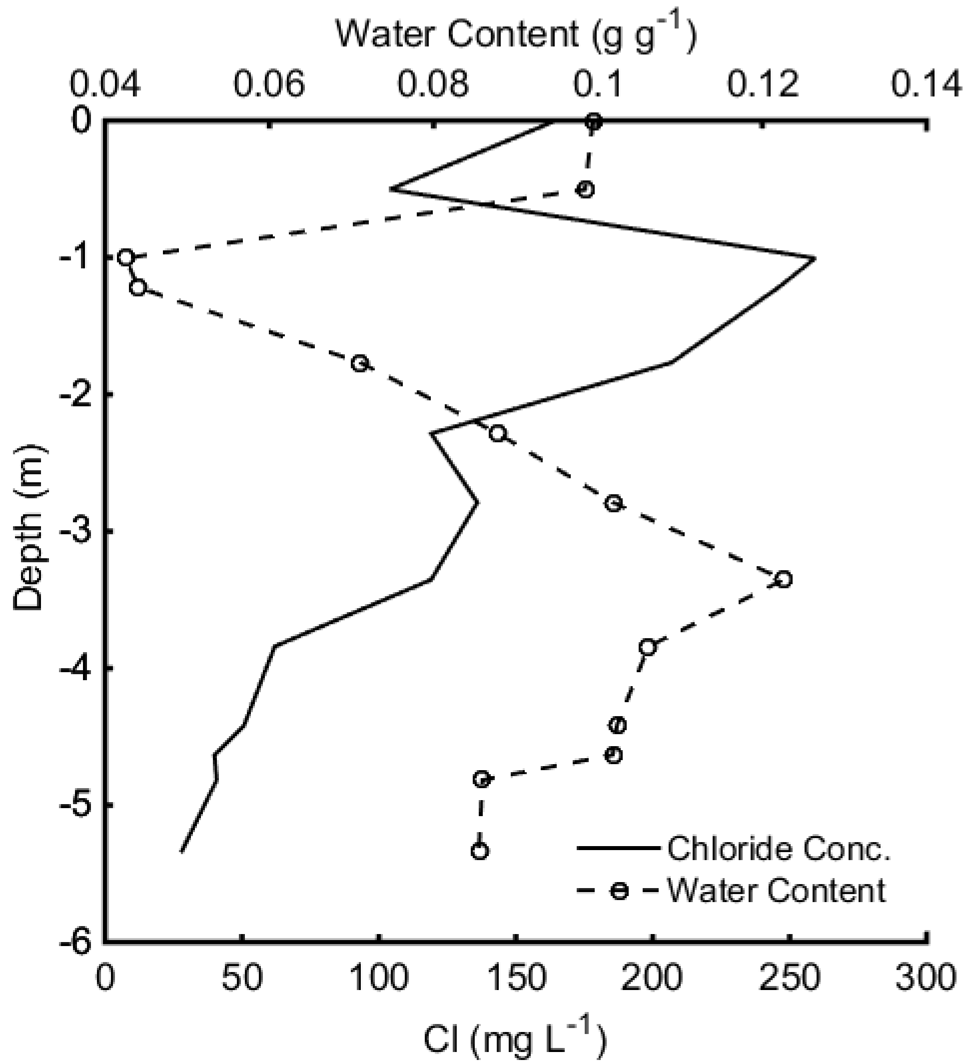

Figure A12. Chloride concentration and gravimetric water content versus depth for the soil cores collected at Arnett, Oklahoma.

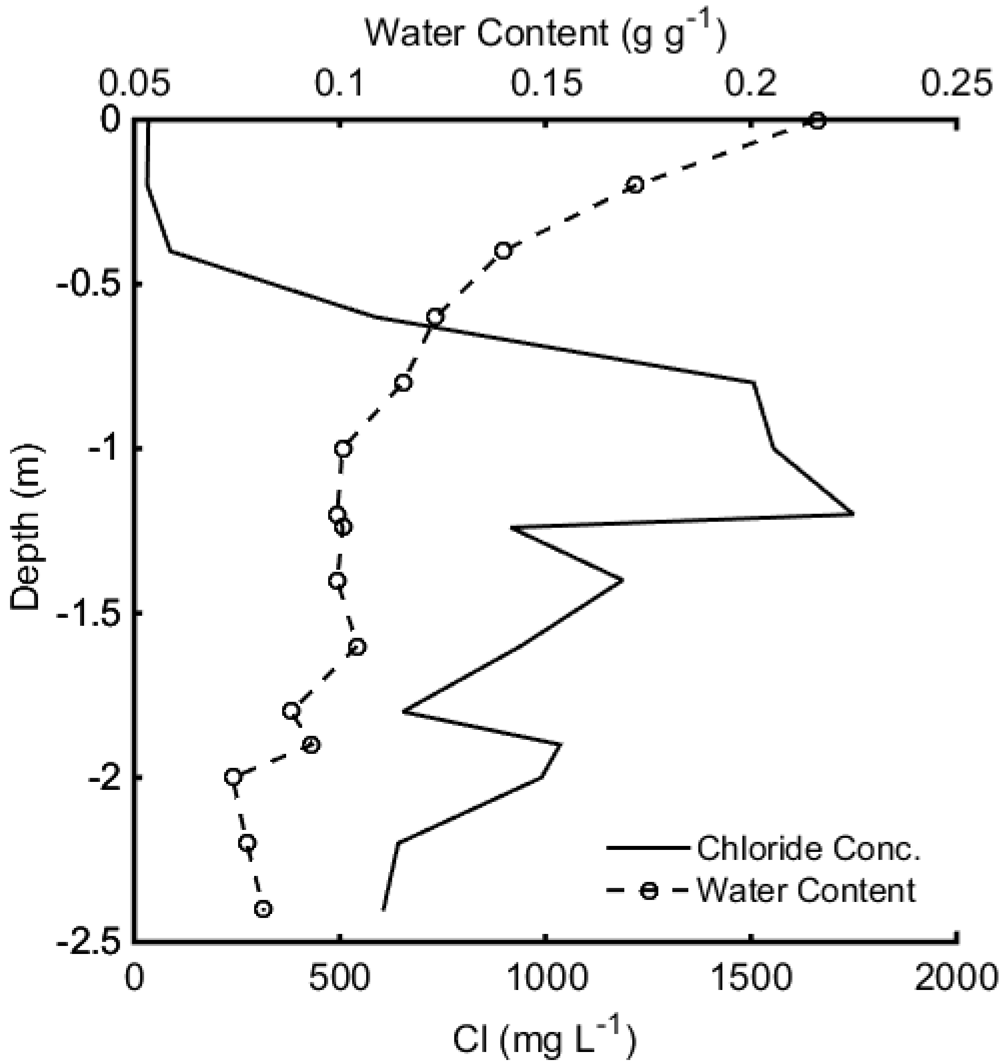

Figure A13. Chloride concentration and gravimetric water content versus depth for the soil cores collected at Boise City, Oklahoma.

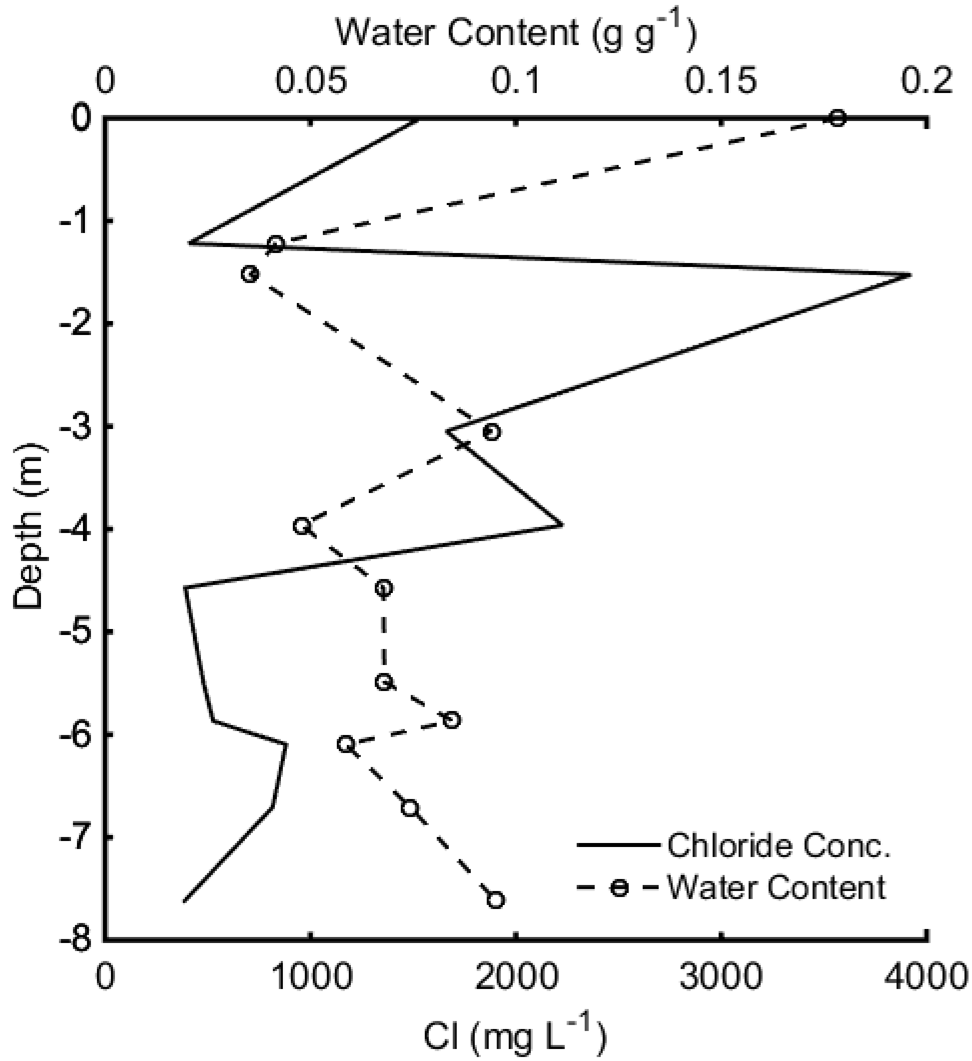

Figure A14. Chloride concentration and gravimetric water content versus depth for the soil cores collected at Fort Cobb, Oklahoma.

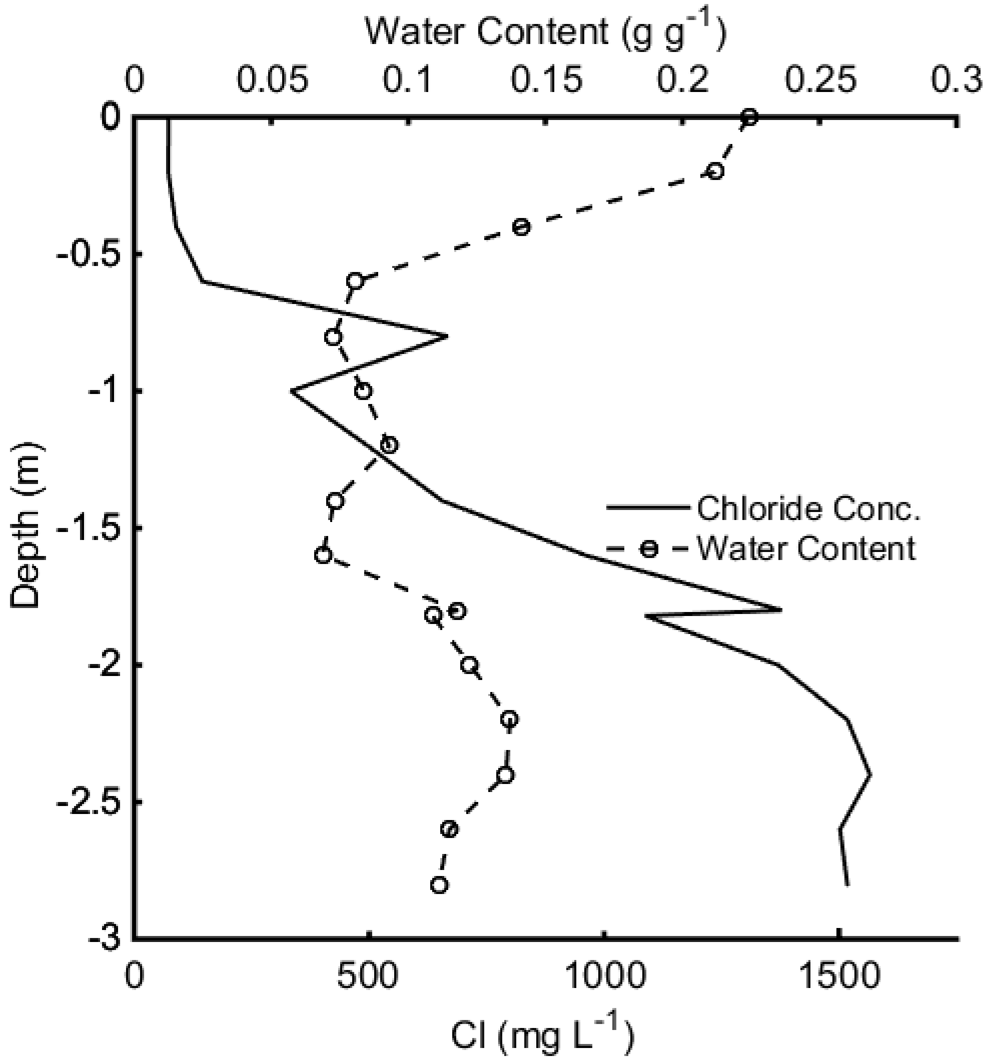

Figure A15. Chloride concentration and gravimetric water content versus depth for the soil cores collected at Freedom, Oklahoma.

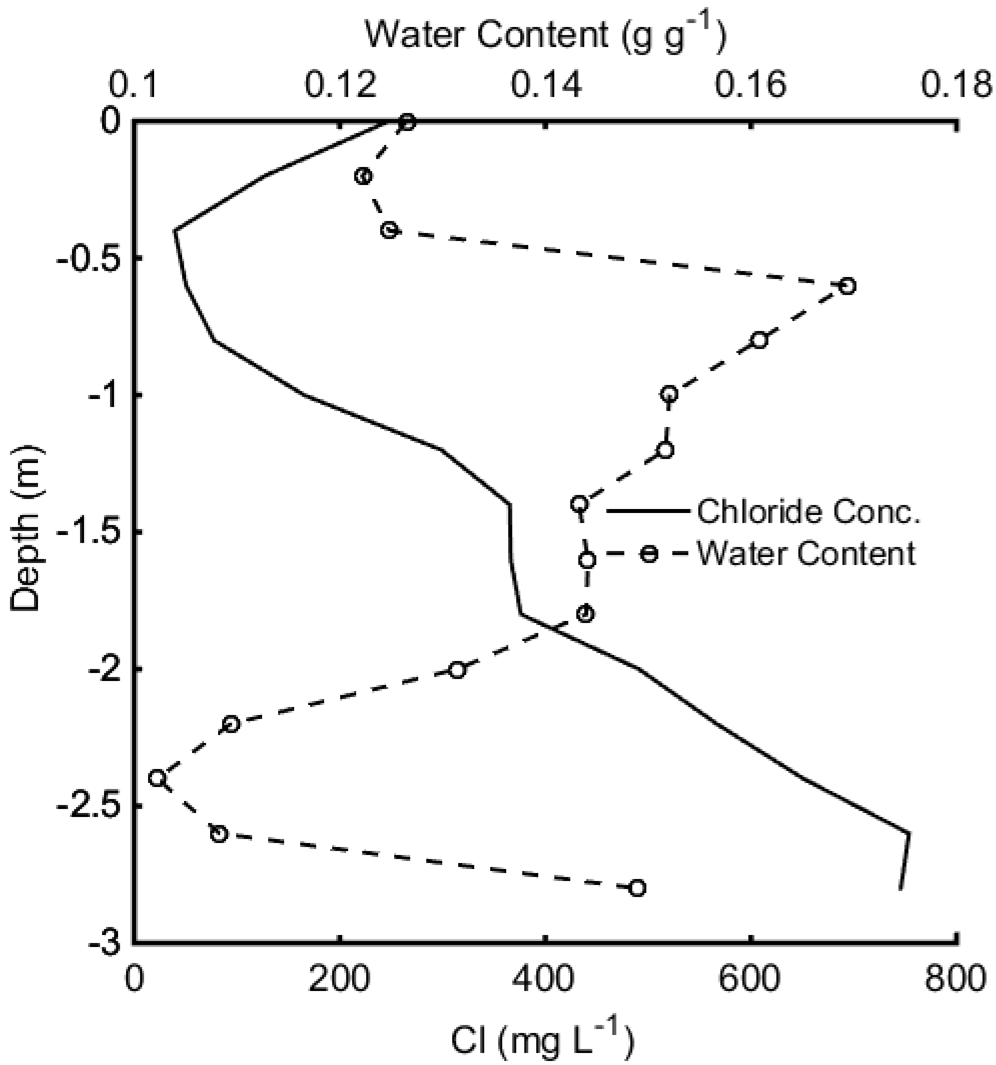

Figure A16. Chloride concentration and gravimetric water content versus depth for the soil cores collected at Hooker, Oklahoma.

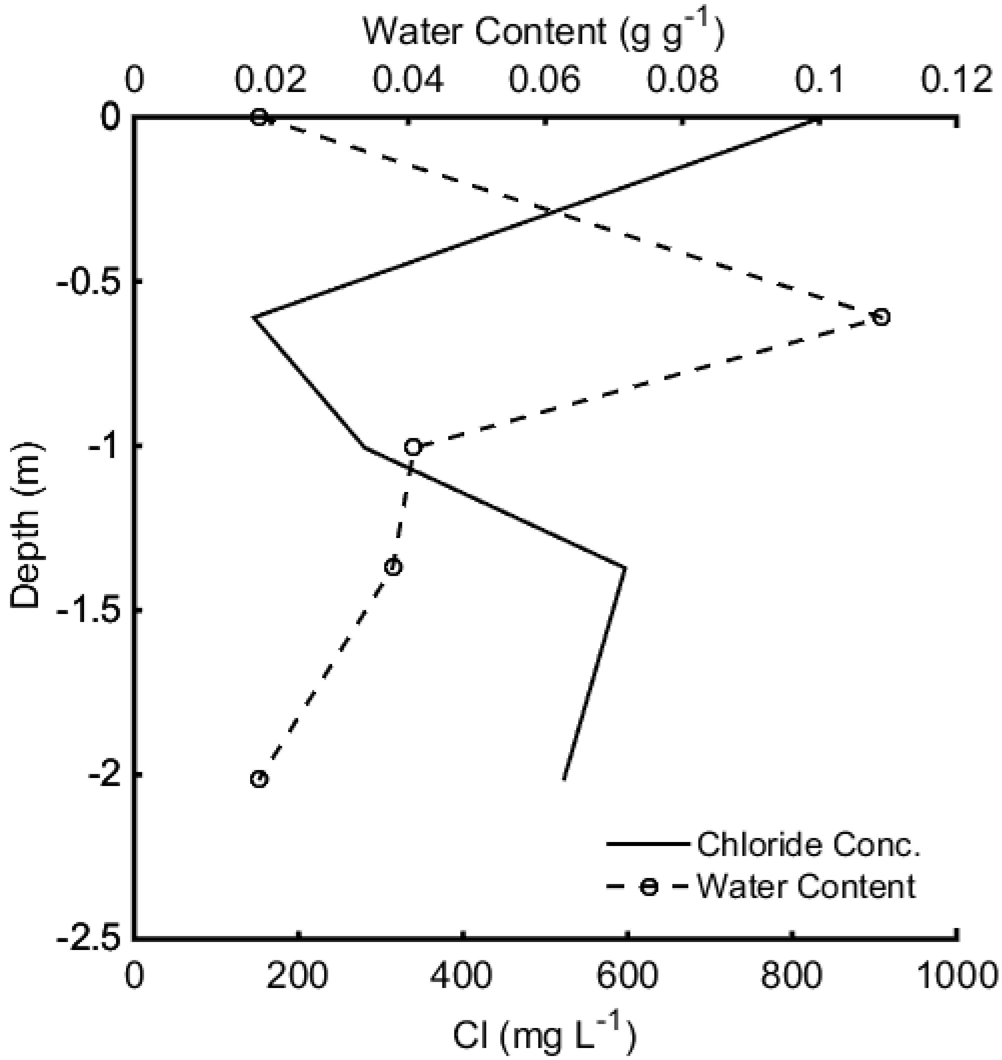

Figure A17. Chloride concentration and gravimetric water content versus depth for the first of two soil cores collected at Slapout, Oklahoma.

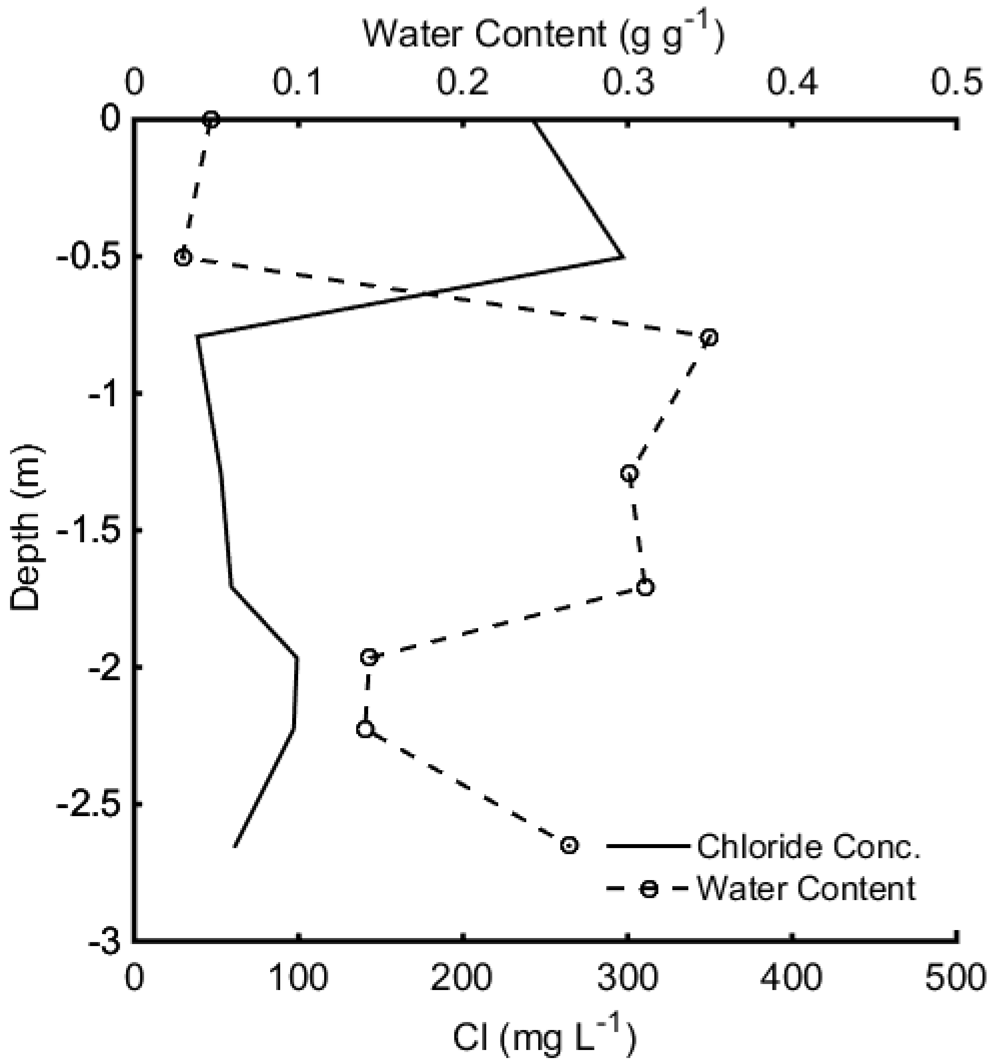

Figure A18. Chloride concentration and gravimetric water content versus depth for the second of two soil cores collected at at Slapout, Oklahoma.

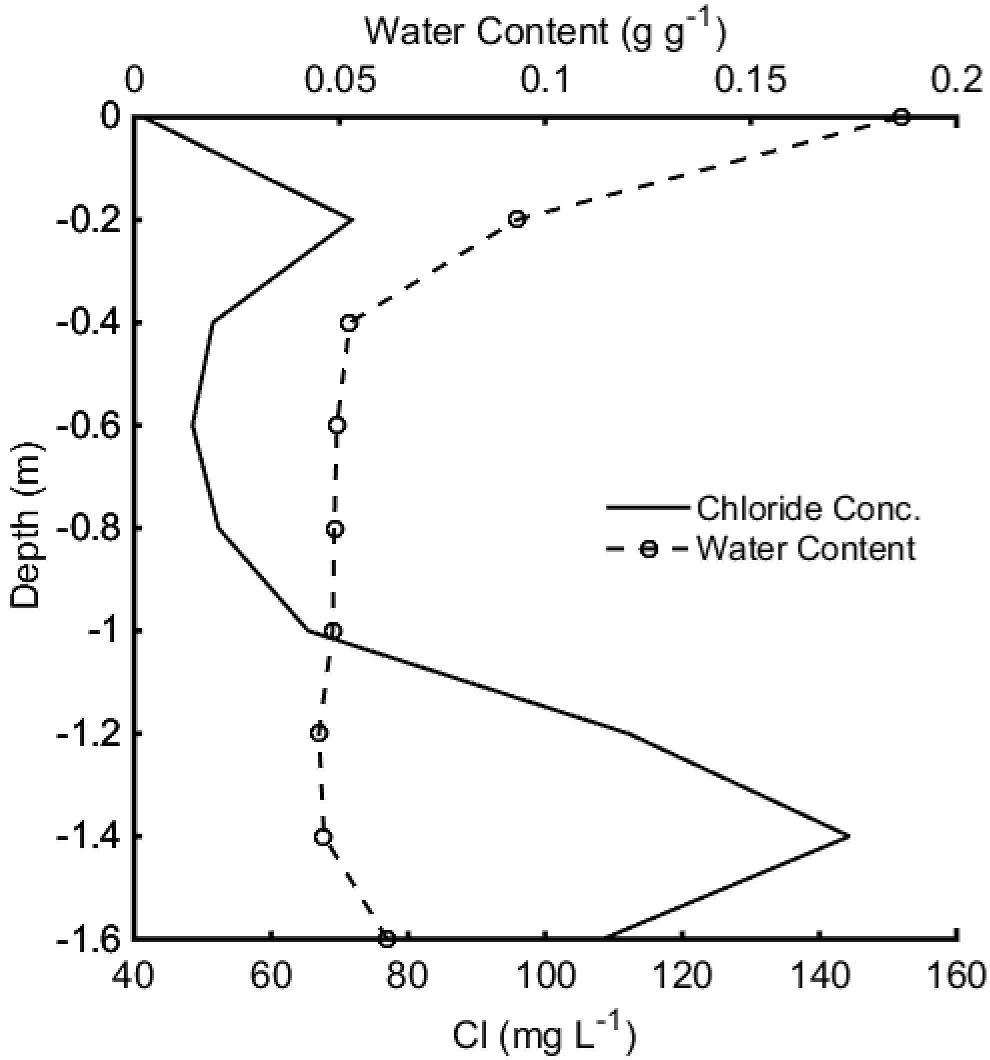

Figure A19. Chloride concentration and gravimetric water content versus depth for the soil cores collected at Woodward, Oklahoma.

VITA

Briana Sallee

Candidate for the Degree of

Master of Science

Thesis: ESTIMATING GROUNDWATER RECHARGE USING THE OKLAHOMA MESONET

Major Field: Plant and Soil Sciences

Biographical:

Education:

Completed the requirements for the Master of Science in Plant and Soil Sciences at Oklahoma State University, Stillwater, Oklahoma in May, 2015.

Completed the requirements for the Bachelor of Science in Environmental Science (Natural Resources option) at Oklahoma State University, Stillwater, Oklahoma in July 2013.