USING STABLE ISOTOPE HYDROLOGY TO PARTITION EVAPOTRANSPIRATION IN THE SAGEBRUSH STEPPE

by

Kellie Jo Rey

A thesis
submitted in partial fulfillment
of the requirements for the degree of
Master of Science in Hydrologic Sciences
Boise State University

May 2016

BOISE STATE UNIVERSITY GRADUATE COLLEGE

DEFENSE COMMITTEE AND FINAL READING APPROVALS

of the thesis submitted by

Kellie Jo Rey

Thesis Title: Using Stable Isotope Hydrology to Partition Evapotranspiration in the Sagebrush Steppe

Date of Final Oral Examination: 09 March 2016

The following individuals read and discussed the thesis submitted by student Kellie Jo Rey, and they evaluated her presentation and response to questions during the final oral examination. They found that the student passed the final oral examination.

James McNamara, Ph.D.	Chair, Supervisory Committee
Matthew Germino, Ph.D.	Member, Supervisory Committee
Alejandro Flores, Ph.D.	Member, Supervisory Committee

The final reading approval of the thesis was granted by James McNamara, Ph.D., Chair of the Supervisory Committee. The thesis was approved for the Graduate College by John R. Pelton, Ph.D., Dean of the Graduate College.

DEDICATION

I would like to dedicate this thesis to all those who do not follow convention or give up when the path is blocked, seeking to understand the truth and elements of the matter. It is these bold humans who often break the mold and pave the way for future generations. It is these people who do not let go of the student mind, always learning and continuing to expand levels of understanding. These lights shine the brightest, these wings fly the farthest.

I would also like to dedicate this effort to my mother Jewel Michelle Rey and my father Gregory Lynn Rey.

ACKNOWLEDGEMENTS

I would like to acknowledge the Germino Lab at the USGS FRESC station in Boise, Idaho. Specifically, Dr. Brynne Lazarus who helped guide the implementation of stable isotope techniques used for this study. In addition, I would like to thank Dr. Matthew Germino who accepted me as his graduate student and helped me find a project that could be completed with no funding whatsoever. I would also like to acknowledge the Pilliod Lab at the USGS FRESC in Boise for the opportunity to broaden my skillset of taxonomic entomology, flexible schedule, and continued employment as it was truly this support that allowed me to continue the pursuit of my masters. I would also like to acknowledge all the field assistants who accompanied me on this scientific adventure: Ashley Rohde, Martha Brabec, Mellissa Hill, Brian Yelen, Catherine Carlise, Joel de Shepper, Kara Kirkpatric, Charles Rohde, and Hadley Kirkpatric. In preparing this manuscript, I find that my life has become so full and enriched with love. To this effect, I would like to thank Eric Smith for his encouragement, support, and willingness to journey together. Many thanks to Martha T. Williams for her professional advice on the manuscript in the final edits: I’ve got a delicious Cabernet and a dinner with your name on it!

ABSTRACT

Evapotranspiration (ET) is a major part of ecosystem water loss. This thesis aims to develop methods that partition soil water loss into evaporation (E) and transpiration (T). Water balance methods may improve with relative contributions of evaporation (E) and transpiration (T). Isotopic fractionation distinguishes soil water loss due to evaporation from that of plant uptake. This provides a means to assess E and T in retrospect rather than only measuring ET fluxes. To measure the isotopic composition of soil water, we used a liquid-vapor equilibration method following Wassenaar (2008). Experimental trials of different soil amounts and equilibration times were performed to adapt the liquid-vapor equilibration method for dry desert soils. We tested a silty loam soil with 10%, 5%, and 2% gravimetric water content (GWC) and found time-to-equilibration was 3, 4, and 5 days, respectively. Second, we tested the ability of a simplified isotope mass balance model to predict 100% E following Wenninger et al. (2010), under controlled conditions with no plants available to remove soil water for T. The simplified model resulted in 99% E (+/- 4.3%). This was the first experiment to test this model under controlled conditions. Third, we used the simplified model to assess changes in E and T across microsites, at a sagebrush steppe field site in southern Idaho. We expected the proportion of E:ET to change with time and vegetation type. Soil water loss and changes in isotopic composition from 0-10 cm were measured in a 72-hour time series. We evaluated ratios of E:ET in sagebrush, bunchgrass, and bare ground microsites in June and September 2014. In September, sagebrush used 18% of soil water from the upper 10 cm for transpiration. Bunchgrass microsite type did not use the near-surface water for transpiration. This method appears to be successful in measuring E:ET ratios in retrospect and may be used to further understand water losses in the sagebrush steppe and improve water balance methods.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

LIST OF ABBREVIATIONS

USGS	United State Geological Survey
FRESC	Forest Rangeland Ecosystem Science Center
USDA	United States Department of Agriculture
NRCS	Natural Resource Conservation Service
ET	Evapotransipration
E	Evaporation
T	Transpiration
E: ET	Proportion of evapotranspiration that is evaporation
EPSCoR	Experimental Program to Stimulate Competitive Research
VSMOW	Vienna Standard Mean Ocean Water
GMWL	Global Meteoric Water Line
GWC	Gravimetric Water Content
WSS	Web Soil Survey
LVE	Liquid-Vapor Equilibration method

CHAPTER 1. PLANT, SOILS, CLIMATE: AN ECOHYDROLOGIC APPROACH

1.1 Evapotranspiration

Ecohydrology is the study of hydrologic effects on ecosystems and how biologic processes affect aspects of the water cycle (Nuttle 2002). The three main aspects common to many ecohydrological investigations are climate, soil, and vegetation, although vertebrate and invertebrate interactions are also often important and studied.

Evapotranspiration (ET) - a major ecohydrologic process - is composed of two terms: evaporation (E) and transpiration (T). Incoming precipitation and outgoing ET, surface runoff, and infiltration affect the water balance in a system. In semi-arid regions of the world, precipitation is variable and limited; what water received is primarily lost to ET. Surface soil moisture varies across the landscape as grasses and shrubs use water resources in different ways and may even alter the hydrologic regime at the microsite level (Ryel et al. 2008, 2010). Vegetation and surface conditions are sensitive to the erratic and changing conditions in the water balance of arid ecosystems (Scheffer et al. 2001; Gutierrez et al. 2006). ET fluctuates with climate and landscape changes as resultant soil moisture varies with disturbance and changing vegetation (Wilcox and Thurow 2006; Obrist et al. 2004; Neilson and Marks 1994). Relative water loss across microsites and vegetation types are important to monitor (Anderson et al. 2011), as surface conditions continue to change.

Currently, ET is measured as a group term that does not distinguish evaporation and transpiration. It is important to partition these two means for water loss into E and T to improve surface water; balance, modeling, and monitoring efforts. By measuring proportions of E:ET, may be possible to better account for water losses in a system.

1.2 Separation of Soil Water Losses to Atmosphere from ET into E:ET

Major available tools to measure or model water loss as ET from the landscape include the eddy flux method and numerical surface flux models (Gutiérrez et al. 2006). These methods do not distinguish E from T but the ET results are used in ecosystem energy balance studies. Energy balance in the sagebrush steppe can have closure rates up to 30% (Allen et al. 2011). The problem may be scale of measurements and heterogeneity of the landscape (Foken 2008). Proportions of E:ET as they vary with vegetation types across the landscape are important to consider. Few studies have tried to quantify contributions of E and T from different vegetation and surface types in semi-arid ecosystems. Partitioning ET into E and T in combination with vapor fluxes increases the ecohydrologic information (Hsieh et al. 1998; Robertson and Gazis 2006; Zhang et al. 2010). Stable isotope hydrology can be used to differentiate soil water removed by evaporation. Using stable isotope hydrology to partition ET is ideal in semi-arid regions where E is greater than T (Griffis 2013). This technique has limits when the evaporative proportion of ET is small (<10%) (Kool et al. 2014).

This thesis aims to develop a set of methods to determine proportions of E:ET in retrospect. Microsite differences in this proportion of E and T will vary in a semi-arid ecosystem. The approach we use is to measure water content and isotopic composition of soil in a time series. The depth at which the isotopic value is the greatest is the evaporative front. This is an important location to identify, as the isotopic composition of the evaporative front is required for use in the mass balance model. Goals of this study are to determine the amount of time needed for liquid-vapor equilibration of a silty loam soil under changing moisture conditions (Chapter 2. “Liquid-Vapor Equilibration”), to test a simplified model to predict 100% evaporation under controlled conditions using stable isotope mass balance equations (Chapter 3. “Partitioning Evapotranspiration”), and apply these methods to partition evapotranspiration into its components, evaporation (E) and transpiration (T), characterizing microsite differences in a natural setting (Chapter 4. “Landscape Application: Partitioning Evapotranspiration across Microsites in the Sagebrush Steppe). We performed experiments that measured shifts in isotopic profiles which we predict will reflect recent evaporation and ratios of E:ET will vary in time and space across microsites in the sagebrush steppe.

1.3 Stable Isotope Hydrology

Isotopes are different species of the same element with the same number of protons but vary in the number of neutrons within the nucleus. Hydrogen and oxygen species exist in natural water and relative abundances vary. Protium (^{1}H) is the common species of hydrogen being 99.99% abundant on this planet and deuterium (^{2}H) is the rare species with 0.015% abundance. The common species of oxygen, ^{16}O is 99.76% abundant and the rare species, ^{18}O is 0.20% abundant. An isotopologue is a particular combination of isotopes that may form a water molecule. The three main isotopologues of water are $^{1}H_2{}^{16}O$, $^{1}H_2{}^{18}O$, and $^{1}H^{2}H^{16}O$. When the molecules change state or phase (i.e. liquid, vapor, solid) they fractionate and change the proportion of isotopologues in a pool of water. The stable isotope ratio is the fraction of rare to common species, so we consider the ratios of $^{2}H/^{1}H$ and $^{18}O/^{16}O$ for reporting water values. Isotopic ratios are in reference to Vienna Standard Mean Ocean Water (VSMOW) and reported in units of permil (‰).

Different isotopologues are distinguishable in natural waters and show patterns of distribution in precipitation at global and local scales (Dansgaard 1964; Gat 1996; Herbin et al. 2007). Atmospheric and hydrologic processes leading to phase change of water alter the proportion of isotopic ratios when fractionation of the molecules occur. Hereafter the term evapofractionation refers to fractionation specific to the evaporative process. When soil water undergoes evapofractionation, the lighter water ($H_2{}^{16}O$) will react faster than the heavier ($H_2{}^{18}O$), leaving the system faster and concentrating heavier isotopes in residual water. The opposite situation occurs when water undergoes condensation, heavier water leaves faster and concentration of the lighter isotopes increases. These concentrations of heavy and light isotopes are referred to as enrichment and depletion, respectively.

Soil water loss (Figure 1.1a) due to plant uptake for transpiration (Figure 1.1b) is discernable from direct evaporation (Figure 1.1c) because plant uptake does not alter the isotopic ratio of soil water (Wershaw, Friedman, and Heller 1966; White et al. 1985; Ehleringer and Dawson 1992; Zhang et al. 2010). As evapofractionation changes the isotopic composition, we can isolate the portion of water lost due to evaporation from that of plant uptake for transpiration (Yakir and da SL Sternberg 2000, Zhang et al. 2010).

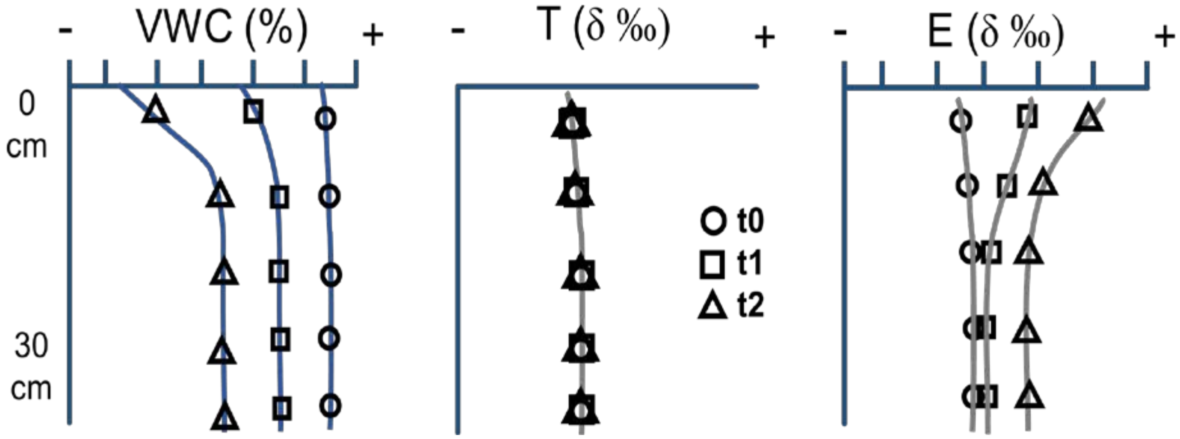

Figure 1.1 (a-c). Theoretical evolution of soil water profiles (a) as they experience water loss only to plant uptake (b) or evaporation (c). The first profile is right after a wetting event (t_0 circles), when isotopic values and water content are uniform. The next step is water loss due to ET (t_1 squares), when the water content decreases, and isotopic values of soil water increase from E. Notably the isotopic composition of soil water is unchanged from T. Continued soil water loss due to ET (t_2 triangles) will further water loss and increase of the isotopic composition.

1.4 Introduction to the Landscape

Study Site

This study site is located in the western United States in the Northern Great Basin ecoregion of southern Idaho (Figure 1.2). The research area is a dissected high lava plateau located 20 miles southwest of Twin Falls, Idaho at the Hollister EPSCoR Sagebrush Flux Site. The flux tower is located at 42°19'26.56"N, 114°42'3.29"W and 4675 feet elevation. See data descriptions and documentation of flux data from Zhao and Allen (2014).

The plant community at this site in Spring 2014 was sagebrush steppe with dominant species being Wyoming big sage (*Artemisia tridentata* Nutt. spp. *wyomingensis* Beetle&Young), Sandberg bluegrass (*Poa secunda* J.Presl), and Bottlebrush Squirreltail (*Elymus elymoides* Raf.). Cheatgrass (*Bromus tectorum* L.) was sparse and a variety of biologic soil crusts and cacti were present. Vegetation was patchy and interspersed with areas of bare ground (Figure 1.3). This area was open cattle range and this plant community remains at high risk for disturbance due to wildfire induced change (Pellant, Abbey, and Karl 2004).

The soil regime of south central Idaho originated from mixed parent materials due to geologic and geomorphic processes. Extrusive volcanic deposits, primarily basalt, form the parent material, with both fluvial and aeolian deposits. This site was characterized as a Shabliss silt loam with 1-4% slope, underlain by a Pliocene-aged olivine basalt and capped with mixed alluvium. The soil texture was 26 % sand, 13% clay, and 61% silt, and had the highest runoff potential, with low infiltration rates and shallow soil over nearly impervious material (NRCS USDA WSS 2015). Soil depth ranged from 30-35 cm with a hardpan layer of caliche and basalt present from 35-50 cm. At 18 cm depth in soil there were noticeable caliche pieces ranging from 4-6 cm diameter.

Soil profiles varied between the “sagebrush” (*Artemisia tridentata* spp. *wyomingensis*), “bunchgrass” (*Elymus elymoides*), and bare ground microsite types. In the sagebrush microsite areas, the upper 4 cm were rich with organic material and a thin moss layer. Sagebrush roots were observed just above weathered basalt from 6-11 cm depth. The bunchgrass microsite type had fine roots that proliferate from 2-14 cm depth, with larger roots beginning at 8 cm depth. Bare ground patches had a delicate structural crust with areas of cryptogrammic crust present. Few root structures intersect the profiles of the bare ground microsite type. This area was considered semi-arid desert with an average rainfall of 311 mm /year, a mean annual air temperature of 9 ^{o}C (48 ^{o}F), and received precipitation in pulses, with intense events that sometimes causing regional flash flooding. At this site, soil water resources were generally in deficit from June through October with peak temperature in July (25 ^{o}C) and freezing conditions from December through February (Figure 1.4). ET flux measured from the local eddy flux station was the greatest from March through June and most variable during spring and summer (Figure 1.5). From the measured ET and precipitation data, it appears that over time more water is leaving the system as ET than enters as precipitation (Figure 1.65) (Germino, Allen, Zhao, Rey et al. unpublished data). The growing gap between ET and precipitation may be due to error in the measurements. Instrument error and associated uncertainties are common in water and energy balance studies as described in Allen et al. (2011). Precipitation is likely under measured at this site with wind, intensity, and spatial variability adding challenges to measure this term accurately (Ciach 2003; McMillan et al. 2011). Understanding more about evaporation and transpiration occur in the sagebrush steppe may further efforts to reduce the gap in water and energy balance studies.

Figure 1.2 Location of field site (red star) relative to Boise, ID (yellow star) in the Northern Great Basin (white outline). Map used with permission of USDA Forest Service, Rocky Mountain Research Station, Boise Aquatic Sciences Lab

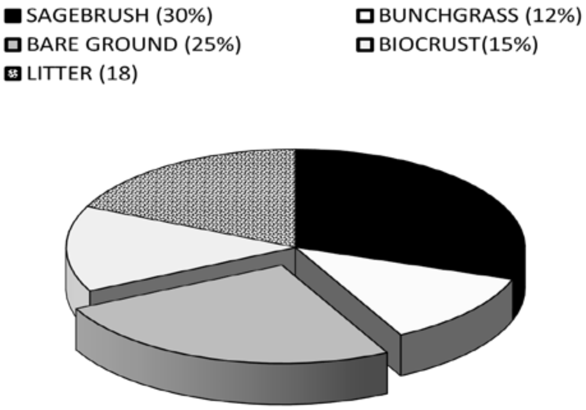

Figure 1.3 Proportion of major vegetation types at the Hollister field site. Sagebrush (30%), Organic Litter (18%), Biological Soil Crust (15%), Bunchgrass (13%), and Bare Ground (25%). Proportion of plant community assessed June 2014.

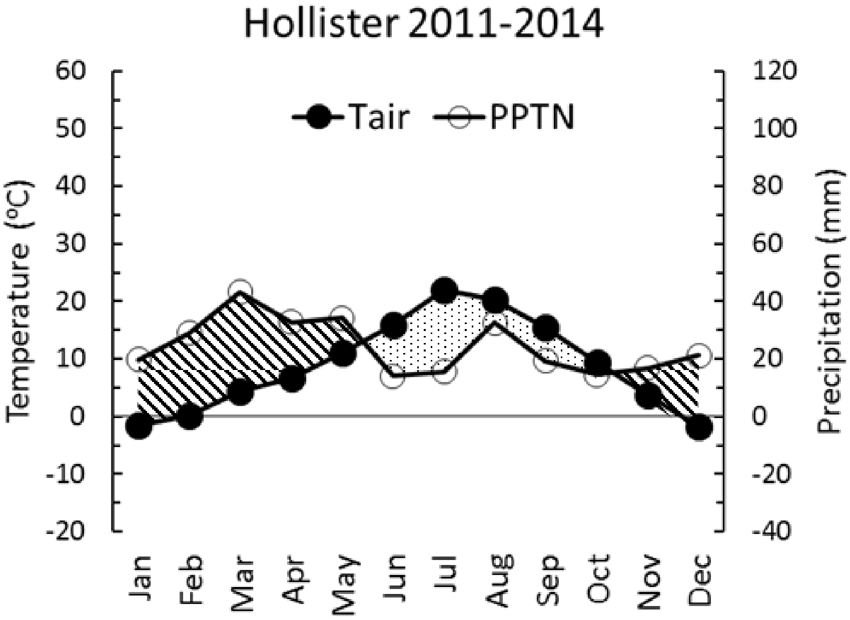

Figure 1.4 Average monthly air temperature (open circles) and total monthly precipitation values (closed circles). Four-year average values were measured at the Hollister Energy Balance Flux Site (Zhao and Allen 2014). The study site is in soil water deficit (lines) from June-October and in surplus (dots) from November-May (Germino, Allen, Zhao, Rey et al. unpublished data).

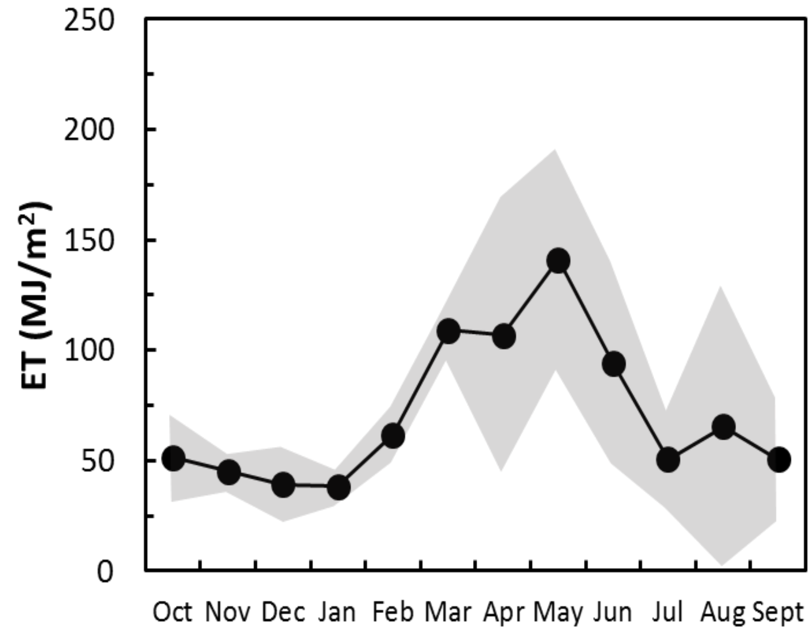

Figure 1.5 Average monthly ET at the Hollister Energy Balance Flux Site for 2011-2014. ET is elevated from March to June with peak values occurring in May. Shading represents the four-year annual variability. ET is most variable at this site from April-June (Germino, Allen, Zhao, Rey et al. unpublished data).

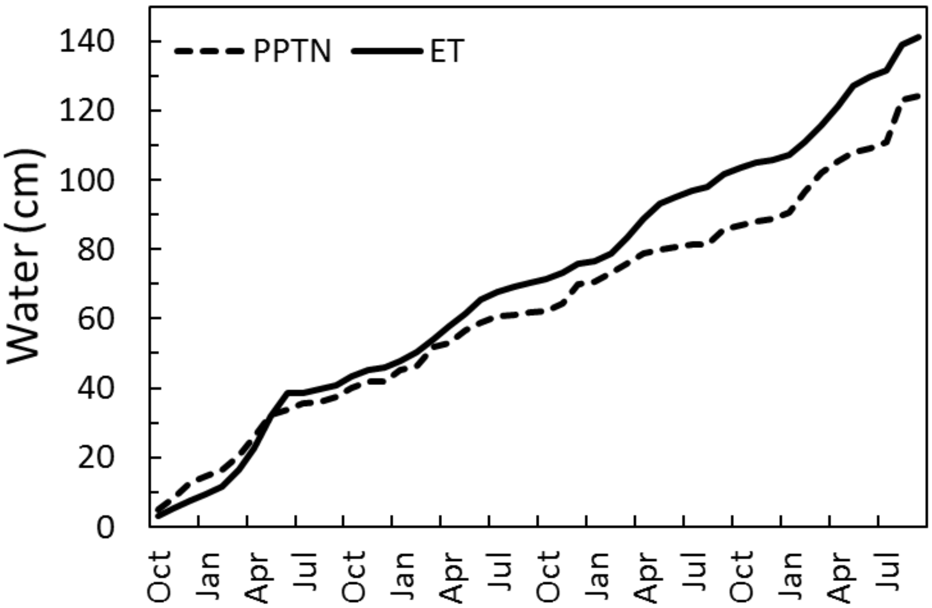

Figure 1.6 A cumulative account of incoming precipitation (PPTN) and outgoing evapotranspiration (ET) over 4 water years 2011-2014 at the Hollister Field Site (Germino, Allen, Zhao, Rey et al. unpublished data).

CHAPTER 2. LIQUID-VAPOR EQUILIBRATION: A METHOD TO MEASURE ISOTOPIC COMPOSITION OF SOIL WATER

2.1 Abstract

In order to employ a liquid-vapor equilibration method to measure isotopic composition of soil water for future experiments, we needed to determine the time it would take for silty-loam soil of various moisture contents to equilibrate. Dry soils are challenging to accurately measure isotopic composition using the liquid-vapor equilibration method. Our goal was to determine how to overcome this challenge with increased volume of soil to ensure the minimum required water in a sample. We tested a silty loam soil with 10%, 5%, and 2% gravimetric water content (GWC) and found time to equilibrate was 3,4, and 5 days respectively. While the 10% and 5% GWC soils reached 100% vapor saturation, the 2% only reached 70 %GWC yet matched the isotopic concentration of our control so we assumed isotopic equilibrium. This is a novel finding for this method, to reach isotopic equilibrium without vapor saturation.

2.2 Introduction

Stable isotopes of water are commonly used in research on plant-soil water relations (Ehleringer and Dawson 1992; Brunel et al. 1995; Phillips and Greg 2003). We are interested in the isotopic composition of soil water and one technique to analyze the isotopic composition is called liquid-vapor equilibration (LVE) or “headspace equilibration” (Wassenaar 2008). This method allows the analysis of isotopic composition without using tedious techniques such as “cold trap” or cryogenic extraction of soil water (Soderberg et al. 2012, Griffis 2013). Rather than extract the water, we take soil from an experimental or natural setting and encapsulate it in plastic where liquid and vapor water may achieve chemical equilibration. Once vapor and liquid are in equilibrium, we analyze vapor water from the headspace and determine the average isotopic composition of soil water in a sample (Hayes 2004).

Saturation of water vapor (RH=100%) means that liquid and vapor are in chemical equilibrium and therefore isotopic equilibrium. Time for a soil sample to reach complete equilibrium is affected by soil texture and water content, so different soil types and ranges of soil moisture should be tested to determine equilibration time. Measuring isotopic composition of water vapor at lower water contents is challenging because there is less water available for headspace saturation and for the liquid and vapor to be in equilibrium. Sandy soils with a low water content (<5% by weight) proved challenging but may be rectified by using larger soil volumes (Wassenaar 2008 and personal communication). Increasing the amount of soil in a sample will increase the amount of water needed to fill the headspace. The goals of this study were to; determine if our dry soil could be isotopically analyzed by increasing the amount of soil and to provide practical guidelines on LVE, for a silty loam soil.

2.3 Methods

To determine time to equilibration for soils of varying moisture contents, we created replicated soils of known gravimetric water content (GWC) and analyzed the headspace vapor every 24 hours. Analysis of isotopic composition and saturation concentration of the vapor in the headspace are used to illustrate liquid-vapor equilibrium. Soil water contents tested (2%, 5%, and 10% GWC) were chosen based on typical low values for this semi-arid ecosystem. Water of known isotopic composition was added to dried room temperature soils, and mixed thoroughly. Three replicates for each soil moisture value at each sampling interval (n=3) were created. To ready samples for isotopic analysis, prepared soils were placed in a quart-sized plastic bag and inflated with dry air, then checked for leaks to ensure a closed system. Bags were placed in an isothermal (20^{o}C) and isobaric (90 kPa) environment to equilibrate. Each soil moisture type was sampled every day over a 7-day period. For the headspace vapor to achieve saturation as required for analysis, the amount of soil in the bags was adjusted to have 20 mL of soil water per bag, as suggested by Wassenaar et al. (2008). At the average temperature and atmospheric pressure of the environment, saturation vapor pressure was about 3 kPa or 33 000 ppm.

Isotopic composition was measured by penetrating the headspace of the plastic bag with a needle attached to a 0.5 m silicon tube. Vapor was pulled directly from the sample bag into the cavity of the Picarro L2130-I water analyzer with a continuous flow and measurements taken every 2 seconds. Final values are reported with a moving average of 2 minutes. After isotopic composition of vapor water (R_{smp_v}) is measured (Eq 2.1) it is then related to the known standard concentration of Vienna Standard Mean Ocean Water (VSMOW) (R_{vsmow}) and reported in standard delta notation δ_{smp_v} with units of permil (‰) using equation 2.2.

$$R_{smp_v} = [^{18}O]/[^{16}O] \qquad \text{eq 2.1}$$

$$\left[\left(\frac{R_{smp_v}}{R_{vsmow}}\right) - 1\right] * 1000 = \delta_{smp_v} \qquad \text{eq 2.2}$$

The isotopic composition of the liquid water (δ_{smp_l}) was calculated using the temperature dependent liquid-vapor fractionation factor (α) equations derived by Horita and Wesolowski (1994) for both ^{18}O (Eq 2.3) and ^{2}H (Eq 2.4). In these equations e is the natural log base and T(^{o}K) is the average laboratory temperature during the time of equilibration. The liquid-vapor equilibrium relationship may be calculated using the appropriate fractionation factor (Eq 2.5).

$$\alpha_{l-v}(^{18}O) = e^{\left(\frac{-7.685+6.7213(10^3T)-1.6664(10^6T^2)+0.35041(10^9T^3}{10^3}\right)} \qquad \text{eq 2.3}$$

$$\alpha_{l-v}(^{2}H) = e^{\left(\frac{1115.8(10^9T^3)-1620.1(10^6T^2)+794.84(10^3T)-161.04+2.9992(10^9T^3)}{10^3}\right)} \qquad \text{eq 2.4}$$

$$\delta_{smp_l} = \alpha * \delta_{smp_v} \qquad \text{eq 2.5}$$

In addition, control samples were made using 20 mL of labeled water and placed in a bag with no soil (GWC= 100%). When isotopic composition of the soil matches that of the control or remains constant, we assume equilibrium was achieved. Data from Wasenaar (2008) demonstrated that pure water should take only 10 minutes to equilibrate. Previous efforts to experimentally determine time to equilibration for wetter soils were more certain of liquid-vapor equilibration due to visible saturation of the headspace. Soil moisture was calculated after isotopic analysis was performed to confirm GWC and amount of soil water available in sample.

2.4 Results

Measured GWC confirmed that all bags contained at least 20 mL of water. Water used in this experiment had the isotopic composition of δ^{18}O -17.60‰ and δ^{2}H -129.84‰ and average laboratory temperature was 20.5^{o}C ±0.2. Under the assumption that 33 000 ppm air is considered saturated with vapor, the 10% soil samples reached saturation by day 3 Figure 2.1d), 5% GWC reached saturation by day 4 (Figure 2.1e), but the 2% GWC soil only reached a maximum of 23 000 ppm on day 6. (Figure 2.1f). The headspace for the 2% soil did not become saturated but the isotopic composition matched the control series (Figure 2.1c). The 10% and 5% GWC series have a more enriched isotopic composition throughout time relative to the $\delta^{18}O$ of the control and did not approach the isotopic composition. While the wetter soils did not match the isotopic composition of the control they did have saturated vapor concentration after 3 and 4 days respectively. The 2% GWC approached the isotopic composition of the control at 5 days, but did not reach the water concentration of the control. (Figure 2.1). Based on the saturation of vapor for the 10% and 5% GWC soils and the isotopic composition of the 2% soils, we will say that these soils reached equilibrium at 3,4, and 5 days respectively. The isotopic composition of drier, 2% and 5% GWC, soils were dependent on concentration, while the 10% and 100% GWC were not (Figure 2.2).

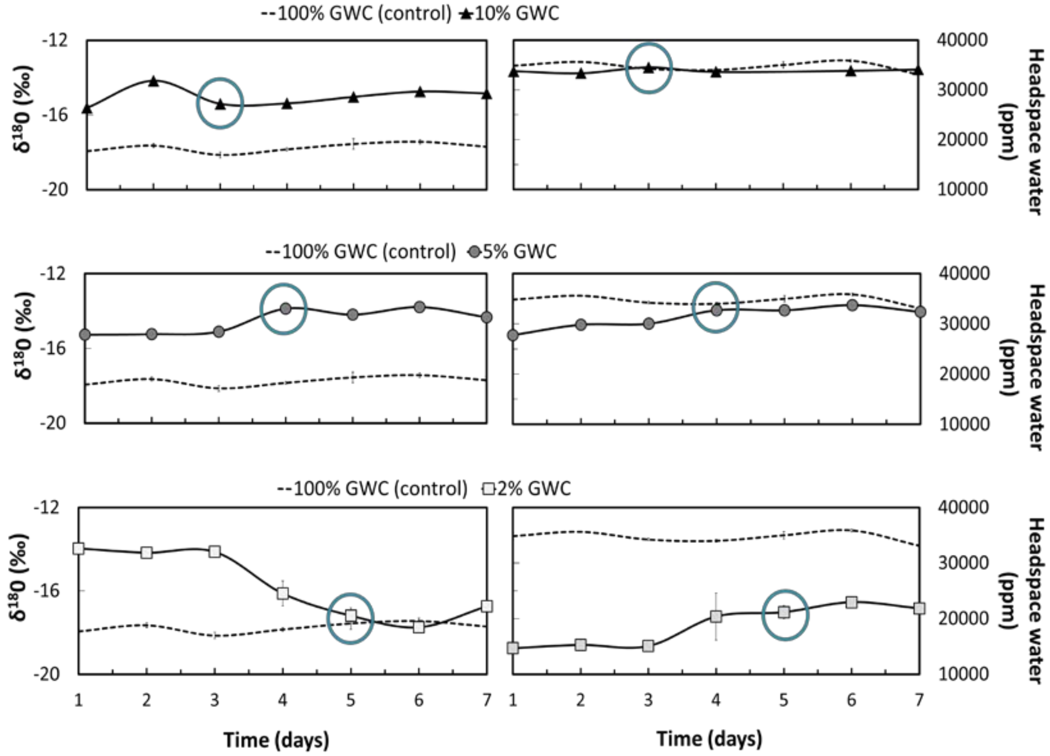

Figure 2.1. (a-f). Isotopic composition (a-c) and water vapor concentration (d-f) of soil with three different water contents in time. Values are compared to the 100% GWC (control) dashed line. The 10% GWC (black triangle) and 5% GWC (grey circle) did not approach the isotopic composition ($\delta^{18}O$) of the control, but matched the water concentration (ppm) after 3 and 4 days respectively. The 2% GWC (white squares) approached the isotopic composition of the control at 5 days, but did not reach the water concentration of the control. Moment of interpreted equilibrium is circled.

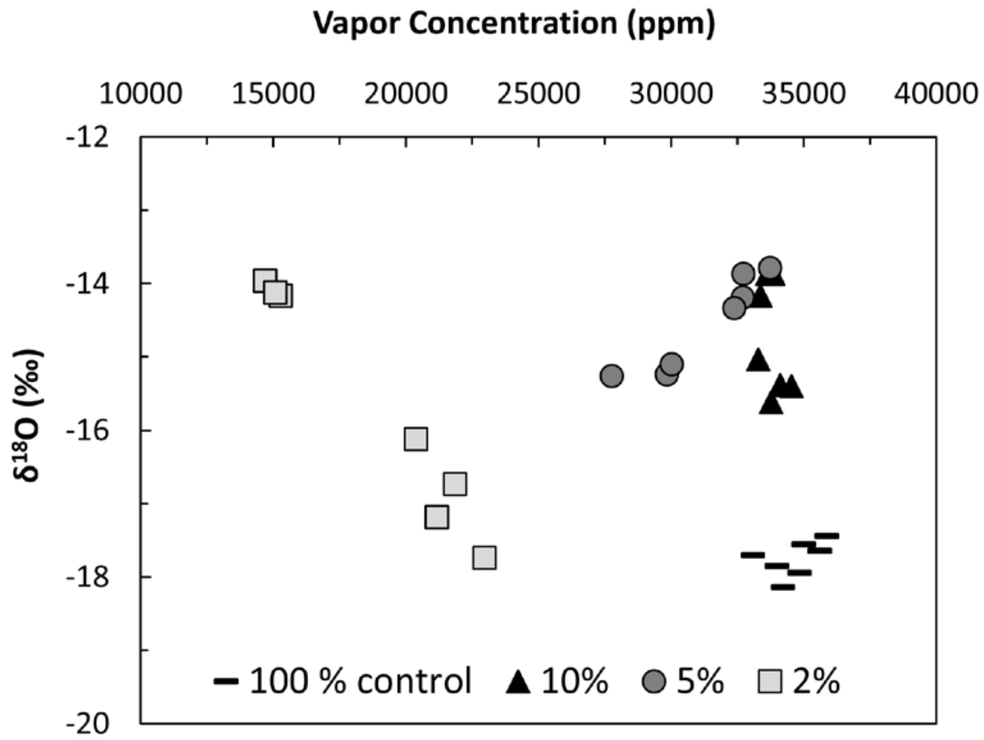

Figure 2.2 The control samples (dash) with only water (100% GWC) cluster in the bottom right saturated and the correct isotopic composition. The 10% GWC samples (triangle) also cluster but in the upper right, at saturation but with a different isotopic composition than the control. Both the 5% GWC (circle) and the 2% GWC (square) soils are not clustered, but trend with changing isotopic value with concentration (R^2=.89, R^2=.97 respectively).

2.5 Discussion

Each soil sample had the minimum 20 mL water in the bag and all but the 2% GWC reached the estimated saturated vapor concentration of 33 000 ppm during the experiment. While the 10% and 5% GWC soils reached 100% vapor saturation by 3 and 4 days respectively (Figure 2.1d-e), the drier soil (2% GWC) only reached 70 % saturation (23 000 ppm). The 2% GWC soils matched the isotopic concentration of our control at 20 000 ppm, so we assumed isotopic equilibrium by day 5 (Figure 2.1f). The isotopic composition of the 5% and 10% GWC soils did not match the control value but remained constant, suggesting equilibrium values (Figure 2.1a-b). Soil water in the 5%

and 10% samples may already have experienced evaporation during preparation causing the isotopic composition to be elevated relative to the control from the beginning. Isotopic values did not increase nor did vapor concentration decrease with time, suggesting that samples remained sealed and the closed system intact. Even though the driest soil (2 % GWC) headspace vapor was likely unsaturated during the experiment it nonetheless appeared sufficient to use in stable isotope studies. This is a novel finding for this method, to reach isotopic equilibrium without vapor saturation.

Overall, this experiment provides a guide on time required for a silty loam soil across a range of moisture contents to reach liquid-vapor isotopic equilibrium. This will be helpful in studies to partition ET on the landscape, where natural soils are collected for study and isotopic composition of soil water is of concern.

CHAPTER 3. PARTITIONING EVAPOTRANSPIRATION

3.1 Abstract

Evapotranspiration (ET) is a major part of ecosystem water balance, and methods to evaluate relative contributions of evaporation (E) and transpiration (T) in ET are needed. Isotopic fractionation can distinguish soil water loss due to evaporation from that of plant uptake, possibly providing means to assess E and T in retrospect rather than directly measuring ET fluxes. We tested the ability of a simplified stable isotope and mass balance model following Wenninger et al. (2010) to predict 100% evaporation in a laboratory soil microcosm. Under controlled conditions with no plants available to remove soil water for T, the simplified model resulted in 99% E (+/- 4.3%). This was the first experiment to test this model under controlled conditions and results will help guide field applications to partition ET in to E and T.

3.2 Introduction

Soil water loss to the atmosphere as evapotranspiration (ET) is generally the largest loss in the water balance of semi-arid ecosystems. ET is composed of evaporation (E) and transpiration (T), and changes in climate or vegetation can shift the balance of T relative to E. Little is understood about what portion of ET is E relative to T in the semi-arid ecosystems (Wilcox and Thorough 2006), and an easy to use model to partition these components will improve water balance models, water loss prediction, and assessment of impacts of vegetation change due to disturbances such as fire and invasive species.

Conservation of both mass and isotopes allows the use of a stable isotope-mass balance model to partition soil water loss to the atmosphere into E and T. Evaporation causes fractionation of isotopes, and the process of evapofractionation is isotopic fractionation due to evaporation. During evapofractionation lighter isotopes vaporize faster and residual soil water becomes enriched in heavy isotopes. In contrast, soil water uptake by plants for transpiration does not fractionate water (Wershaw, Friedman, and Heller 1966, White et al. 1985, Ehleringer and Dawson 1992, Zhang et al. 2010). Isotopic enrichment of soil water combined with decreased water content should indicate evaporation. Depletion of soil water not accompanied by isotopic enrichment indicates transpiration, assuming no drainage or runoff.

Isotope hydrology and mass balance models using soil, plant, and air water have been used to infer differences in E and T on the landscape level (Hsieh et al. 1998; Robertson and Gazis 2006; Wenninger et al 2010; Zhang et al. 2010), but this approach has not been tested or verified under simplified conditions without vegetation, where infiltration, storage, runoff, and ET are each controlled. To distinguish soil water loss between E and T, the product of isotpoic composition and mass fraction of residual soil water before and after ET has occured is calculated. The difference in the products is the amount of water lost due to E.

We tested the ability of the simplified Wenninger et al (2010) model to predict ratios of E:ET using a microcosm of bare soil in which all water lost was only possible to E. All other means for water loss were controlled for. We predict the simplified model will result in 100% E and 0% T. Successful demonstration of this simplified isotope mass balance model would be an important step towards establishing validity of this method and application to other field studies.

3.3 Methods

We used open top plastic containers with no vegetation to create a microcosm in which the only soil water loss would be evaporation. Each container (945 cm^3) was mixed with a loess and mixed alluvium derived 26% sand, 13% clay, 61% silt soil and wet to a gravimetric water content (GWC) of 25%. Soil was wetted to ensure enough water as required for isotopic measurement after evaporation. Three containers were used as the initial condition (t_1) samples. Three other containers were placed under a 1000 Watt solar lamp with constant convection from a fan and were sampled 12 hours later (t_2). Containers were devoid of any vegetation and had no water inputs, surface runoff, infiltration, or transpiration between t_1 and t_2, so all water lost was due to E.

To sample soils, each container was cut open along the side to allow extraction of 2 cm layers from 0-10 cm. Samples were bagged and inflated with dry air, then placed in an isothermal environment for 48 hours to achieve liquid-vapor equilibration. This method requires the headspace to become fully saturated (RH=100%), for liquid and vapor water to be in isotopic equilibrium. Time to equilibration was experimentally determined based on soil type and estimated soil moisture, following Wassenaar (2008). An equilibrium fractionation factor was calculated using temperature of the controlled environment, and used to convert the isotopic composition of vapor water into liquid (Chapter 2, equations 2.3-2.5).

Isotopic composition of vapor water from the headspace was measured using a Picarro L2130-*I* water analyzer, reporting values in delta permil (δ ‰) notation, relative to Vienna Standard Mean Ocean Water (VSMOW) following Hayes (2004). A Temperature dependent liquid-vapor fractionation factor (α) was used following Horita and Wesolowski (1994) to calculate the isotopic composition of liquid in equilibrium with measured vapor water (refer to chapter 2 methods). Kinetic fractionation due to temperature gradients in the soil were not considered in this study since the focus is not on water fluxes, but rather proportions of water loss inferred from residual soil water following evaporation. Previous studies partitioning E:ET only used oxygen isotopes (Hsieh et al. 1998; Robertson and Gazis 2006; Wenninger et al. 2010), because deuterium does not improve results when partitioning ET (Haverd et al. 2011). While this study only uses oxygen isotopes in the following calculations, we report both δ^{18}O and δ^{2}H values. The original equations as presented by Wenninger et al. (2010) are described (Eq 3.1-3.6) along with the simplified equations this study offers (Eq 3.7-3.8). The mass balance model was simplified by removing water inputs and outputs that were experimentally eliminated, specifically downward infiltration (*z*), incoming precipitation (*r*), and surface runoff.

$$m_i + m_r = m_f + m_e + m_{nf} + m_z \qquad \text{Eq 3.1}$$

$$x_j = \frac{m_j}{m_T} \qquad \text{Eq 3.2}$$

The mass of each water component (*m*) is denoted with the respective subscript *j* = (*i, f, r, e, nf, z*) initial, final, rain, evapofractionated, non-fractionated, or downward-infiltrated water. The fraction of water for each component (*x*) is the ratio of that component's mass relative to the total mass of water (T) in the soil column (Eq 3.2). The overall stable isotope mass balance equation is the summed product of each model component's δ^{18}O composition and mass fraction relative to the total water (Eq 3.3). The

δ_{nf} term was algebraically factored with the assumption that the isotopic composition of non-fractionated water loss was equal to that of the infiltrated water, which is represented by the average of the initial and final isotopic composition (Eq 3.4).

$$x_i\delta_i + x_r\delta_r = x_f\delta_f + x_e\delta_e + x_{nf}\delta_{nf} + x_z\delta_z \qquad \text{Eq 3.3}$$

$$\delta_{nf} = \delta_z = \frac{\delta_i+\delta_f}{2} \qquad \text{Eq 3.4}$$

The fraction of water that is evapofractionated (x_e) and non-fractionated (x_{nf}) is the fraction of evaporation and transpiration respectively (Eq. 3.5-3.6). There is no downward infiltration (x_z=0) or transpiration/runoff (x_{nf}=0).

$$x_e = \frac{x_i\delta_i+x_r\delta_r-x_f\delta_f-(x_{nf}+x_z)\delta_z}{\delta_e} \qquad \text{Eq 3.5}$$

$$x_{nf} = x_r + x_i - x_e - x_f - x_z \qquad \text{Eq 3.6}$$

In equation 3.5, the δ_r and x_r terms are set to zero making the terms $x_r\delta_r$ zero. With the added interpretation, that δ_z is zero; $(x_{nf}+x_z)\ \delta_z$ becomes zero. Mass fraction of soil water for the initial and final time step were an average of the respective microcosm profile, and all replicates were averaged across each layer in the profile (n=3). Initial isotopic composition (δ_i) of liquid soil water was averaged across the profile. Within the profile, the maximum isotopic value, or evaporative front was identified and both liquid and vapor water isotopic δ^{18}O values at this location were used in the mass balance calculations. The measured vapor composition from the evaporative front was used for δ_e and the calculated liquid value was used for δ_f. We solved for the fraction of water lost due to evaporation (x_e) and transpiration (x_{nf}) in our experiment without incoming water, infiltration, runoff, or transpiration, using the simplified equations 3.7-3.8. Once we have a mass fraction of water evaporated, the remainder of mass used for non-fractionated processes (transpiration) is calculated and converted to depth of water (mm). Using the fractions of water lost to evaporation and transpiration the ratio of E:ET is calculated using equation 3.9.

$$x_e = \frac{x_i\delta_i - x_f\delta_f}{\delta_e} \qquad \text{Eq 3.7}$$

$$x_{nf} = x_i - x_e - x_f \qquad \text{Eq 3.8}$$

$$E: ET = \frac{x_e}{x_e + x_{nf}} \qquad \text{Eq 3.9}$$

Error was propagated through equations 3.7-3.9 using the standard error (σ) of each mass fraction and isotopic value. Error in the final equation 3.9 was calculated using the addition and multiplication of measured quantities as shown in equation 3.10.

$$\sigma E: ET = |E: ET| \sqrt{\left(\frac{\sigma x_e}{x_e}\right)^2 + \left(\frac{\sigma(x_e + x_{nf})}{(x_e + x_{nf})}\right)^2} \qquad \text{Eq 3.10}$$

Our approach assured that under initial conditions (t_1), soil water pools at any depth within the profile were homogenous, meaning water and its isotopes were evenly distributed within a soil depth. We assume mass balance does not distinguish between mobile or immobile soil pore water and all water in soil is exchangeable.

3.4 Results

Both the water content and isotopic profiles showed water loss and evapofractionation through the whole profile (Figure 3.1a-c). The average initial GWC of the soil in the container was 25%, and after 12 hours of continuous evaporation the average GWC was 17% (Figure 3.1a). Almost 30% (28.9%) of the water (3.6 mm) was lost during the sampling interval.

Average initial and final liquid δ^{18}O values in the soil profile were -16.63‰ ±0.25 and -14.58‰ ±0.26, respectively (Figure 3.1b). Average initial and final liquid water δ^2H composition in profile was -129.8‰ ±0.19 and -128.3‰ ±0.52 respectively. Average soil water δ^2H values from 0-2 cm and 2-4 cm at t_1 fell within the standard deviation of t_2 values (Figure 3.1c), which indicates no change over time. The residual water in surface soils (0-2 cm) became enriched in δ^{18}O by 3‰ and depleted in δ^2H by 1‰. The maximum isotopic value (evaporative front) was identified between 0-2 cm according to δ^{18}O values, but at 8-10 cm for δ^2H. (Figures 3.1 b-c). Final isotopic composition of δ^{18}O vapor water at the evaporative front was -21.86‰ ±0.26 and liquid water was -14.58‰ ±0.15. Based on these isotopic outcomes, the estimated proportion of water lost to evaporation was 99% ±4.3%, giving a ratio of 99:1, E:ET.

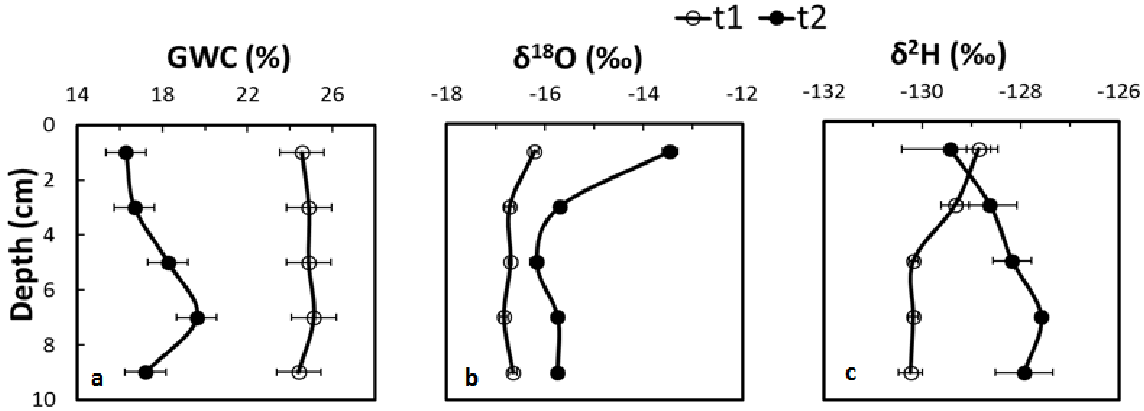

Figure 3.1 (a-c). Water lost during 12-hour microcosm experiment, between initial t_1 (open circles) and final t_2 (closed circles). Water loss is described in three ways: (a) Gravimetric Water Content (GWC), (b) isotopic enrichment in δ^{18}O, and (c) δ^2H.

3.5 Discussion

We predicted 100% E and 0%T due to blocked runoff, transpiration, and infiltration of water from occurring in subject soils, but observed 99% E and 1% T. The t_2 isotopic profile of δ^2H values at t_2 in Figure 3.1c do not follow the same pattern of t_2 surface δ^{18}O values being more enriched in Figure 3.1b. The lack of δ^2H enrichment at the surface is likely due to tendency of the H to exchange within the molecule being less than the O. The preferential order of isotope replacement is ^{16}O, ^{18}O, and ^{2}H, so the δ^2H values will change slower compared to δ^{18}O. If the experimental duration were longer, we might see the δ^2H values more enriched near the surface like the δ^{18}O profile.

Error in the E:ET ratio according to equation 3.10 is 4.3 %, which encompasses standard error in replication of measured values. This error may be due to evaporation from the sides as soil began to separate from the container. Soil shrinkage during evaporation should be considered and methods revised for future controlled lab experiments, such as a seal to prevent air flow from sides. The t_1 δ^2H profile suggests evapofractionation was already occurring in the microcosm before the first soil measurements and t_2 profile suggests that condensation may have occurred at soil surface (Figure 3.1c). Kinetic fractionation due to temperature gradients in the soil was not considered in this study since the focus was not on water fluxes, but proportion of water lost. In future studies, thermocouple data and a kinetic fractionation factor could be used to improve calculation of evaporation rates.

We observed the predicted profile evolution of soil water loss and isotopic composition forming the basis of this research as discussed in Chapter 1. The simplified mass balance equation we used appears to accurately predict water efflux due to E when no runoff, infiltration, or transpiration occurs, but methods may need to be refined with a more comprehensive microcosm to explore other model inputs. A generalized model will offer a step towards application to a larger scale and the subsequent thesis chapter discusses application of this technique to partition evapotranspiration in a natural setting.

CHAPTER 4. LANDSCAPE APPLICATION: PARTITIONING EVAPOTRANSPIRATION ACROSS MICROSITES IN THE SAGEBRUSH STEPPE

4.1 Abstract

Evapotranspiration (ET) is a major part of the water balance in semi-arid ecosystems. Changes in climate and vegetation patterns will shift the proportion of evaporation (E) and transpiration (T), but few studies have quantified their relative contributions to ET. Isotopic fractionation distinguishes soil water loss due to evaporation from that of plant uptake used for transpiration. This provides a means to assess E and T in retrospect rather than only measuring ET fluxes. In June and September 2014, we evaluated ratios of E:ET in sagebrush, bunchgrass, and bare ground microsites at a sagebrush steppe field site in southern Idaho. We expected that soil water removal would occur only through E on bare ground and for bunchgrass when senesced in September. Soil water loss and changes in isotopic composition from 0-10 cm were measured in a 72-hour time series following wetting, and E and T were quantified using a simplified isotope mass balance model following Wenninger et al. (2010). Sagebrush used 5% of the soil water in the upper 10 cm for transpiration in June and 18% in September. The bunchgrass used 8% of near-surface soil water in June, but did not use any in September. Only evaporation and no transpiration occurred on bare ground microsite. This method appears to be successful in measuring E:ET ratios in retrospect. The use of both components E and T may further understanding of water loss in the sagebrush steppe as plant communities continue to change.

4.2 Introduction

Evapotranspiration (ET) is the largest annual water loss in semi-arid regions of the world (Wilcox and Thurow 2006), with upwards of 100% or more of incoming precipitation leaving as ET (Flerchinger and Cooley 2000; Germino and Allen unpublished). It seems imperative to understand contributions of evaporation (E) and transpiration (T) in semi-arid environments as plant communities are changing across a landscape, shifting the amount of bare soil and plant canopy gaps. Little is understood about the importance of E and T in the sagebrush steppe ecosystem (Wilcox and Thurow 2006); and less is known about these contributions at the microsite scale. We expect changes in E relative to T to be an important ecosystem impact. Methods to quantify ET exist but few studies have quantified the relative contributions of E and T to ET.

Stable isotope hydrology can help distinguish soil water loss between evaporation and plant water uptake for transpiration. Evaporation causes isotopic fractionation of soil water but plant uptake for use in transpiration does not (Wershaw, Friedman, and Heller 1966, White et al. 1985, Ehleringer and Dawson 1992). During evaporation, lighter isotopes vaporize faster and residual soil water becomes enriched in heavy isotopes. Evapofractionation is isotopic fractionation due to evaporation. Enrichment of heavy isotopes at the soil surface combined with soil water loss should indicate evaporation. Soil water loss not accompanied by isotopic changes indicates plant uptake for transpiration, assuming no drainage or runoff.

In the semi-arid sagebrush steppe we expect most hydrologic fluxes occur near the surface (Mathieu and Bariac 1996a) and infiltration to deeper soil only occurring during large precipitation events and spring snow melt (Schwinning and Sala 2004; Gazis and Feng 2004). Shifts in precipitation and temperature will result in changing vegetation and surface conditions. ET will fluctuate with landscape changes as soil moisture varies with time, disturbance, and changing vegetation (Neilson and Marks 1994; Obrist et al. 2004). With the specialized interactions between plants, soils, and water in the sagebrush steppe, it is important to understand how contributions of E and T shift with changing landscapes.

Tools available to measure or model water loss as ET from the landscape include the eddy flux method and numerical surface flux models (Gutiérrez et al. 2006). These methods generally do not separate E from T and furthermore have issues with energy balance closure in which energy inputs and outputs may have disagreement up to 30% (Allen et al. 2011). The problem may be scale of measurements and heterogeneity of the landscape (Foken 2008). Few studies have tried to quanitfy contributions of E and T from different vegetation and surface types across the landscape, microsites may differ in soil water lost to the atmosphere at both temporal and spatial dimensions. We can measure the isotopic composition of soil water within a profile and know the depth of isotopic maximum, or the evaporative front, where the most evaporation has occurred using methods pioneered by Barnes and Allison (1988) and Mathieu and Bariac (1996a, b). Isotopic composition of soil water at the evaporative front is an important parameter in the stable isotope mass balance model used to partition E and T (Rothfuss 2010). The ability to partition E:ET using $\delta^{18}O$ in soil water was first demonstrated with a course resolution of 10 cm increments by Hsieh et al. (1998). Robertson and Gazis (2006) and Wenninger et al. (2010) provide alternative solutions for fraction of water loss due to E and assumed isotopic composition of non-fractionated (transpired) soil water. We choose to follow Wenninger et al. (2010) but offer a simplified model for a condition in which runoff precipitation and deep infiltration are eliminated.

Our objective was to retrospectively quantify contributions of E and T from three microsite types in the sagebrush steppe ecosystem: sagebrush, bunchgrass, and bare ground, that vary strongly in E and T providing ideal differences to test our model at two different times during the year. We predicted that evergreen sagebrush would always be transpiring (E:ET <1), whereas bare soil would only have evaporation (E:ET =1), and bunchgrass would exhibit transpiration in June (E:ET <1) but not when senesced in September (E:ET =1). Following these predicted scenarios, we also predict that when surface soils are experiencing water loss due to both E and T gravimetric water content (GWC) would be less than if water loss was due to only E. Differing proportions of water loss will also have different infiltration rates across surface soils of microsites and across season.

4.3 Methods

To partition ET at the microsite scale, we use a simplified isotope mass balance model previously tested under a controlled environment designed to predict 100% E. We obtained E:ET by adding water to microsite plots in both June and September and measuring the isotopic composition of soil water in a 72-hour time series. To measure isotopic composition of the soil water, we employed a liquid-vapor headspace equilibration technique (Wassenaar 2008). This method was particularly helpful, as soil water in an arid ecosystem often exists in both vapor and liquid form near the surface. Soil texture and type was designated according to the Bouyoucos (1963) hydrometer method and the NRCS USDA Web Soil Survey respectively.

Experimental Design

To adequately represent the vegetation in the area, 20 -1 m^2 sampling areas were randomly spaced within the 250 m^2 study area. These sample areas were photographed directly downwards from 1.5 m above and were independent of any other sample areas. To measure the proportion of vegetation we used the nadir imagery and SamplePoint software developed by the USDA (Booth et al. 2006) to assign a 100 point grid over the image. With this grid, we calculated percent cover (%). The microsite array for soil water manipulation (Figure 4.1) was created for each microsite type to have had 10 random sampling locations, divided in half for the beginning and end of the 72-hours (n=5). For each June and September, sampling event a new microsite array of 30 locations was created such that no microsites were sampled twice. Random samples located near the road were moved. Samples microsites were defined as: “sagebrush”, Wyoming big sagebrush (*Artemisia tridentata wyomingensis*); “bunchgrass”, Bottlebrush Squirreltail (*Elymus elymoides*); and “bare ground”, which contained no plants and any underground root connections were severed with a thin blade saw blade. Effort was made to sever roots to at least 10 cm depth.

We created a 30 microcosm array using a 20 cm diameter PVC cylinder tapped into the ground to at least 4cm depth and focus infiltration of 1.5 L of labeled water down into the soil. The cylinder improved soil saturation of the 4 710cm^3 column to 11cm depth with an initial GWC of at least 20%. Saturating the soil created a uniform isotopic and GWC condition from 0-11 cm depth. Water was added to soil columns at pre-dawn to reduce evaporative potential simulating a natural saturation event followed by constant ET with no precipitation added between the beginning (t1) and end (t2) of the 72 hours.

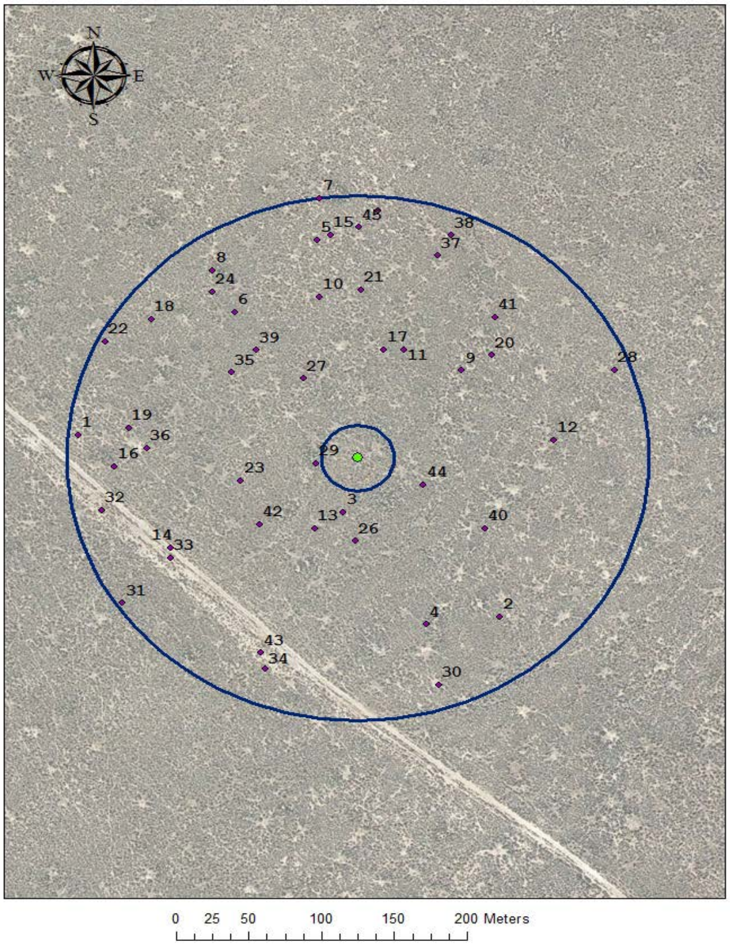

Figure 4.1. Randomized points in a 250m^2 area surrounding the eddy flux tower at the Hollister Sagebrush Energy Balance site, in the center. Each point was assigned a microsite type and flagged for the manipulated experiment. Only 30 points were used at each sampling event and points were repositioned for the September sampling event such that no point was sampled twice

Soil columns were sampled for GWC and isotopic composition immediately after saturation for an initial (t_1) profile. Three days later, the remaining microsites were sampled for a final (t_2) profile. The wetting front was not identified because it was unnecessary for the model. The assumption is that the entire soil profile has been wetted, with the added water infiltrated at t1, and the experiment begins at this point. These soil water profiles in series were used to calculate soil water loss and isotopic composition for input to the mass balance equations. The soil profile of non-irrigated samples (t_0) was also collected before the experiment began to inform the natural ambient soil moisture and isotopic composition. Depths of samples in the soil column were chosen to efficiently sample yet adequately measure the isotopic composition at the evaporative front. To capture the isotopic signal of the evaporative front each sample above and below should be no farther than 2 cm (Rothfuss et al. (2010); Dubbert et al. 2013). Several studies and personal observation suggest the evaporative front within the upper 10 cm (Mathieu and Bariac 1996a; Yamanaka and Yonetani 1999; Heitman et al. 2008). Soil samples were taken in 2 cm increments and are as follows: 0-2, 2-4, 4-6, 7-9, and 9-11 cm. A knife and hand shovel were used to excavate each horizon into a plastic bag. Samples were sealed and placed in an isothermal container and transported to the laboratory for isotopic analysis.

Rain catchment jars were placed in the ground to capture any precipitation, even though there was no forecast for rain, and used mineral oil to reduce evaporation of any collected water. Energy balance fluxes (net radiation, sensible heat, latent heat (ET), and soil heat), air and soil temperatures, and precipitation data were collected from the eddy flux tower at the Hollister Sagebrush Energy Balance Site (Zhao and Allen 2014). 72 hour totals of these data were calculated (Table 4.2) and used to characterize climate and energy fluxes during this experiment and get an average evaporation from the landscape to compare to evaporation from manipulated microsite.

The Model

This model uses the difference in water content and isotopic composition between two points in time, regardless of prior conditions, to solve for proportion of water loss due to E and T. The original equations as presented by Wenninger et al. (2010) are described (Eq 4.1-4.6) along with the simplified equations, this study offers (Eq 4.7-4.8). The mass balance model was simplified by removing controlled terms, downward infiltration (*z*), incoming precipitation (*r*), and surface runoff. The mass of each water component is represented by the *m* term with initial, final, rain, evapofractionated, non-fractionated, or downward infiltrated denoted by the respective subscript *j* = (*i, f, r, e, nf, z*).

$$m_i + m_r = m_f + m_e + m_{nf} + m_z \quad \text{Eq 4.1}$$

$$x_j = \frac{m_j}{m_T} \quad \text{Eq 4.2}$$

The fraction of water for each component (x) is the ratio of that component's mass relative to the total mass of water (T) in the soil column (Eq 4.2). The overall stable isotope mass balance equation is the summed product of each model component's $\delta^{18}O$ composition and mass fraction relative to the total water (Eq 4.3).

$$x_i\delta_i + x_r\delta_r = x_f\delta_f + x_e\delta_e + x_{nf}\delta_{nf} + x_z\delta_z \quad \text{Eq 4.3}$$

Isotopic composition of soil water removed by transpiration (δ_{nf}), a process which does not fractionate water, is determined from equation 4.4.

$$\delta_{nf} = \delta_z = \frac{\delta_i + \delta_f}{2} \quad \text{Eq 4.4}$$

The fraction of water that is isotopically evapofractionated (x_e) and non-fractionated (x_{nf}) is the fraction of evaporation and transpiration respectively (Eq. 4.5-4.6).

$$x_e = \frac{x_i\delta_i + x_r\delta_r - x_f\delta_f - (x_{nf} + x_z)\delta_z}{\delta_e} \quad \text{Eq 4.5}$$

$$x_{nf} = x_r + x_i - x_e - x_f - x_z \quad \text{Eq 4.6}$$

Replicates were averaged across each layer in profile (n=3) for a microsite representative profile of water content and isotopic values. Total mass (m_T) of soil water for the initial and final time steps were an average of the respective microcosm profile. The $\delta^{18}O$ isotopic maximum of soil water in the profile, or evaporative front was identified and both liquid and vapor values were used in calculations. The measured vapor composition from the evaporative front was used for δ_e. Isotopic composition of liquid soil water was averaged from the whole microcosm profile for δ_i, but for the final time step δ_f isotopic composition of the evaporative front was identified and used. Averaging $\delta^{18}O$ values in profile to use for δ_e and δf may not give real results so the value at the active point of evaporation is necessary.For this study we experimentally controlled for *r* and *z*, eliminating those parameters from the model and solved for x_e and x_{nf} using simplified equations 4.7-4.8. The mass fraction of water evaporated was determined, and then the mass fraction of water transpired was solved for. Using the fractions of water lost to evaporation and transpiration the ratio of E:ET was calculated using equation 4.9.

$$x_e = \frac{x_i\delta_i - x_f\delta_f}{\delta_e} \qquad \text{Eq 4.7}$$

$$x_{nf} = x_i - x_e - x_f \qquad \text{Eq 4.8}$$

$$E{:}ET = \frac{x_e}{x_e + x_{nf}} \qquad \text{Eq 4.9}$$

Error in E:ET was calculated using the standard error (σ) of $x_e$and x_{nf} and the addition and multiplication of measured quantities principle (Eq 4.10).

$$\sigma E{:}ET = |E{:}ET|\sqrt{\left(\frac{\sigma x_e}{x_e}\right)^2 + \left(\frac{\sigma(x_e + x_{nf})}{x_e + x_{nf}}\right)^2} \qquad \text{Eq 4.10}$$

<u>Determining Isotopic Composition in Laboratory</u>

All isotopic values were measured using the Picarro L2130-I water analyzer at the USGS FRESC in Boise, ID. Isotopic composition of vapor water from the headspace is reported in delta permil (δ ‰) notation, relative to Vienna Standard Mean Ocean Water (VSMOW) following Hayes (2004). An equilibrium fractionation factor was calculated using the temperature of the controlled environment following Horita and Wesolowski (1994) to calculate the isotopic composition of liquid in equilibrium with measured vapor water. Details of these methods are presented in Chapter 2.

Samples with high organic concentrations may confound the data and must be identified for correction. Standards used to calibrate isotopic values should have a background concentration of organics (C_2H_6 and CH_4) representing a zero plane. If the concentrations in a sample fell below the concentration values, we disregarded having an organic effect. If samples fell above this plane, we would correct isotopic values with appropriate shift determined by the Picarro Software. For isotopic analysis, samples were not affected by organic contaminants ethane or methane. Sample concentrations for ethane were not detected and methane values fell within the range of standard concentrations.

Assumptions

Our methods assumed that all soil water was both mobile and mixed within each layer of the profile, especially at t_2. Then mass balance values should not reflect any immobile soil water interstitially bound in the soil matrix. We assumed the PVC tubes we used blocked surface runoff and focused downward infiltration of the added water to saturate to 11 cm depth. Rationale for measuring soil only to a depth of 11 cm was the assumption that the upper 10 cm was representative of the first 20 cm and deeper soil only recharged during snowmelt or heavy spring rain and not likely experiencing evaporation (Hsieh 1998, Gazis and Feng 2004).

The translated assumptions into the model was then there was no downward infiltration (x_z=0). In equation 4.5, the δ_r and x_r terms are set to zero making the terms $x_r\delta_r$ zero. We were trying to solve for x_{nf} in equation 4.6 so we may remove this term with the added interpretation that δ_z is zero, $(x_{nf}+x_z)\delta_z$ becomes zero.

4.4 Results

Water Loss and Changes in Isotopic Composition

The relative proportion of the microsites measured from the vegetation survey across the study area in June 2014 was, sagebrush 45%, bunchgrass 18%, and bare ground 37%. In between sampling events, our instruments recorded ~ 150mm of precipitation, increasing the ambient soil moisture content by 7%. Following infiltration of the added water (t_1), the microsites had increased soil moisture compared to the ambient (t_0) levels of 5-10% GWC near the surface. Each microsite type reached the minimum desired GWC of 20% (up to 35 % g/g) throughout the 0-10 cm soil profile after infiltration. The added water took 3.5 times longer to infiltrate on the bare ground microsite compared to the other microsite types. Overall in September relative to June, all microsites took a significantly greater amount of time to infiltrate, ranging from 1.5-2.5 times greater ($p<0.05$) (Figure 4.2). After a period of 72 hours (t_2), water loss occurred in all the microcosms. On average, the microsite types lost 8.4 mm ± 1.1 of water in June and 8.2 mm ± 1.4 in September (Sagebrush), 8.6 mm ± 1.5 in June and 7.4 mm ± 1.6 in September (Bunchgrass), 4.7 mm ± 1.7 in June and 5.9 mm ± 1.1 in September (Bare Ground) (Figure 4.3). The amount of water evaporated as an average on the landscape measured from the eddy flux tower during a 72-hour period in June (1.72 mm ± 0.08) was much less than that in September (4.90 mm ± 0.06) (Table 4.1).

Both bare ground and bunchgrass microsite types lost almost all added water in June near the soil surface, drying to ambient surface moisture contents, whereas soil water contents near the surface remained elevated under sagebrush (Figure 4.4). In September, ambient (t_0) soil moisture content was greater but the water loss pattern (from t_2 -t_1) was similar to June. Overall, t_2 soil moisture profiles do not reach t_0 soil moisture content near the surface, so not all added water had been lost. However, bare ground profiles do reach ambient soil moisture content at 10 cm (Figure 4.4 d-f).

In June, isotopic composition of soil water at t_0 for all microsite types was similar in shape and location of isotopic maximum at 4-6 cm depths (Figure 4.5). After the added water infiltrated, the shape of the t_1 isotopic profile showed overall more depleted values. After 72 hours (t_2), sagebrush was the most enriched (-15‰), then bare ground (-17‰), and bunchgrass least enriched (most depleted) near surface (-19‰). Isotopic composition of soil water at bunchgrass microsites had the same t_0, t_1, and t_2 surface value (Figure 4.5b).

In September, ambient isotopic composition across microsite types were dissimilar in profile and location of evaporative front (Figure 4.5d-f). The evaporative front was at 2-4 cm in sagebrush and bunchgrass microsites and at 4-6 cm for bare ground. September t_2 surface isotopic values (0-2 cm) were comparable across microsite types, around -8‰.

Microclimate during observation periods

Climate data from flux towers during the 72-hour experimental time intervals showed average soil temperature from 0-30cm was 1.3 ^{o}C less in June than September, but air temperature was 10 ^{o}C greater. In June sensible heat (H) was 3 times greater, net radiation (R_n) 1.7 times greater, and soil heat flux 3 times greater, but ET was 35% less. Average soil moisture content from 0-30 cm was 7% wetter in September than June and experienced 3 times as much ET. Evaporation rates from microclimate data were less in June (0.57 mm/day) than September (1.6 mm/day). Precipitation measured during the 72-hour interval was the same as the standard error and considered negligible (Table 4.1). While precipitation during the 72-hour experiment was negligible, this site received 149 mm between June and September.

E:ET Ratios Across Microsite Types in Early and Late Summer

E:ET ratios for sagebrush, bunchgrass, and bare ground were 95%, 92%, and 100% in June, and 82%, 99%, and 100% in September respectively (Table 4.2). E:ET ratios in September showed sagebrush lost 18% of soil water in the upper 10 cm to non-fractionating processes (transpiration). Sagebrush microsite experienced 13% more transpiration, bunchgrass experienced 8% less transpiration, and bare ground remained the same with 0% transpiration in September compared to June. Near surface water loss for bunchgrass microsites decreased and E:ET ratio increased to 99% E. Bare ground lost more water in September but the ratio of E:ET remained the same. Sagebrush lost almost the same amount of water in June and September, but used more water for transpiration in the upper 10 cm.

Looking at deuterium excess, the ambient t_0 soil water had an equation of $y = 3.8x - 48.6$ in June and $y = 7.4x + 17.7$ in September (Figure 4.6). Relative to the Global Meteoric Water Line (GMWL) with the equation of $y = 8x + 10$, the reduced slope in June indicated evaporation as the primary process and the positive shift in September reflected a strong summer precipitation signal.

Table 4.1 Eddy Flux Tower data for experimental duration. Energy balance parameters; Net Radiation (Rn), Soil heat Flux (G), Sensible Heat (H), and Latent Energy (LE), are all given in a 72-hour total energy flux (MJ/m^2). Both Volumetric Water Content (VWC) and Soil Temperature (Tc) are average of 0-30 cm. Precipitation (P) values are negligible, as the sum equals the standard error of multiple sensors (Germino, Allen, Zhao, Rey et al. unpublished data)

June	R_n MJ/m^2	G MJ/m^2	H MJ/m^2	λE MJ/m^2	T_{air} °C	T_{soil} °C	VWC %	SWP kPa	P mm
Σ_{72hr}	45.59	3.94	30.62	4.22	19.55	19.34	20.26	-316.50	0.51
σ	10.24	17.21	0.25	0.20	0.03	1.18	0.49	40.84	0.51
Sept	R_n MJ/m^2	G MJ/m^2	H MJ/m^2	λE MJ/m^2	T_{air} °C	T_{soil} °C	VWC %	SWP kPa	P mm
Σ_{72hr}	27.16	1.32	10.90	12.03	9.51	20.66	27.06	-262.64	0.64
σ	**3.55**	**9.26**	**0.26**	**0.15**	**45.84**	**1.15**	**0.58**	**43.87**	**0.64**

Table 4.2 Fractions of evaporation (x_e) and non-fractionating processes (x_{nf}) for sagebrush, bunchgrass, and bare ground microsites as calculated from mass balance equations 1-2. E: ET $= \frac{x_e}{x_e + x_{nf}}$.

June	**Sagebrush**	**Bunchgrass**	**Bare Ground**
x_e	0.69	0.76	0.47
x_{nf}	0.04	0.07	0
E:ET	0.95	0.92	1
Sept	**Sagebrush**	**Bunchgrass**	**Bare Ground**
x_e	0.7	0.6	0.49
x_{nf}	0.15	0.01	0
E:ET	0.82	0.99	1

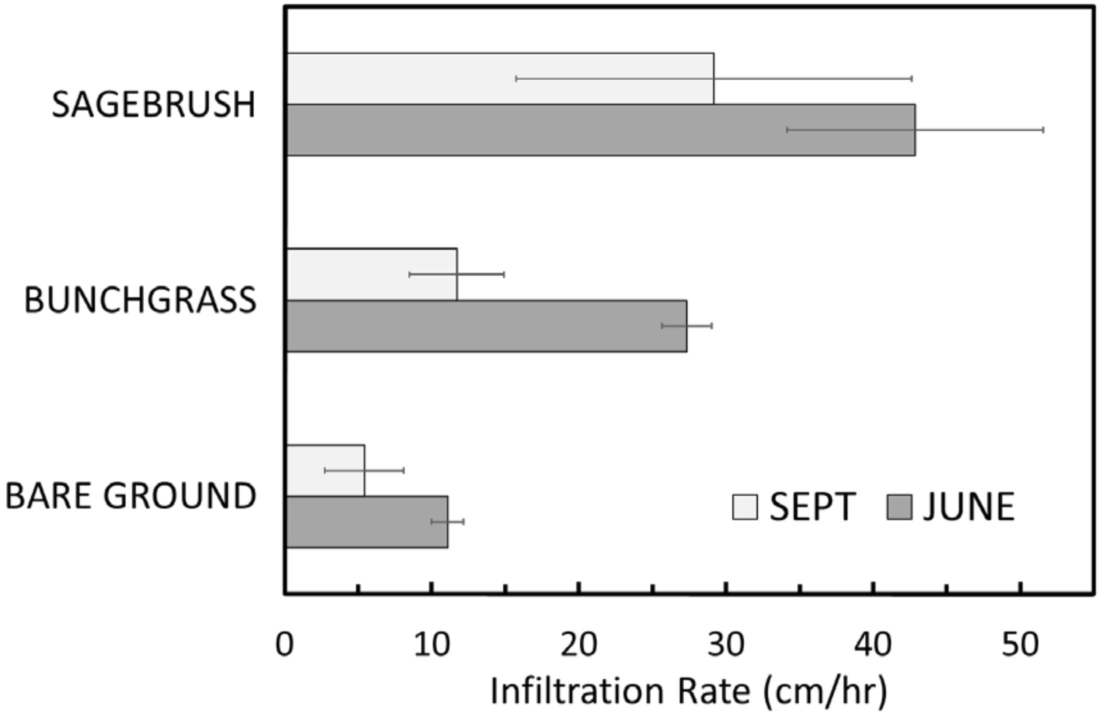

Figure 4.2 Infiltration rate by microsite type and sampling event. Sagebrush the highest infiltration rate (cm/hr) and infiltration rates decreased in September

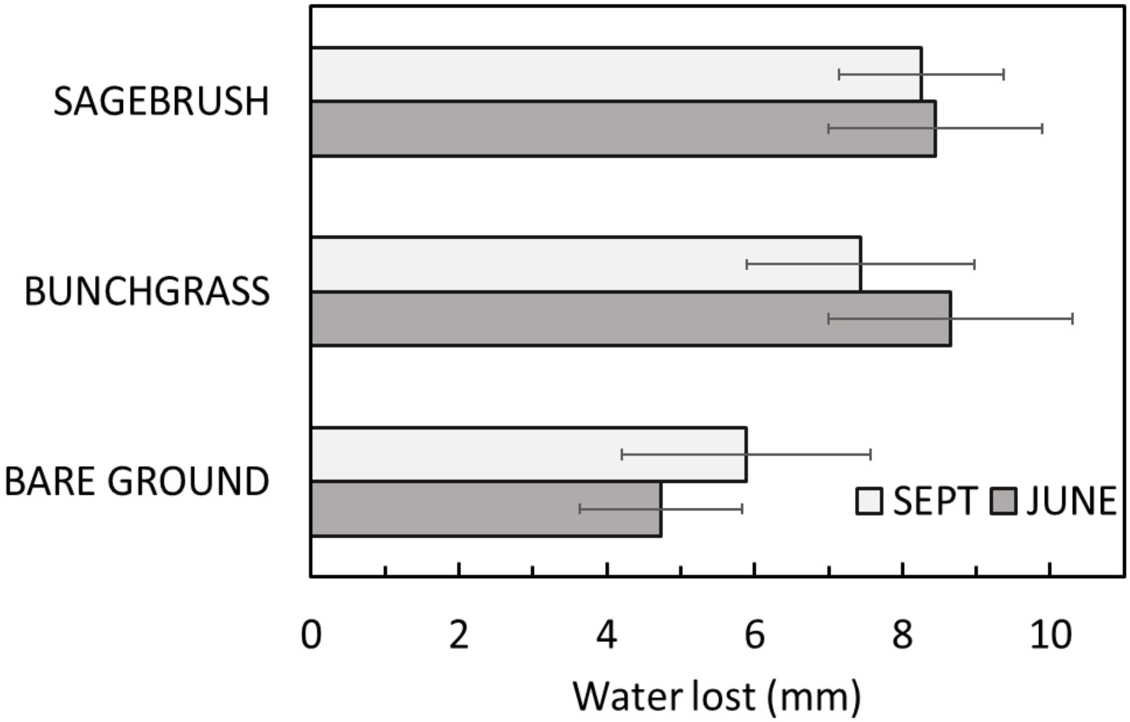

Figure 4.3 Water lost in soil profiles during the 72-hour field experiment for microsites in both June and September.

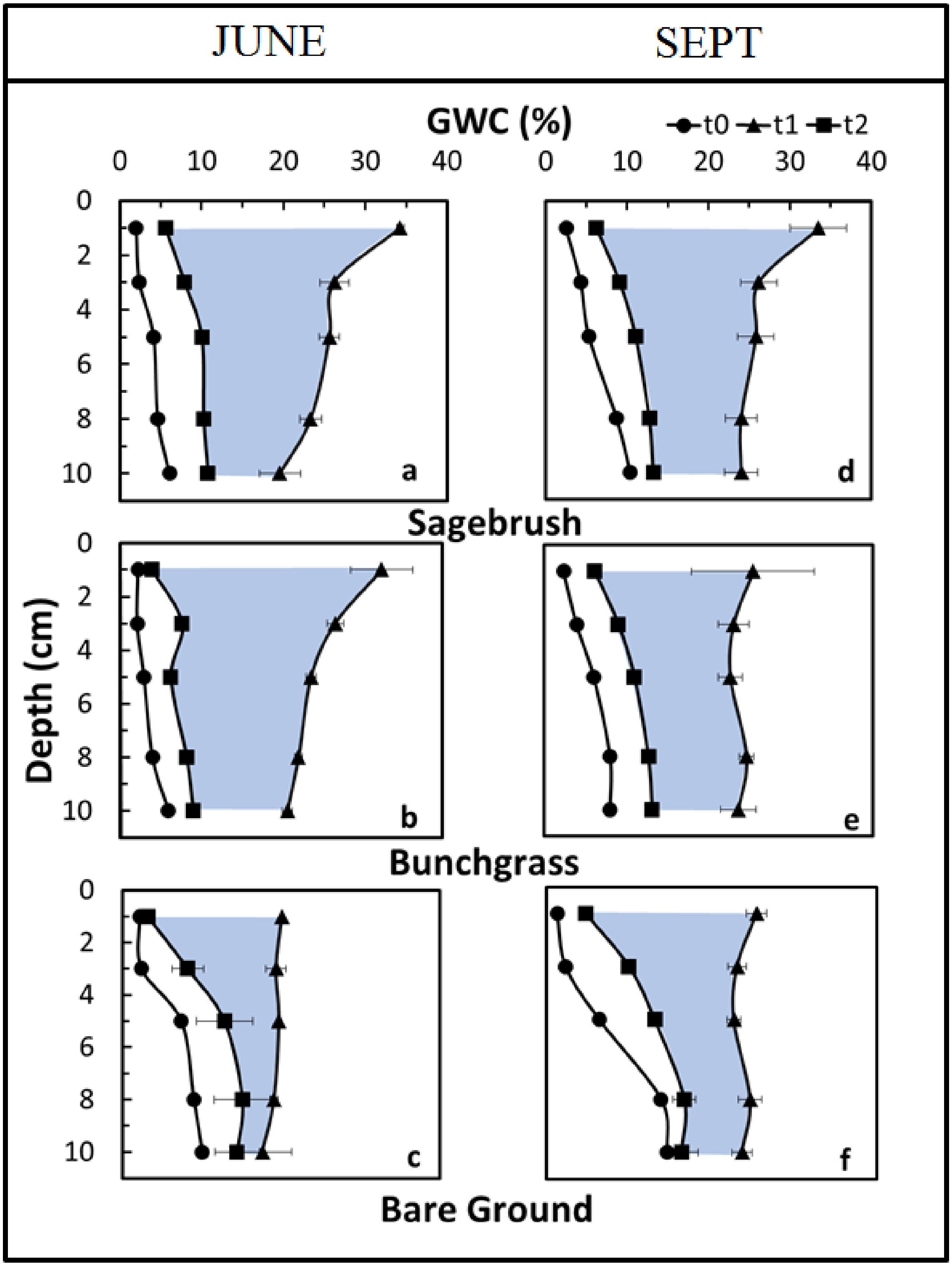

Figure 4.4 (a-f). Time series of gravimetric water content GWC (% g/g) for the three microsites: sagebrush, bunchgrass, and bare ground in a 10 cm profile. Initial condition is t_0, immediately after infiltration is t_1, 72 hours later represents t_2, with June data in left panel and September in right.

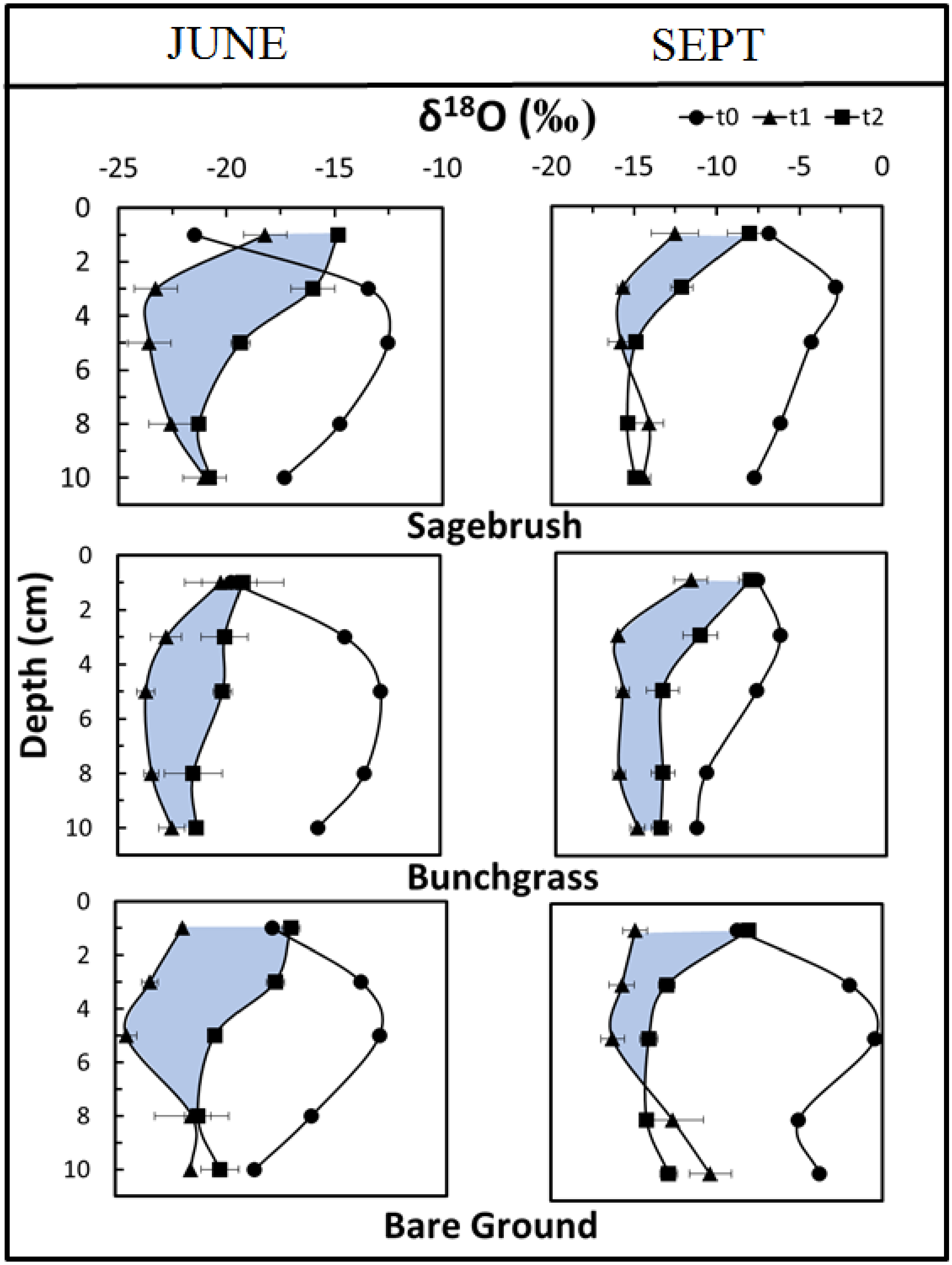

Figure 4.5 (a-f). Time series of isotopic composition for δ^{18}O for the three microsites: sagebrush, bunchgrass, and bare ground in a 10 cm profile. June data is the left panel and September is on right. Ambient condition is t_0, immediately after infiltration is t_1, 72 hours later represents t_2, with Shaded area represents shift in isotopic composition due to evaporation

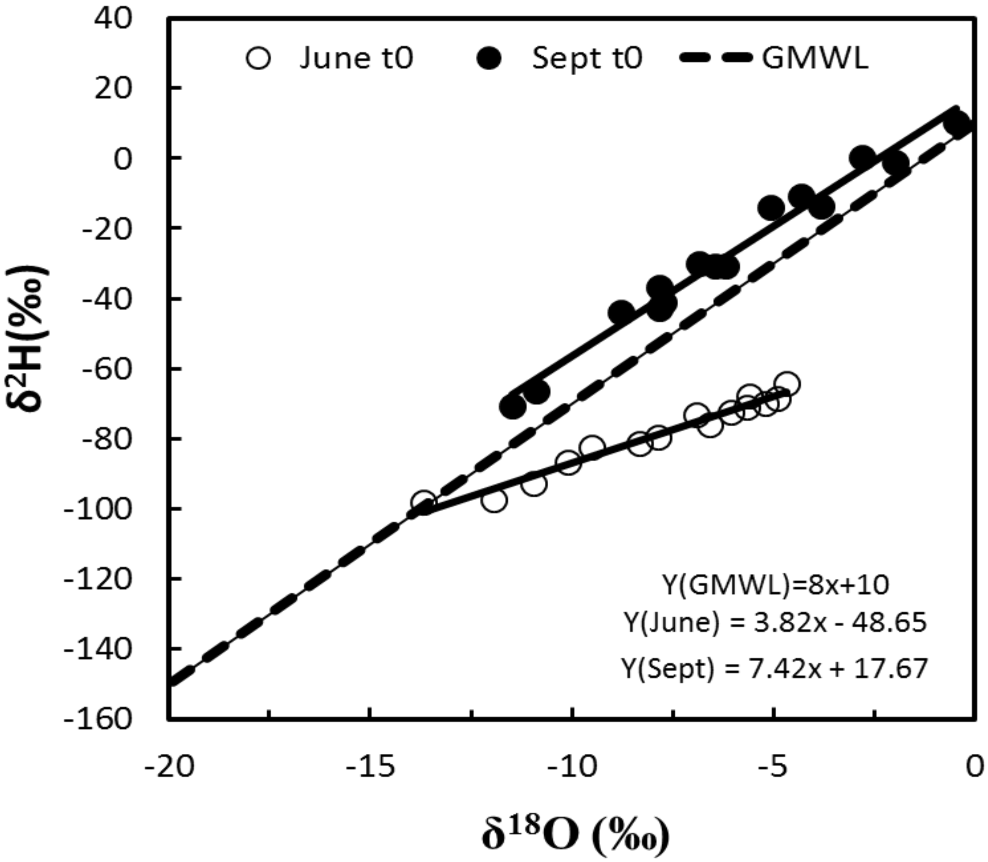

Figure 4.6 Ambient conditions (t_0) of soil water for June (open circle) and September (closed circle) sampling events relative to the global meteoric water line (GMWL). The equation for the line in June shows a clear evaporative signal with the reduced slope, whereas, the equation for the line in September shows a strong summer precipitation signal with a parallel slope, shifted towards the origin.

4.5 Discussion

In late spring (June) and summer (September) we measured changes in soil water content and isotopic composition for three microsite types in the sagebrush steppe ecosystem (sagebrush, bunchgrass, and bare ground). We used eddy flux tower data to measure the local ET flux and climate. The isotopic profiles evolved as predicted in Chapter 1 "Plants, Soils, Climate: and ecohydrologic approach”. After a wetting event occurred the soil water became uniform in GWC and δ^{18}O. Then as ET occurred, GWC decreased and δ^{18}O values became enriched. With time GWC approached the soil water content from before the wetting event (t0). Soil water isotopes became enriched but even more so near the surface (0-2 cm). The isotopic profile of the ambient soil water was more depleted at the surface (0-2 cm) and the evaporative front was deeper (4-6 cm). This is likely due to condensation of vapor at the surface with time (Yamanaka and Yonetani 1999) and we would expect that the isotopic profiles would follow this after a longer period of time.

Sagebrush lost a comparable amount of water from 0-10 cm between June and September, whereas bunchgrass lost less, and bare ground lost more soil water. The increased water loss from the bare ground microsite type reflects the increased ET flux from the landscape in September (Table 4.2). Infiltration was different across both microsite type and season. The September soil columns took an average of 1.8 times longer to infiltrate. Natural soils were wetter in Sept and had greater water loss and lower infiltration rates than June.

Contributions of E and T from were retrospectively quantified during the spring and the summer. We predicted that evergreen sagebrush would always be transpiring (E:ET <1), whereas bare soil would only have evaporation (E:ET =1), and bunchgrass would exhibit transpiration in June (E:ET <1) but not when senesced in September (E:ET =1). Most of evapotranspiration in the sagebrush steppe is from evaporation with seasonal differences in transpiration across microsite types. In June all microsite types lost soil water mostly due to evaporation with little transpiration but in September transpiration increased 13% under the sagebrush microsites. As predicted sagebrush remained evergreen with E:ET <1 in both June and September, and bare ground only experienced evaporation (E:ET=1). We predicted that by September bunchgrass would senesce (no T) and soil water loss in the upper 10 cm would be only to E. As predicted in September, soil water loss due to T under bunch grass approached 0% (E:ET=.99), yet the bunch grass was green.

The bunchgrass did not use the available near surface soil water for T yet was obviously green, so we interpret the grass used deeper soil water resources, as expected for the bunchgrass summer water use strategy (Ryel et al. 2010). The longer the drought period, the more time grass will take to resume water uptake in the shallow soil after a saturation event (Wraith et al. 1995) and the bunchgrass at our study site had not yet resumed water uptake 30 days following a major saturation event. The 13% increase in plant water uptake in the upper 10 cm of soil from sagebrush shows the opportunistic use of late summer precipitation (Ehleringer and Dawson 1992; Kurc and Small 2007) and a rapid resumption of shallow soil water (Ryel et al. 2010). We also predicted that when surface soils experienced water loss due to both E and T, gravimetric water content (GWC) would be less than if water loss was due to only E. This is the case where bunchgrass and sagebrush, both having experienced E and T, lost more water than bare ground which only lost water to E. As natural soil water content increased from June to September so did natural transpiration on the landscape

Overall, methods to retrospectively partition ET into E and T were successful and provide guidance for future ecohydrologic studies. Kinetic fractionation effects due to temperature gradients in the soil were not considered in this study since the focus was not on water fluxes, but rather proportions of water lost to E relative to T. In future studies, thermocouple data and a kinetic fractionation factor could be used to calculate evaporation rates. While results of the E:ET ratios give fraction of evaporation and we aren’t directly measuring transpiration, we assume that the remainder of initial water less the evaporation is due to transpiration. Attempts to mitigate potential sources or water loss other than transpiration or evaporation were made by guiding the infiltration of water directly into the soil column, adding the right amount of water to wet a known volume to a desired moisture level, and severing root connections in the upper soil surface. However, downward infiltration may not have been accounted for during September, the soil column may have been hydro active below 11 cm due to the large precipitation event that occurred in August. This study suggests that while this method is challenged at small portions of evaporation (Kool et al.2014), it may be challenged at small portions of transpiration. Refinement of the method is needed with an increase in scale of application. Future studies should be done to measure E:ET capturing natural wetting events followed by constant ET, over the course of all seasons. More seasonal E:ET studies should be done in the sagebrush steppe as improved knowledge of surface hydrology in this region may help narrow the error associated with water balance efforts. Truly understanding patterns and quantities of water loss may help inform broader applications such as restoration efforts and the study of ecosystem dynamics.

CONCLUSION

This master's thesis in the discipline of hydrologic sciences used an ecohydrologic approach to simplify and improve methods for partitioning evapotranspiration (ET) into evaporation (E) and transpiration (T). I used stable isotope hydrology and developed a simplified isotope mass balance model to partition ET and tested the model under both a controlled and natural environment.

I tested soils of varying moisture contents to ensure that the method for isotopic analysis was suitable for the arid soils of the sagebrush steppe. I used a liquid-vapor equilibration method to analyze isotopic composition of vapor in the headspace of a soil sample. The goal was to determine how long soils of varying water content would take to reach liquid-vapor equilibration and also isotopic equilibration. I tested soil with 10%, 5%, and 2% gravimetric water content and found that these soils took 3, 4, and 5 days respectively to equilibrate. This became a guideline for time needed for soil samples in the following experiments to reach liquid-vapor equilibration.

I tested a simplified isotope mass balance model to determine if under controlled conditions the model would predict 100% E:ET. To do this I created soil microcosms and controlled for runoff, downward infiltration, incoming precipitation, and transpiration. I was able to partition ET into E and T with a E:ET ratio of 0.99: 0.10 (4.3% error) and consider this a successful test of the model.

With the same controlled conditions except for transpiration, I applied this simplified method to a natural setting to study microsite and seasonal differences in E and

T at a sagebrush steppe field site near Hollister, Idaho. Could I measure changes in water content and isotopic values in the sagebrush steppe to calculate E:ET ratios and see differences in microsites and seasons? By studying sagebrush, bunchgrass, and bare ground microsites in late spring and late summer, I was able to calculate differences in E:ET ratios across microsites and seasons. The simplified mass balance equations seemed to work and the ratios followed the predicted scenarios. The soil water content and isotopic profiles evolved as predicted, in that when evaporation occurred water content decreased and isotopic values became enriched, more so near the surface (0-2 cm). I captured the shift in transpiration increasing across the landscape and sagebrush and bunchgrass microsites shifting the depth at which the plants accessed soil water to use for transpiration. Sagebrush shifted from using deeper water resources (>10 cm) in June to using soil water within the upper 10 cm in September. Bunchgrass used some water within the upper 10 cm in June but used even less in September even though the grasses were obviously green and transpiring, implying the use of deeper water resources.

Results from this study support the use of a simplified isotope mass balance model to interpret relative changes in E and T across a landscape. This thesis will aid future ecohydrologic studies concerning proportions of soil water loss due to evaporation and transpiration in semi-arid ecosystems. At a larger scale I hope this study will support a change in thinking of ET as a singular parameter in ecosystem water balance and instead as two separate processes that may be retrospectively assessed.

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