DISSERTATION

A STUDY OF LONG-TERM SOIL MOISTURE DYNAMICS: ASSESSING BIOLOGICALLY AVAILABLE WATER AS A FUNCTION OF SOIL DEVELOPMENT

Submitted by

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ABSTRACT

A STUDY OF LONG-TERM SOIL MOISTURE DYNAMICS: ASSESSING BIOLOGICALLY AVAILABLE WATER AS A FUNCTION OF SOIL DEVELOPMENT

Forecasting ecosystem responses to global change is highly uncertain in light of the alarming rates of climate change predicted by the scientific community. Rising CO_2 concentrations not only cause increased warming, but may also influence the amount and distribution of rainfall in terrestrial ecosystems. This in turn affects plant growth and the ability of ecosystems to perform important functions including nutrient cycling and decomposition.

Soil moisture is considered the major control of ecosystem structure and function, and it is considered the most limiting resource to biological activity in semi-arid grassland ecosystems. Total soil moisture potentials are controlled by edaphic properties such as texture, structure, micro-porosity, bulk density, soil depth, clay mineralogy, and organic matter content. Physical and chemical properties interact with hydrologic inflows and outflows to control soil moisture causing the soil to act as a store and regulator in the water flow system of the overall ecosystem. Thus, the soil acts as both temporary storage of precipitation inputs and as a regulator controlling the partition between inputs and the major outflows: evapotranspiration, runoff, leaching, and flow between organisms. Understanding the pedologic controls of water retention is critical in considering the long-term dynamics of ecosystems and projecting the consequence of global change.

The focus of my dissertation is twofold: to elucidate the change in water holding characteristics of soils through pedogenesis and to quantify how global change will impact soil moisture in the U.S. Great Plains. In order to best address my research questions, I began by

studying two established soil-chronosequences in northeast Colorado and central Wyoming to assess the characteristics of soil's physical and chemical properties. I examined how they control the biologically available water holding capacities that change predictably as a function of soil age. Next, I examined other notable soil chronosequences across the western United States to test the millennial evolution of soil water holding capacities through various climates and soil parent materials. Finally, I used a soil moisture simulation to spatially model the historical, contemporary, and future projections of soil moisture on the Great Plains.

I found in semi-arid ecosystems that three broad stages of soil development exist and are linked to landscape ages that are ecologically and biogeochemically significant: aggrading, equilibrium, and retrogressive stages. Soils in the *aggrading stage* are typically weakly developed, have genetically simple horizon differentiation, and minimal water retention. Prominent clay and carbonate features are expressed in the *equilibrium stage* soils which show more complex soil horizonation, structure, aggregation, and porosity. Within these intermediate soils, the capacity to store water reaches a maximum. Declining or *retrogressive stage* soils show losses of clays and carbonates, have undergone extensive leaching, and the soil's capacity to store water is at a minimum compared to the aggrading and equilibrium stages. Furthermore, I confirmed when modeling soil moisture in the Great Plains that course-textured landscapes store less soil water and when accompanied with disturbance are more vulnerable to climate change.

Overall, my dissertation focuses on understanding the role pedogenesis has on soil water holding characteristics and how global change impacts semiarid landscapes. My results have helped improve understanding of long-term ecosystem biophysical feedbacks through quantifying soil moisture retention characteristics across soil age and climatic processes by linking soil water properties to climatic and pedogenic variables.

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CHAPTER 1

INTRODUCTION AND DISSERTATION OBJECTIVES

1.1 Background

Soil moisture is the amount of water stored in the vadose zone, or the unsaturated portion of the soil. In regard to the global hydrological cycle, the overall quantity of soil moisture is small (approximately 0.005% of the Earth's total water volume; Jones 1997), yet it is recognized as a significant component of the climatological, hydrological, and ecological systems. Soil moisture availability is considered a major driver of plant productivity globally as soil water provides plant-available water. Globally, insufficient soil water is considered the factor most limiting to plant growth (Noy-Meir 1973), and understanding how soil moisture can be used without significantly damaging natural ecosystems is one of the keys to sustainable land-use and the prevention of ecosystem decline.

Soil composition varies greatly spatially and temporally, but it almost always includes material in the solid phase (including organic and inorganic material), water phase (including solutes), and gaseous phase (including oxygen, carbon dioxide, and nitrogen). The water soil phase does not completely fill pore spaces within the vados zone, and is held tightly to mineral grains and organic matter through capillary and adsorptive forces. Rather than the total physical amount of water, soil water in tension between the energy bounds of field capacity (~ –0.033 MPa) and the permanent wilting point (~ –1.5 MPa) characterizes the biologically transpirable pool of water. Quantities of plant available water are often more important to ecologists as it is considered the functional pool of water regulating the functioning of ecosystem (Ryel et al. 2008).

The amount of water stored in the soil is highly dependent on soil characteristics. The key edaphic properties controlling soil moisture storage includes: texture, structure, micro-porosity, density, soil depth, organic matter content, clay mineralogy, and landscape position. All these characteristics interact with hydrologic inflows and outflows to augment soil moisture with the soil both storing and regulating water flow system within the ecosystem. It acts both as a temporary store for the precipitation inputs and as a regulator controlling the major outflows: runoff, evapotranspiration, and flow between organisms. Particularly in drier climates, the soil transforms non-continuous rainfall or snow into a semi-continuous supply of water for plant growth.

When considering effects of climate change on soil moisture, responses will not only vary with the magnitude of climate (or land use) change, but also with soil moisture storage characteristics. The plant available water-holding capacity (PAWC) characteristics change across ecosystems and landscapes and thus will affect changes in soil moisture deficits. Soils with lower water-holding capacity attain soil moisture deficits earlier and pose a greater sensitivity to climate change. Alternative, soils with higher water-holding capacities have the potential to dampen the effects of climate change. It is thought that arid and semi-arid ecosystems are particularly vulnerable to global change since their soils are generally characterized by poorly developed soil profiles, low soil water contents, high potential evapotranspiration, and consequently relatively low rates of percolation and leaching.

In semi-arid and arid ecosystems soil moisture is considered the major control of ecosystem structure, function, and diversity and is recognized as the greatest limiting resource to biological activity (Noy-Meir 1973). This is because arid and semi-arid climates have low, variable, and unpredictable precipitation patterns (Schwinning and Sala 2004); although as a general rule, both water and nitrogen may be co-limiting resources (Hooper and Johnson 1999; Austin and Sala 2002) with phosphorus limitations in certain ecosystems (Walker and Syers 1976). Pulses or annual averages of precipitation are often considered good predictors of production in arid and semi-arid ecosystems at regional scales (Webb et al. 1978; Ogle and Reynolds 2004). At local site and landscape scales, however, plant community composition and production is greatly influenced by soil water availability associated with soil landscape position, infiltration capacity, water holding capacity, and species composition (Walter 1973; McAuliffe 1994; Reynolds et al. 2004; Swemmer et al. 2007). Furthermore, soil heterogeneity often complicates the deciphering of mechanisms and processes behind observed patterns in plant production. A detailed knowledge of plant physiology, as well as comprehensive knowledge of both soil pedology and hydrology are important in interpreting responses in soils with complex development.

Water dynamics of arid and semi-arid soil systems are poorly understood often due to complexity of those ecosystem's soil profile development. Modeling studies of soil water dynamics in water limited systems often rely on predictive relationships between easily measured soil properties (such as particle size distribution and organic matter) and difficult to measure soil hydrological properties (hydraulic conductivity and soil water retention characteristics) to generate the parameters necessary to simulate infiltration. While there is a growing amount of research studying soil-plant feedback mechanisms of the state factors of soil development (Jenny 1941) including topography, vegetation, climate, and parent material (Burke et al. 1999; Gutierrez-Jurado et al. 2006; Schlesinger et al, 1990; Wilcox et al. 2003; Davidson et al. 2000), consideration soil age has typically been limited to ecosystem development (Peltzer et al. 2010).

Landscape age and ecosystem response over longer time scales (thousands to millions of years) are often studied through soil chronosequences (Jenny 1941, 1994) where the differences in other factors of soil development are minimized and only soil age is allowed to vary. Studies of soil chronosequences have yielded significant insights into pedogenesis and long-term ecosystem development (Stevens and Walker 1970, Walker and Syers 1976, Thompson 1981, Vitousek 2004, Wardel et al. 2004). Absent of rejuvenating disturbances, ecosystems properties such as production, decomposition, and nutrient cycling do not remain constant, and biophysical feedbacks exert great control on soil retention characteristics. Soil-age--plant interactions in water limited ecosystems present a significant opportunity to understand both the short term (runoff response, seasonal growth) and long-term processes (soil evolution, plant succession) operating within watersheds.

1.2 Research objectives

The overarching goal of this research is to develop a mechanistic understanding of the range and variability of soil moisture conditions and to quantify global change impacts (e.g. climate change) on soil moisture in the western Great Plains. In order to address this objective, I developed a pedogenically based model of water holding characteristics across millennial time scales established from various soil chronosequences in the Western United States. I then used soil moisture simulations to spatially model the historical, contemporary, and future projections of soil moisture on the Great Plains. The greatest degree of vulnerability of an ecosystem to climate change is reflected in the soil's current hydrologic regime (e.g. soil pedological properties, landscape properties, and climatic variables). In this study I focus most of my analyses on soil moisture regimes and argue that soil moisture can be utilized as a master variable or the key determinant of ecosystem structure and function. Specific objectives of this research were to:

1) Develop a pedohydrologic model of soil profile development as a function of time and define unique pedogenic phases, grouping soil into distinct hydrologic units (e.g., soil hydrologic regimes) to identify soil properties and landscape elements that are either resilient or vulnerable to global change drivers.

2) Identify trends of moisture characteristics across established soil chronosequences in the arid and semi-arid western United States across different climates.

3) Create a chronicle of soil moisture and soil climate for the North American Great Plains as a function of monthly temperature, monthly precipitation, soil properties, and annual production for the contemporary historic record (1895-2012).

4) Forecast soil climate regimes across landscape gradients and quantify vulnerability to projected climate change scenarios in the Central Great Plains.

1.3 Dissertation Format

My dissertation contains six chapters: an overall introduction including research objectives (this chapter), four manuscript chapters (Chapters 2, 3, and 4), and a general conclusions section (Chapter 6). Chapters 2, 3, and 4 were written as separate manuscripts to be submitted for publication in the peer-reviewed scientific journals listed below. Therefore, some redundancy exists among these three manuscript chapters. Chapter 2, "Pedogenic controls on plant available water across semi-arid chronosequences," will be submitted to the journal Geoderma. Chapter 3, "Vulnerability of Landscapes to Climate Change: Assessing a 118-year chronology of soil moisture dynamics in Grassland Ecosystems of the Great Plains, North America, USA," is in preparation for publication with a journal yet to be determined. Chapter 4, "A long-term perspective of the dry boundary for the Great Plains of North America," will be submitted as a note to the SSSA journal Soil Horizons. These chapters will be submitted for publication starting in 2015. Methods are discussed in each manuscript, so there is no separate methods section in the dissertation. Each manuscript also has its own introduction and conclusions section, which contain a review of the literature relevant to the topic of that particular manuscript. Thus, the introduction and literature review (Chapter 1), as well as the general conclusions (Chapter 5), are both relatively brief and attempt to give an overview of the entire dissertation and summarize and integrate the results from the four manuscript chapters.

1: REFERENCES

Austin, A.T., and O. E. Sala, 2002, Carbon and nitorgegn dynamics across a natural precipitation gradient in Patagonia, Argentina, Jouranl of Vegetation Science, 13:351-360.

Burke IC, WK Lauenroth, R. Riggle, P Brannen, B Madigan, S Beard, 1999, Grazing, Microsite, and Plant Species in Controlling Spatial Patterns, Ecosystems, 2:422-438.

Davidson EA, SE Trumbore, R Amundson, 2000, Biogeochemistry: soil warming and organic carbon content, Nature 408: 789.

Dingman SL, 2002, Physical hydrology. 2nd ed. Prentice Hall, Upper Saddle River, NJ. Physical hydrology. 2nd ed. Prentice Hall, Upper Saddle River, NJ. 2002.

Gutierrez-Jurado HA, Vivoni ER, Harrison JBJ, Guan H (2006) Ecohydrology of root zone water fluxes and soil development in complex semiarid rangelands. Hydrol Process 20:3289–3316

Hooper, DU, L Johnson, 1999, Nitrogen limitation in dryaldn ecosystems: responses to geographical and temporal variation in precipitation, Biogeochemistry 46: 247-293.

Jenny, H, 1941, Factors of Soil Formation: A System of Quantitative Pedology, McGraw-Hill, New York.

Jenny, H. 1994. Factors of soil formation. A system of quantitative pedology. Dover Press, New York, New York, USA.

Jones, JAA, 1997, Global Hydrology: Processes, resources and environmental management, Addison Wesley Longman, Harlow, UK.

McAuliffe, JR, 1994, Landscape evolution, soil formation, and ecological patterns and processes in Sonoran Desert bajadas. Ecol Monogr 64:111–148.

Noy-Meir, I. 1973. Desert ecosystems: environment and producers. Annual Review of Ecology and Systematics 4: 23-51.

Ogle and Reynold, 2004, Plant responses to precipitation in desert ecosystems: integrating functional types, pulses, thresholds, and delays. Oecologia 141:282–294.

Peltzer DA, DA Wardle, VJ Allison, WT Baisden, RD Bardgett, OA Chadwick, LM Condron, RL Parfitt, S Porder, SJ Richardson, BL Turner, PM Vitousek, J Walkker, LR Walker, 2010, Understanding ecosystem retrogression, Ecological Monographs 80(4) 509-529.

Reynolds, JF, Kemp PR, Fernández RJ, Ogle K (2004) Plant functional-type responses in the warm deserts of North America: “drinking from the same cup” in pulsed environments. Oecologia.

Ryel RJ, Ivans CY, Peek MS, Leffler AJ (2008) Functional differences in soil water pools: a new perspective on plant water use in water-limited ecosystems. Progress in Botany 69:397-422.

Schlesinger, W. H., J. F. Reynolds, G. L. Cunningham, L. F. Huenneke, W. M. Jarrell, R. A. Virginia, and W. G. Whitford. 1990. Biological feedbacks in global desertification. Science 247(4946):1043–1048.

Schwinning, S. and Sala, OE, 2004, Hierarchy of responses to resource pulses in arid and semi-arid ecosystems. Oecologia, 141: 211-220.

Stevens, P. R., and T. W. Walker. 1970. The chronosequence concept and soil formation. Quarterly Review of Biology 45:333–350.

Swemmer AM, Knapp AK, Snyman HA, 2007, Intra-seasonal precipitation patterns and above-ground productivity in three perennial grasslands. Jouranal Ecology 95:780–788.

Thompson, CH, 1981, Podzol chronosequences on coastal sand dunes of eastern Australia, Nature 291:59–61.

Walker, T W, J K. Syers, 1976, The fate of phosphorus during pedogenesis, Geoderma 15:1–19.

Walter H, 1973, Vegetation of the earth in relation to climate and the eco-physological conditions, Springer, New York.

Wardle, DA, LR Walker, RD Bardgett, 2004, Ecosystem properties and forest decline in contrasting long-term chronosequences, Science 305:509–513.

Webb WS, Szarek S, Lauenroth W, Kineson R, Smith M, 1978, Primary productivity and water use in native forest, grassland, and desert ecosystems, Ecology 59:1239–1247.

Wilcox, B. P., Breshears, D. D., & Allen, C. D. (2003). Ecohydrology of a resource-conserving semiarid woodland: Effects of scale and disturbance. Ecological Monographs, 73, 223–239.

Vitousek, P. M. 2004. Nutrient cycling and limitation. Hawaii as a model system. Princeton University Press. Princeton, New Jersey, USA.

CHAPTER 2

PEDOGENIC CONTROLS ON BIOLOGICALLY AVAILABLE WATER IN SEMI-ARID ECOSYSTEMS[1]

2.1 Summary

Dry land ecosystems are defined by the interaction of water with soils, plants, and landscapes in regions that are particularly sensitive to variations in water availability. While these interactions have been studied across a wide range of topographic and biologic gradients, less is known about the biophysical controls exerted by soil age on ecosystem processes through soil-age-plant hydrologic functioning. Furthermore, profile heterogeneity can lead to important differences in the soil hydraulic properties, water balance, and ecosystem production. This study outlines the mechanisms affecting water retention characteristics, soil development, and hydrologic fluxes in semi-arid ecosystems across two soil age gradients. We illustrate pedologic controls on alluvial terraces in the Colorado Piedmont of north-east Colorado and in the Wind River Basin, Wyoming. Soil age's consequence on root zone hydrologic fluxes and soil development in the two landscapes were assessed using soil observation and hydraulic properties from pedotransfer functions (PTFs). Our results show that in semi-arid ecosystems, broad stages of soil development exist and are linked to landscape ages that are ecologically and biogeochemically significant: an aggrading, an intermediate phase where plant available water capacity reaches a maximum, and a regressive phase where long-term weathering has altered soil properties (e.g., structure, texture, and chemical composition) and where plant available water

[1] This manuscript is being prepared for publication to the journal Geoderma with co-authors RM Bergstrom, PH Martin, AK Knapp, OA Chadwick, and EF Kelly.

declines. Soil texture was a strong determinant in available water holding capacity with the Wind River's sandy soils holding less available water than the loamy soil gradient in the Colorado Piedmont. We also found that in intermediate soil ages where clay and carbonates continue to increase, biologically available water begins to decrease. At the oldest soil stages, water retention, as well as other soil properties such as organic matter, nutrient availability, clays, and carbonates, steadily decrease. Our results indicate that the quantity of biologically available water is directly tied to soil age and this study helps improve the understanding of long-term ecosystem biophysical feedback through soil moisture retention characteristics and climatic processes.

2.2 Introduction

Long-term ecosystem development is controlled by biotic and abiotic factors including climate, parent material, and substrate age (Vitousek 2004), and although these factors exert great influence over the nutrients sources in the ecosystem (Chadwick et al. 1999), the consequences are that the biologically available water capacity in the soil is largely unknown across time scales of thousands to millions of years. Since soil moisture is considered the major control of ecosystem structure, function, and diversity, and is recognized as the greatest limiting resource to biological activity in semi-arid ecosystems (Noy-Meir 1973), understanding soil-age-plant interactions is necessary to understand both the short-term (runoff response, seasonal growth, and water balance) and long-term processes (soil evolution, plant succession, and ecosystem development) operating within landscapes. Mechanisms of plant-available water capacity and water storage as they relate to the state factors of topography, vegetation, climate, and parent material tend to be well understood (Burke et al. 1999; Gutierrez-Jurado et al. 2006; Schlesinger et al. 1990; Wilcox et al. 2003; Davidson et al. 2000), yet studies have not investigated the role soil age has on long-term dynamics of biologically available water.

Absent of rejuvenating disturbances, ecosystem properties such as production, decomposition, and nutrient cycling do not remain constant over millennial time scales. Typically pool sizes and flux rates of organic carbon and nitrogen progressively increase from the early to middle stages of pedogenesis. This is accompanied with development of soil horizons, structure, and a maximal biomass phase. The progressive phase is often followed by a decline or regressive phase characterized by significant reductions in storage and cycling rates of ecosystem carbon, as well as the weathering of clays and carbonates on the most highly weathered substrates (Stevens and Walker 1970; Syers et al. 1970; Robertson and Vitousek 1981; Riley and Vitousek 1995). Overall, the transition from progressive-to-decline phase during pedogeniesis is better understood in humid ecosystems (Wardle et al. 2004), though recent research is emerging in arid and semi-arid ecosystems (Austin et al. 2004; Selmants and Hart 2008).

Progressive pedogenesis generally leads to higher soil water holding capacities. For example, the weathering of course fragments into clay sized particles increases plant available water storage (Hook and Burke 2000), and the associated increases in organic matter accumulation via soil aggregation and adsorption on mineral surfaces enhances soil structure and water holding capacity, and increases infiltration (Kaye and Burke 2002). Soils in dryland systems are generally characterized by coarse textures, low soil water contents, high potential evapotranspiration, and consequently relatively low rates of percolation and leaching (D'Odorico and Porporato 2006). Furthermore, arid climates have low, variable, and unpredictable precipitation patterns (Schwinning and Sala 2004). Because timing and quantity of water fluxes, including precipitation and evapotranspiration, strongly determines the rate and trajectory of soil pedogenesis (Chadwick and Chorover 2001), ecosystem development of semi-arid and arid ecosystems then are fundamentally different, progressing on much slower time scales than humid ecosystems.

This study takes advantage soil age gradients isolating time as a soil forming factor (Jenny 1941, 1980) by selecting genetically related groups of soils evolving with similar vegetation, topography, parent material, and climate. This soil-chronosequence translates the spatial differences between soils into temporal differences by substituting space for time (Pickett 1989). Studies of soil-chronosequences have yielded significant insights into pedogenesis and long-term ecosystem development (Stevens and Walker 1970; Walker and Syers 1976; Thompson 1981; Vitousek 2004; Wardel et al. 2004). While there is criticism of indiscriminant use of chronosequences (Picket 1989; Philips 1993; Johnson and Miyanishi 2008), consistent, robust results have emerged from a range of chronosequence studies worldwide, at least at the ecosystem level (Peltzer et al. 2010).

In this chapter, we will use soil age to determine the range of available water holding characteristics across two established long-term soil-chronosequences in semi-arid systems. Specifically, we will look at key edaphic properties controlling soil moisture contents such as soil texture, carbonate and clay accumulation, organic matter content, and profile development–all of which can strongly interact with precipitation inputs to reduce or augment soil moisture. Our study was designed to address three fundamental research questions: (1) How does biologically available water change with respect to pedogenesis and soil development within semi-arid ecosystems? (2) What are the key properties in semi-arid ecosystems controlling water retention characteristics? (3) Do trends of production emerge across soil-age gradients linked to the change in available water capacity?

2.3 Materials and Methods

Two soil-age gradients (Figure 2-1) were established and characterized as post-incisive on strath river terraces in semi-arid steppe climates of the northern Colorado Piedmont of Colorado, USA (40.728 N, 103.579 W) and along the Wind River near Bull Lake, Wyoming, USA (43.210 N, -109.047W). These sites were selected as they share similar steppe climates while varying in texture: the Colorado gradient having more loamy soils and the Wyoming gradient exhibiting sandier soils (see table 2-1 for soil descriptions for the two chronosequences). Soil morphological data from both sites are from unpublished data describing the Colorado Piedmont (Loadholt 2002) and the Wind River Basin in (Peacock 1994).

The first soil chronosequence is from six modal soil pedons forming a 2,000 to 600,000 year gradient in northern Colorado along Horsetail Creek, a tributary of the South Platte (unpublished data from Loadholt 2002). Replication was obtained by two additional soil cores located approximately two meters from the central pedon which were collected to examine spatial variability of soil properties. The Horsetail Creek chronosequence (HCC) soils are within a 16 km area on alluvial terraces ranging from between 7 and 70 meters above the modern stream elevation. Mean annual temperature and precipitation records covering 1981-2010 from a weather station nearby are 8.6°C and 321 mm (Lauenroth et al. 2008). The terraces consist of granitic and metamorphic sand and gravel veneered surfaces derived mostly from glacial melt water from the Rocky Mountains, and geologically have been identified as associated with interglacial periods (Scott 1982). Deposits range in thicknesses between <1 m to > 30 m. Vegetation on the sites is dominated by *Bouteloua gracilis* and *Buchloe dactyloides* typical of the short grass steppe ecosystem. Sites were selected from geomorphically stable topographic positions, being level to gently sloping surfaces (Loadholt 2002).

The second soil-chronosequence was selected from seven soil geomorphic surfaces forming an 18,000 to 1,320,000 year old chronosequence along terraces above the Wind River in central Wyoming (unpublished data from Peacock 1994; Chadwick et al. 1994). Replication of soils was based on local topographic vegetation variations from within each terrace. Wind River chronosequence (WRC) sites range between 6 and 223 meters above the modern stream elevation. The soil climate is arid to semi-arid and the estimated mean annual temperature and precipitation are -5.3°C and 254 mm (PRISM 2013). Terraces are thinly mantled alluvium from mixed volcanic, crystalline, sedimentary, and quartzite gravels (Hancock et al. 1999). Vegetation is a mix of *Artemisia* sp., *Agropyron* spp., cactus, and *Stipa comata*.

Biologically available water was measured based upon the difference between volumetric water content held at field capacity (θ_{FC}) and water held at the wilting point (θ_{WP}). Samples of θ_{FC} were measured or modeled based on water retained at -0.033 MPa, and samples of θ_{WP} were determined at the traditional wilting point of -1.5 MPa (Romano and Santini 2002). Although 4.0 MPa could be an appropriate measurement of the wilting point for desert adapted plant species (Senock et al. 1994; Pockman and Sperry 2000), with some estimates even as low as -10 MPa (Odening et al. 1974), we used a hydraulic wilting point of -1.5 MPa for two reasons. Firstly, it is the traditional value and a common output from hydraulic models and pedotransfer functions (PTFs), and secondly, since our temporal resolution is across geologic time scales we consider this range of soil water as the active water pool controlling pedogenesis and biological activity.

Soil hydraulic properties for both HCC and WRC were estimated using the PTF model "k-Nearest Neighbor" Version 1.00.02 (Nemes et al. 2008) at water contents at the -0.033 mPa and at -1.5 MPa matric potentials. This non-parametric technique used US-NRCS-SCS Soil Characterization Database (Soil Survey Staff 1997) as reference soils data for pattern recognition based on sand, silt, and clay, as well as soil bulk density (BD) and soil organic matter (OM) contents (Nemes et al. 2006). Application input data included texture, BD, and OM variables, then assessed using a bootstrap technique to perform multiple subset sections on the reference dataset.

Measured hydraulic values were also available for the HCC soils (Loadholt 2002), which were conducted with natural clods using pressure-plate desorption procedures (Soil Survey Staff 1996). Water retention differences are reported for both sites as the volume fraction ($cm^3\ cm^{-3}$) of water retained in the soil between -0.033 MPa and -1.5 MPa suction.

Biologically available water (BAW) holding capacity values were calculated for each soil pedon using weighted mean percent profile data indices (Birkland 1999) by summing each horizon across the pedon to maximum rooting depth to find total available water holding capacity. Available water capacity is the capacity of the soil to hold water for use by most plants, calculated as:

$$BAW = \sum_{i=1}^{n} WRD_{\,i} \times z_i \tag{1}$$

where BAW is the total available water capacity for the pedon up to root-restricting layer, *I* is the soil horizon (1, 2, 3…*n*), WRD is the calculated water retention difference in the *i*th horizon, and z_i is the horizon thickness of the *i*th horizon (cm, although it is commonly expressed in mm). If multiple pedons of the same age existed within a soil age gradient, an average of those values was used to represent BAW. A t=0 from each soil-chronosequence, representing initiation of soil formation were based upon modeling the C horizon of the youngest soil in the age gradient. Hydraulic properties of this horizon were then multiplied by a depth of 100 cm to get t=0 of the sequence. Error for each soil's BAW was calculated by square rooting the sum of squares (RSS) for standard deviations reported from the "k- Nearest Neighbor" PTF model. This was done for each horizon, multiplied by the horizon's thickness, and RSS for the soil profile.

2.4 Results

Variations in the degree of soil development across the chronosequences result in directional changes in soil biologically available water with time. Physical and chemical properties for soil profiles from both soil chronosequences are summarized in tables 2-2 and 2-3. Soils of WRC and HCC exhibit expected age related morphologic characteristics showing increasing soil profile development, accumulation clay, and pedogenic carbonates. Soils from HCC are mostly loamy to sandy loam with an increasing > 2 mm fraction on older terraces. Soils from HCC terraces through the first five ages of the chronosequence show a thickening of the A horizon, and an increase in clay content, carbonates, and bulk density (Table 2-2, Figure 2-2). Calcium carbonates were found in all but the oldest terrace, with the greatest enrichment of pedogenic carbonates occurring on the HCC5 (see table 2-1 for numbering designation) showing stage III carbonate formation. This is reflected in the soil pH which is mostly alkaline (7.0-8.8) except on the oldest HCC terrace (HCC6) where surface horizons were mildly acidic with a pH of 5.8. Overall, HCC6 soil profile saw decreases in carbonates and clay compared to younger soils.

Soil parent material on the Wind River terraces has high sand content (Table 2-3). Wind River soils had many textural discontinuities due to alluvial bedding showing increases in the > 2mm fraction around 30-40 cm depth. Although WRC soils developed in an area with low precipitation, they have seen some pedogenic clay accumulation (Figure 2-2). Calcium carbonates are the most distinguishing soil profiles characteristics, with rinds forming on clasts and calcic horizons forming in older soils. Soils on terraces WRC1 (see table 2-1 for numbering designation) have horizons with stage II carbonate accumulation and all subsequent terraces show stage III carbonate accumulation with WRC9 terrace showing the greatest percentage of carbonates with a horizon with 44% $CaCO_3$. Horizon thickness and depths are quite variable. Soils are mostly alkaline with pH ranges between 7.1 and 8.6. Soil organic carbon was also variable and did not show a trend with age.

Modeled hydraulic soil properties from soil in the HCC (figure 2-5) show relationships to measured water retention values (R^2 = 0.884). Since most of the NRCS soil database used to estimate field capacity and permanent wilting point are not in arid and semi-arid values, hydraulic values could be improved by modifying the soils database to exclude soil in dissimilar climates.

Wind River site soils show a strong decreasing water storage trend with depth. The Horsetail Creek soils show a similar pattern where surface horizons have higher BAWHC than the deeper horizons; however, these soils have the highest water holding capacity in carbonate rich horizons at the middle profile depths. Horizons with higher clay and organic matter contents also show greater BAWHC relative to parent material, or C horizon. Higher clay contents are concentrated in argillic horizons which occur in all the WRC soils after the WRC4 (139ky) soil and after the HCC3 (17ky), although in the oldest HCC site shows less clay in the B horizon (11.8%). All soil profiles in HCC show decreasing organic matter with depths, while some lower WRC horizons have higher organic content then the surface horizon. The WRC soils all have lower clay and higher sand percentages then HCC soils, and thus BAWHC is greater in HCC soils with an average of 15.5% (cm^3 cm^{-3}) plant available water content versus 11.8% (cm^3 cm^{-3}) in WRC soils.

On both gradients, we noticed an increase in available water capacity with age (figure 2-2), followed by a plateau and subsequent decline in AWHC. Soils from HTC showed a greater range among sites with a maximum of 295 mm of potential water storage on the 17 ky surface to a low of 145 mm of potential water storage on the oldest surface of 600 ky. Soils on WRC terraces show a peak of BAWHC of 177 mm at 870 ky (WRC8) and decreases below the youngest terrace to an BAWHC of 145 mm. Wind River soils exhibit a lower BAWHC than HTC soils due to lower clay and organic content. Modeled *time-zero points* of the youngest parent material showed that HCC (217 mm) held about 100 mm more plant available water then WRC (114mm). Weathering then subsequently lowered AWHC on the HTC soils below the zero point (figure 2-3); where WRC's oldest soil was shown to hold more water for plants then the modeled zero point.

Remotely sensed values for plant production are shown in figure 2-4. Soil heterogeneity yielded high variances so trends can only be inferred without significance. An increase in production was observed on cropland and grassland landforms between the early phase (modern flood plain and Piney Creek Alluvium, recent to 2ky) and middle phase (Broadway and Louviers Alluvium, 10ky to 130 ky). A decrease is then identified into the final developmental phase (Slocum and Verdos Alluvium, 240ky to 600ky). Cropland identified from NLCD in phase 3 did not show declines at the same level as grassland systems, where its decline appears similar to phase 1 cropland. Grassland phase 3 production showed declines where it dropped below production values in phase one.

2.5. Discussion

Studies of soil development in the context of long-term ecosystem dynamics over thousands to millions of years have identified consistent and directional changes in soil properties, namely increases in organic matter, nitrogen, soil complexity, and biomass. These increases are followed by decreases in these key properties in older, highly weathered soils (Peltzer et al. 2010) although the transition from increasing to decreasing varies between these two sequences. Our results show that the quantity of biologically available water in soil is directly tied to soil age and is linked to the physical and climatic process controlling plant-soil feedbacks and that those feedbacks regulate long-term ecosystem development.

As noted in humid environments, absence of any rejuvenating processes, a general trend with pedogenesis appears with progressive increases in organic matter, nitrogen, and biomass followed by a decline in these properties in older and more highly weathered soils (Vitousek 2004). Based on these long-term biogeochemical cycles, Vitousek and others (1997) proposed that these developmental stages could be portioned into broad functional pedologic phases (building, sustaining, and degrading phases) of ecosystem development. Our results of biologically available water support application of this model in semi-arid environments.

Based upon our results, in water limited systems we note three broad stages of soil development that are linked to landscape age that are believed to be ecologically and biogeochemically significant: progressive, intermediate, and regressive stages. First, the *progressive* or building stage begins with a new substrate and soils are typically weakly developed. Although mineral transformations have occurred, the soil is genetically simple (little horizon differentiation). Surface and subsurface materials are pedogenically and hydrologically similar and the soil's capacity to retain water is at minimum relative to later stages of pedogenesis. In early stages of soil genesis, soils also accumulate organic matter, form pore spaces, and increase soil aggregation. This leads to dilation of soil horizons resulting from organic matter accumulation, root growth, decay, and animal activity (Brimhall et al. 1988). This is similar to young soil's humid chronosequences depending on the local disturbance leading to a new parent material. In young soils when soil genesis is dominated by progressive processes (for example: horizonation, developmental upbuilding, and soil deepening), soil regressive processes (specifically haploidization) are ongoing minimally, yet are being overwhelmed by progressive processes.

In the *intermediate* stage in semi-arid ecosystems, clay and carbonates become dominant features in the soil profiles and there is significant pedological and hydrological differentiation between surface and subsurface horizons, thus the soils capacity to store water is at a maximum relative to early or later stages of pedogenesis. Intermediate soils are in a sort of quasi-equilibrium where progressive soil processes are balanced with regressive processes. Argillic horizons are present in both chronosequences in the intermediate stage as well as build up. Peak of the intermediate stage where biologically available water is at a maximum is around 17 ky in the HCC and around 870 ky in the WRC. This delay in the WRC likely reflects slower weathering rates and reduced biological activity in the more arid ecosystem of the Wind River.

The *regressive* or declining stage, is perhaps the least understood and has seen a flurry of studies recently (Wardle et al. 2004; Vitousek 2004; Peltzer 2010; Slements and Hart 2008), and is best understood in humid ecosystems. In this stage, most weatherable primary minerals have been transformed into secondary forms, soils experience loss of clay (relative to soils in the intermediate stages of development), and $CaCO_3$ starts to decline due to weathering, loss of soil buffering capacity, and higher soil acidity. Pedological and hydrological differentiation starts to decline and the system's capacity to store water is intermediate relative to other phases described above. Our results show that the timing of ecosystem decline is directly linked to climate with HCC beginning between 17ky and 130ky while the WRC declines between 1090 and 1110ky. Selmants and Hart (2008) found similar results in a volcanic soil age gradient in Northern Arizona (MAP = 340mm, MAT = 11°C) where ecosystem retrogression began between 750ky and 3000ky, suggesting that climate strongly governs the timing of ecosystem development and decline.

Soil texture also has a strong influence over water retention characteristics. WRC had significantly lower silt and clay percentages than HCC. When comparing parent materials between the two sites, WRC has between 90-95% sand with HCC boasting half those percentages. However, sandy soils have been found on the Great Plains to be higher in productive ecosystems when precipitation is <370 mm yr^{-1} then loamy soils (Sala et al. 1988). In semi-arid systems, the major losses of soil water occur via bare soil evaporation. However, where sandy soils occur, bare soil evaporation is lower than in loamy soils because water penetrates deeper into the soil. Runoff also is lower in sandy soils than in loamy soils. In more humid regions, substantial water losses occur via deep percolation, which is reduced in soils with high water-holding capacity (Noy-Meir 1973). Therefore, in dry regions, sandy soils with low water-holding capacity have more water available for plant growth than soils with higher water-holding capacity.

Soil profile characteristics play a role in water storage and deep water percolation, and are limited by lower hydraulic conductivity of argillic and calcic horizons; however these horizons—which are much more common in semiarid soils—have been found to hold large quantities of water. Petrocalcic horizons in New Mexico were shown to hold as much as 4 times the course textured parent material (Duniway et al., 2007). This is important because the location of plant available water within the soil profile affects shallow-rooted plant species differently than deep rooting systems. Rates of $CaCo_3$ additions to soils in the Wind River region were shown to be significantly lower (Hancock et al. 1999) than reported for other arid regions (figure 2-2).

Decline in C on older substrates has been attributed to either a decline in productivity and/or an increase in C turnover rates. Although in some chronosequences there is a trend of increasing clay and carbonates content while we have shown in WRC and HCC that water holding capacity is declining. Possible explanations of this could be changes in soil density, porosity or by changes to the clay mineralogy. It is also unclear if soil development leading to retrogression in semi-arid systems is defined by the loss of parent substrate-derived elements from the weathering zone and formation of barriers to nutrient access, or if these processes are simply slower in seasonally dry or arid areas (Peltzer et al. 2010). Selmants and Hart (2008, 2010) have suggested that decline is only possible over long time frames which compound the effect of major historical climate shifts, leading to the need to consider long-term precipitation variables (Hotchkiss et al. 2000).

2.6 Conclusions

Our results have several important implications for the role of soils in terrestrial ecosystem development. It is only in the last few years that ecosystem decline due to pedogenesis has been associated with semi-arid and arid ecosystems (Selmants and Hart 2008; Peltzer et al. 2010). Our analysis of BAW across soil-age gradients shows that soil retrogression occurs in semi-arid landscapes in the in semi-arid soils of the high plains as well. Interestingly, our results also show that biologically available water capacity starts to decline in these soils even while clay and organic carbon content continue to increase with soil substrate age, possibly due to alterations in bulk density and soil structure. Our results indicate that the quantity of biologically available water is directly tied to soil age, and this study helps improve the understanding of long-term ecosystem biophysical feedback through soil moisture retention characteristics and climatic processes.

2.6: TABLES AND FIGURES

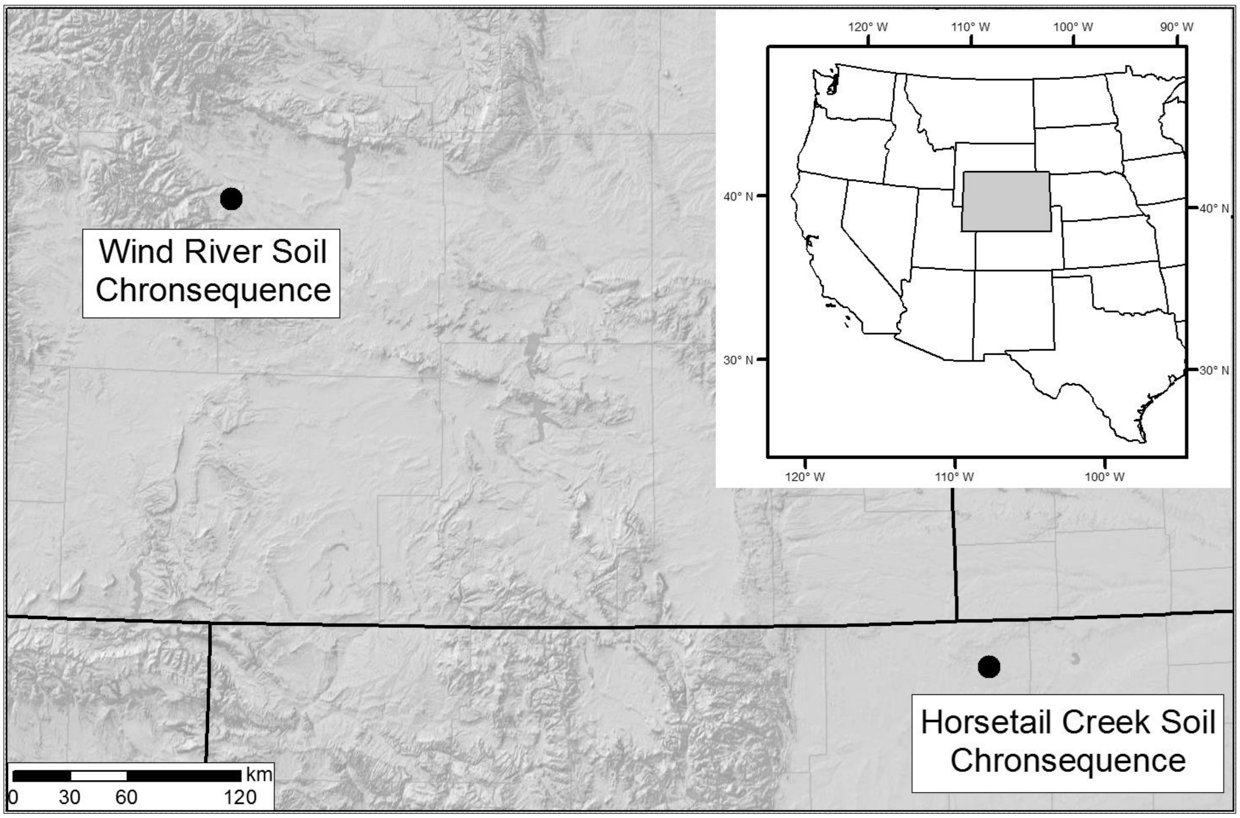

Figure 2-1. Locations of soil-chronosequences in Colorado and Wyoming.

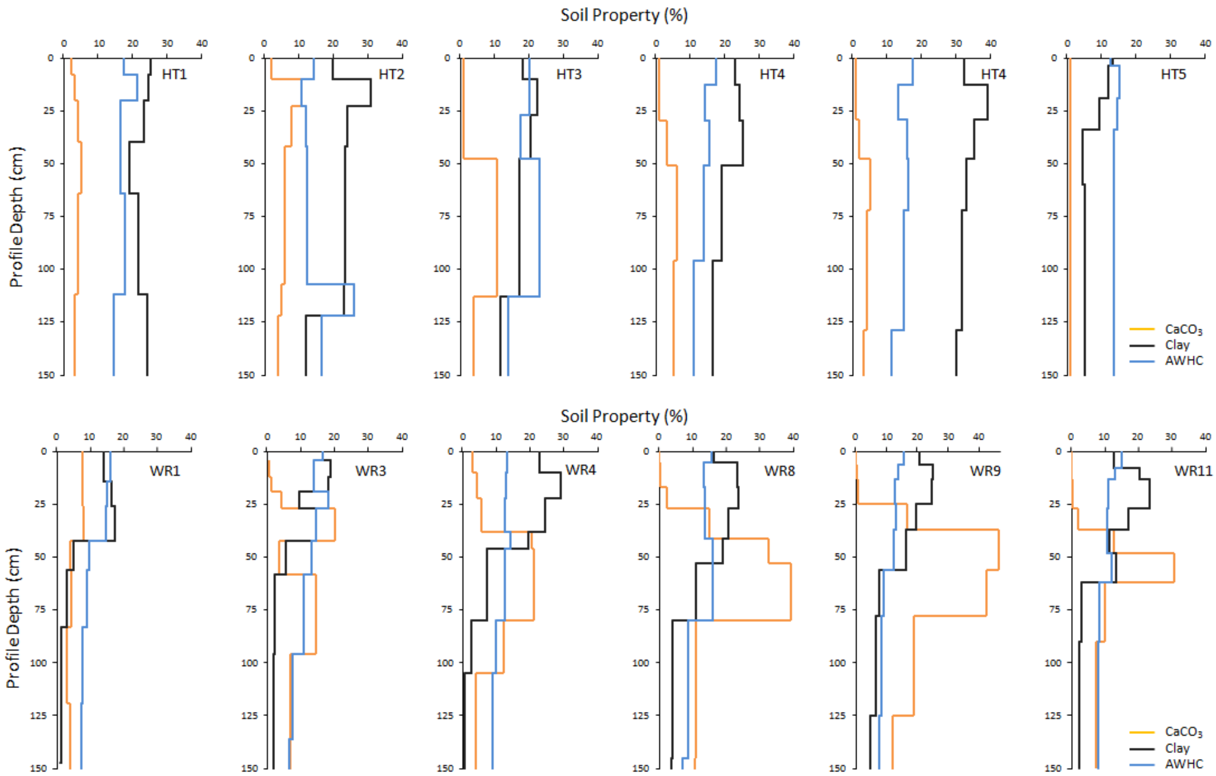

Figure 2-2. Depth profile plots of carbonates, clay, and total avilable water holding capacity (difference of %FC and %WP) for representative soils of different soil surface from each terrace. See Table 2-1 and 2-4 and text for age determaintions.

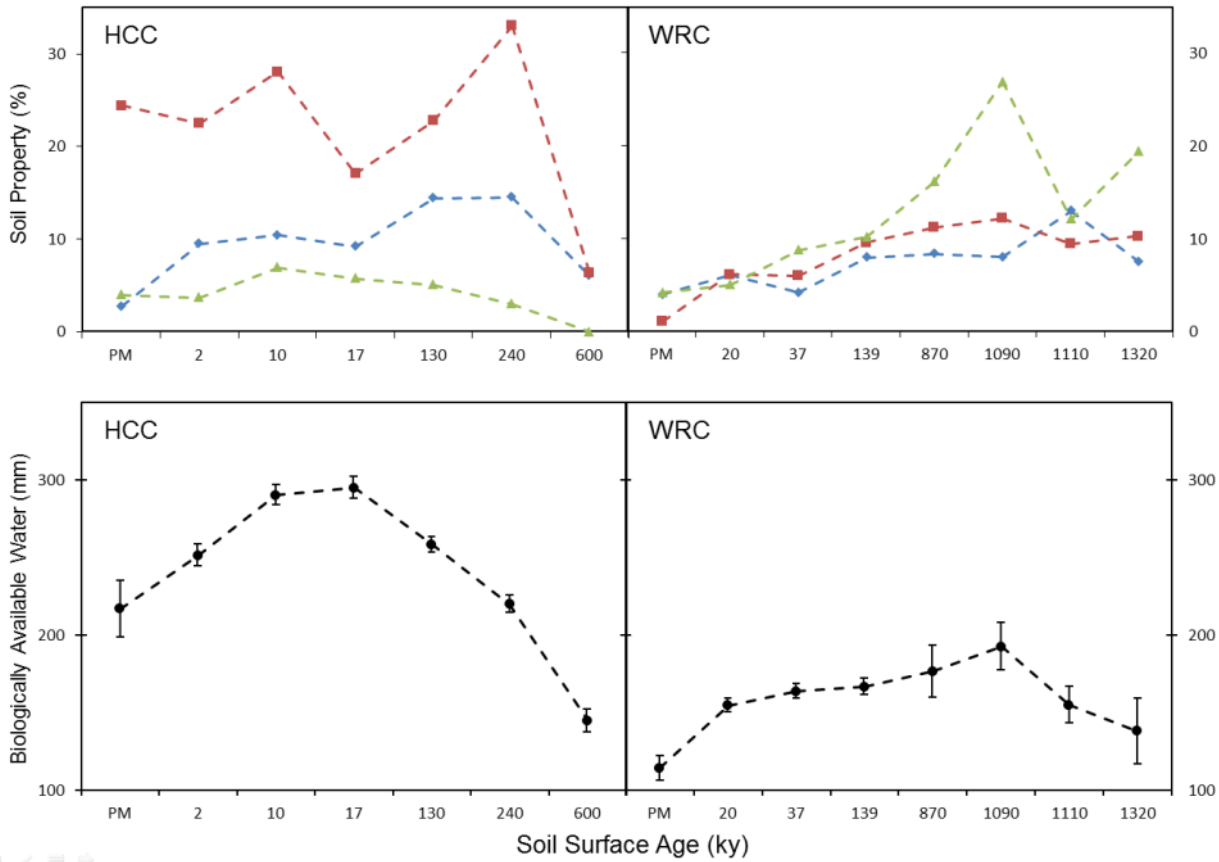

Figure 2-3. Biologically available water and soil physical properties across soil-age gradients at the Horsetail Creek chronosequence (HCC) in northern Colorado and the Wind River chronosequence (WRC) in central Wyoming. Approximate soil age is in thousands of years and PM is the modeled parent material of the youngest C horizon as a zero point. Data is expressed in weighted mean percent indices for the profile with clay (red), organic matter (blue expressed as $\%^{10}$), carbonates (green). BAW expressed in mm of water.

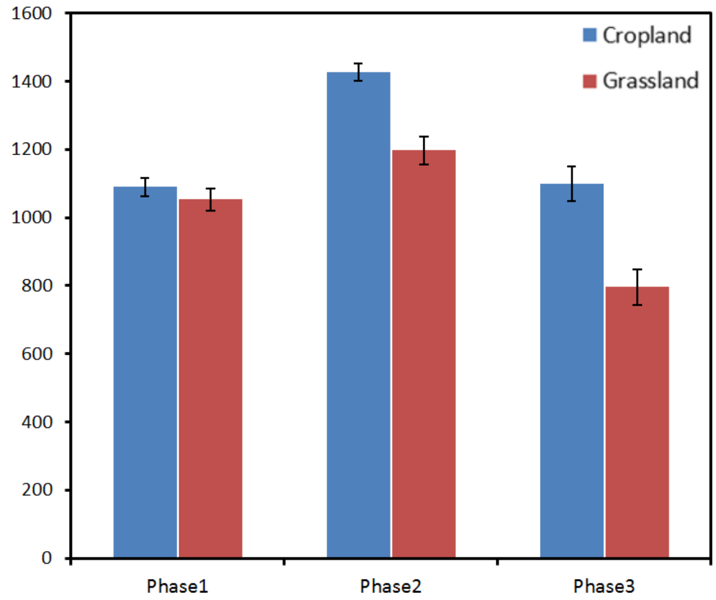

Figure 2-4. Plant production that was remotely sensed across landform age gradients near the Horsetail Creek chronosequence using NRCS-SSURGO rangeland production values. Error bars represent standard error.

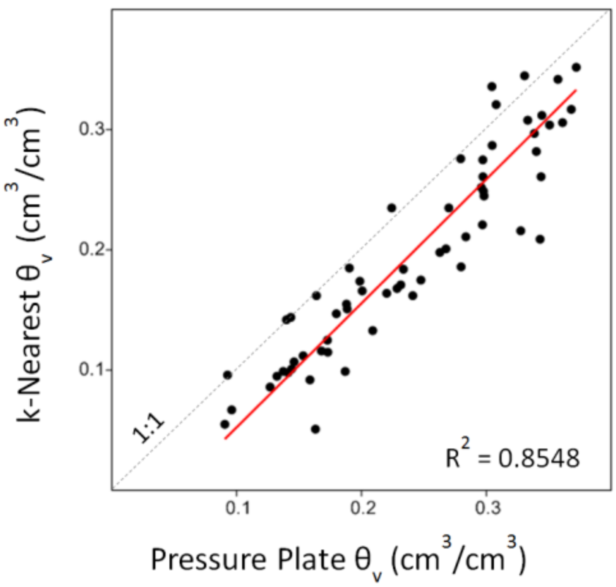

Figure 2-5. Linear regression of k-Nearest modeled and measured soil hydraulic properties of the Horsetail Creek chronosequence. Volumetric water parameters model lower than k-Nearest predicted values

Table 2-1: Morphology and physical features of soils in the Horsetail Creek chronosequence (HCC) and the Wind River chronosequence (WRC).

Terrace	Soil Classification[1,4]	Terrace Unit[1,4]	Elev. above Modern Stream (m)[3,4]	Age (kyrs)[1,2,5]
Horsetail Creek Chronosequence				
HC1	Haplocalcidic Usteochrept	Post-Piney Creek	7	<2
HC2	Aridic Haplustalf	Piney Creek	14	<10
HC3	Aridic Argiustoll	Wind Rework Broadway	18	13-20
HC4	Aridic Argiustoll	Slocum (younger)	19	130
HC5	Aridic Argiustoll	Slocum (older)	28	190-300
HC6	Aridic Haplustalf	Verdos	70	600
Wind River Chronosequence				
WR1	Typic Camborthid	LN4	6	18
WR3	Typic Calciorthid	BLN14	40	100
WR4	Typic Haplargid	PB1	43	130
WR8	Typic Haplargid	BLS3	156	156
WR11	Typic Haplargid	CR5	189	1090
WR12	Typic Haplargid	CR6	192	1110
WR13	Typic Haplargid	BLS2	223	1320

[1] Loadholt, 2002; [2]Blecker, 1994; [3]Madole, 1998, [4] Peacock, 1994; [5]Chadwick, 1997;

Table 2-2: Physical, chemical, and morphological characteristics of soil profiles on terraces of Horse Tail Creek.

	Depth	Texture (%)			C	$CaCO_3$	pH	P_{OD}	ρ_{33}	Water Content		WRD
	(cm)	sand	silt	clay				g/cm^3	g/cm^3	-33 kPa	-1500 kPa	cm^3/cm^3
HT_1 – 2 ka												
A	0-8	44.3	30.5	25.2	1.60	2	7.8	1.28	1.14	0.308	0.155	0.174
Bw1	8-20	41.9	33.4	24.7	0.95	3	8.0	1.27	1.1	0.345	0.151	0.213
Bw2	20-40	50.9	25.7	23.4	0.75	4	8.2	1.19	1.09	0.297	0.147	0.164
Bk1	40-64	56.0	24.9	19.1	0.84	5	8.6	1.25	1.16	0.287	0.144	0.166
Bk2	64-112	49.9	28.5	21.6	0.69	4	8.4	1.24	1.11	0.321	0.162	0.176
Bk3	112-162	45.6	30.3	24.1	0.13	3	8.0	1.24	1.14	0.312	0.185	0.145
C	162-200	44.5	31.1	24.4	0.28	4	8.5	1.22	1.1	0.342	0.174	0.185
HT_2 – 10 ka												
A	0-10	54.0	26.1	19.9	1.08	2	8.0	1.41	1.27	0.211	0.098	0.144
Bt	10-23	42.4	26.6	31.0	0.84	11	8.1	1.53	1.33	0.216	0.133	0.110
Btk	23-42	54.7	21.2	24.1	0.93	8	8.3	1.43	1.12	0.209	0.099	0.123
Bk	42-107	51.1	25.3	23.6	0.30	6	8.3	1.46	1.34	0.186	0.092	0.126
2Bk	107-122	42.8	34.1	23.1	0.47	5	8.2	1.48	1.31	0.252	0.051	0.263
2BCk	122-182+	73.1	14.7	12.2	0.27	4	8.2	1.38	1.21	0.235	0.096	0.168
HT_3 – 17 ka												
A	0-10	52.1	29.7	18.2	1.17	0	6.8	1.29	1.19	0.275	0.101	0.204
Bt1	10-27	52.3	25.2	22.5	1.13	0	7.3	1.43	1.34	0.276	0.112	0.202
Bt2	27-48	53.4	25.9	20.7	0.64	0	7.8	1.54	1.28	0.235	0.107	0.177
Bk1	48-113	38.0	44.6	17.4	0.87	11	8.3	1.26	1.39	0.336	0.142	0.233
Bk2	113-168	70.5	17.7	11.8	0.30	4	8.7	1.43	1.43	0.168	0.067	0.141
C	168-200	56.1	30.4	13.5	1.07	5	8.6	1.24	-	0.249	0.095	0.185
HT_4 – 130 ka												
A	0-13	48.1	28.9	23.0	1.18	0	7.1	1.32	1.17	0.261	0.116	0.173
Bt	13-30	51.0	24.6	24.4	0.78	0	7.8	1.52	1.23	0.221	0.115	0.142
Btk	30-51	51.3	23.5	25.2	0.47	3	8.0	1.44	1.38	0.245	0.125	0.154
Bk	51-96	59.5	21.5	19.0	0.39	6	8.5	1.47	1.2	0.198	0.099	0.138
BCk	96-162	61.8	21.7	16.5	0.48	5	8.5	1.47	1.4	0.162	0.086	0.109
C	162-177+	74.3	12.5	13.2	0.45	2	8.3	1.50	1.2	-	-	-
HT_5 – 240 ka												
A	0-13	33.4	34.1	32.5	2.12	0	6.6	1.26	1.03	0.352	0.184	0.173
Bt1	13-29	36.1	24.7	39.2	1.58	0	7.4	1.39	1.13	0.317	0.201	0.131
Bt2	29-48	44.5	20.0	35.5	1.00	2	8.1	1.41	1.2	0.306	0.175	0.157
Btk	48-72	43.2	23.6	33.2	0.40	5	8.3	1.37	1.21	0.304	0.171	0.161
Bk	72-129	45.3	23.0	31.7	0.73	4	8.5	1.39	1.26	0.282	0.164	0.149
BCk	129-200	46.0	24.0	30.0	0.82	3	8.2	1.25	1.18	0.261	0.166	0.112
HT_6 – 600 ka												
A	0-4	63.7	23.0	13.3	1.50	0	5.8	[1]1.40	[1]1.27	[2]0.191	[2]0.074	1.271
Bt1	4-19	85.9	2.3	11.8	1.01	0	7.0	[1]1.79	[1]1.52	[2]0.126	[2]0.055	1.516
Bt2	19-34	86.8	3.9	9.3	0.42	0	7.3	1.69	1.44	[2]0.109	[2]0.044	1.444
BC	34-60	93.6	1.9	4.5	0.19	0	7.3	[1]1.50	[1]1.36	[2]0.093	[2]0.025	1.362
C	60-200	93.4	1.4	5.2	0.13	0	7.4	[1]1.50	[1]1.36	[2]0.096	[2]0.026	1.362

[1]Estimated based on texture; [2]Modeled using Pedotransfer function "K-Nearest Neighbor"

Table 2-3: Physical, chemical, and morphological characteristics of soil profiles on terraces of the Wind River.

	Depth	Texture (%)			C	$CaCO_3$	pH	P_{OD}	Water Content		WRD
	(cm)	sand	silt	clay				g/cm^3	-33 kPa	-1500 kPa	cm^3/cm^3
WR1 – 20 ka											
A	0-14	58.6	27.4	14.0	0.91	7.9	7.7	1.20	0.296	0.1355	0.242
Abk	14-26	50.9	32.8	16.3	0.62	7.7	7.7	1.30	0.286	0.1352	0.212
Bk1	26-42	47.0	35.5	17.5	0.48	8.0	7.9	1.30	0.284	0.136	0.276
2Bk2	42-56	89.0	5.8	5.2	0.36	4.1	8.0	1.40	0.148	0.0482	0.175
2Bk3	56-83	93.0	3.7	3.3	0.29	4.6	7.9	1.50	0.128	0.0365	0.335
3Bk4	83-119	96.1	2.4	1.5	0.18	3.2	7.8	1.60	0.103	0.0248	0.407
4Bk5	119-147	95.3	3.3	1.4	0.21	4.1	8.3	1.60	0.102	0.0263	0.307
4Bk6	147-190	95.8	3.1	1.1	0.40	4.2	8.5	1.60	0.1	0.0251	0.032
WR3 – 37 ka											
A	0-4	52.0	31.6	16.4	0.21	0.2	7.1	1.20	0.309	0.1468	0.070
AB	4-12	50.2	31.1	18.7	0.91	0.4	7.2	1.36	0.277	0.144	0.130
Bw	12-19	51.8	30.0	18.2	0.93	1.2	7.5	1.38	0.283	0.1464	0.118
2Bk1	19-27	55.9	34.7	9.4	0.30	4.0	7.7	1.30	0.279	0.0994	0.168
2Bk2	27-42	59.2	26.3	14.5	0.20	20.2	7.6	1.30	0.276	0.1302	0.257
3Bkm	42-58	86.6	7.8	5.6	0.13	3.6	8.2	1.29	0.187	0.0557	0.244
3Bkq1	58-96	95.0	2.7	2.3	0.13	14.5	8.2	1.40	0.139	0.0319	0.512
3Bkq2	96-136	94.8	3.4	1.8	0.10	6.7	8.5	1.60	0.104	0.028	0.435
3BCk	136-210	95.2	3.1	1.7	0.12	6.7	8.5	1.70	0.093	0.027	0.140
WR4 – 139 ka											
A	0-10	51.9	25.3	22.8	0.77	2.9	7.5	1.36	0.28	0.1537	0.155
Abtk	10-22	45.8	25.0	29.2	0.75	4.2	7.8	1.36	0.321	0.1955	0.184
Btk1	22-38	50.2	25.3	24.5	0.65	5.6	8.0	1.38	0.288	0.1659	0.242
2Btk2	38-46	62.6	18.0	19.4	1.19	20.4	8.1	1.30	0.278	0.1372	0.132
3Bkm	46-80	82.7	10.2	7.1	0.86	21.1	8.1	1.29	0.2	0.0716	0.508
3Bk1	80-105	91.9	5.7	2.4	0.92	12.2	8.0	1.40	0.137	0.038	0.311
4Bk2	105-164	96.3	3.3	0.4	0.14	3.9	8.3	1.50	0.115	0.0264	0.540
WR8 – 870 ka											
A	0-5	58.8	24.7	16.5	0.78	0.2	6.9	1.20	0.302	0.1422	0.086
Bt	5-17	54.2	22.3	23.5	0.99	0.6	7.1	1.36	0.294	0.1629	0.192
Btk1	17-27	54.2	22.2	23.6	0.78	2.6	7.7	1.35	0.294	0.1637	0.158
Btk2	27-41	56.6	22.7	20.7	0.81	15.0	7.9	1.41	0.286	0.1502	0.242
Bkm1	41-53	61.4	19.7	18.9	0.78	32.7	8.1	1.18	0.306	0.1501	0.198
2Bkm2	53-80	73.9	14.9	11.2	0.64	39.4	8.3	1.17	0.246	0.0907	0.442
2Bkq1	80-145	92.3	3.6	4.1	0.25	11.2	8.6	1.50	0.127	0.0396	0.764
2Bkq2	145-185	92.0	4.4	3.6	0.22	10.6	8.6	1.60	0.108	0.0354	0.052
WR9 – 950 ka											
A	0-6	50.1	29.2	20.7	0.71	0.2	7.1	1.20	0.321	0.1656	0.101
Bw	6-13	48.4	26.3	25.3	0.94	0.4	7.2	1.29	0.318	0.1822	0.110
Bt1	13-25	50.7	24.6	24.7	0.89	0.7	7.4	1.35	0.290	0.1663	0.181
2Bt2	25-37	57.7	22.7	19.6	1.04	16.9	7.6	1.45	0.267	0.1424	0.195
2Bkm1	37-56	68.1	15.5	16.4	0.48	46.7	7.8	1.48	0.245	0.1254	0.302
2Bk1	56-78	83.0	9.2	7.8	0.97	42.8	7.9	1.50	0.153	0.0609	0.274
3km2	78-125	86.4	6.8	6.8	0.37	19.0	8.4	1.78	0.140	0.056	0.633
3Bk2	125-175	89.3	5.8	4.9	0.21	12.2	8.5	1.70	0.117	0.0429	0.282
WR11 – 1110 ka											
A	0-8	57.4	29.8	12.8	0.68	0.1	6.8	1.30	0.284	0.127	0.152
AB2	8-13	58.3	21.3	20.4	0.81	0.2	7.3	1.40	0.281	0.151	0.132
Bt	13-27	58.3	18.3	23.4	0.47	0.6	7.2	1.50	0.277	0.167	0.112
2Btk	27-37	70.4	12.5	17.1	0.43	2.3	8.1	1.60	0.227	0.122	0.109
2Bk1	37-48	77.8	10.6	11.6	0.41	12.7	7.8	1.50	0.189	0.085	0.108
2Bkm1	48-62	73.4	13.3	13.3	0.44	30.7	8.1	1.30	0.222	0.098	0.121
2Bkm2	62-90	84.5	12.2	3.3	0.35	10.1	8.4	1.60	0.133	0.049	0.085
3Bk2	90-120	90.9	6.5	2.6	0.28	7.5	8.6	1.70	0.118	0.038	0.081

2: REFERENCES

Austin, A.T., and O. E. Sala, 2002, Carbon and nitorgegn dynamics across a natural precipitation gradient in Patagonia, Argentina, Jouranl of Vegetation Science, 13:351-360.

Birkland, P.W., 1999, Soils and Geomorphhology, Oxford Press, 430 p.

Brimhall, G.H., Chadwick, O.A., Lewis, C.J., Compston, W., Williams, I.S., Danti, K.J., Dietrich, W.E., Power, M.E., Hendricks, D., Bratt, J., 1991b. Deformational mass transport and invasive processes in soil evolution. Science 255, 695–755.

Burke I.C., W.K. Lauenroth, R. Riggle, P. Brannen, B. Madigan, S Beard, 1999, Spatial variability of soil properties in the shortgrass steppe: the relative importance of topography, grazing, microsite, and plant species in controlling spatial patterns, Ecosystems 2:322-438.

Chadwick, O.A., et al, (ed) 1994. Quaternary Geology of the Wind River Basin, Wyoming, Friends of the Pleistocene - Rocky Mountain Cell, 141p.

Chadwick, O.A. L.A. Derry, P.M. Vitousek, B.J. Hubert, L.O. Hedin, Changing sources of nutrients during four million years of ecosystem development, Nature 397: 491-497.

Chadwick, O.A, J. Chorover, 2001, The chemistry of pedogenic thresholds, Geoderm 100: 321-353.

Chadwick, O.A., R.D. Hall, F.M. Phillips, 1997, Chronology of Pleistocene glacial advance in the central Rocky Mountains, GSA Bulletin, v109, n11, p. 1443-1452.

Davidson, E.A., L.V. Verchot, J.H. Cattanio, I.L. Ackerman, J.E.M. Carvalho, 2000, Effects of soil water content on soil respiration in forests and cattle pastures of eastern Amazonia, Biogeochemistry 45:53-69.

Dobos, R.R., H.R. Sinclair Jr, M.P. Robotham, 2012, National commodity Crop Productivity Index (NCCPI) User Guid, Version 2, USDA-NRCS.

D’odorico, P, A. Porporato (ed), 2006, Dryland Ecohydrology, Springer, 341 p.

Duniway, M.C., J.E. Herrick, H.C. Monger, 2010, Spatial and Temporal Variability of Plant-Available Water in Calcium Carbonate-Cemented Soils and Consequences for Arid Ecosystem Resilience, Oecologia, 163:215-226.

Gutiérrez-Jurado, H. A., E. R. Vivoni, J. B. J. Harrison, H. Guan, 2006, Ecohydrology of root zone water fluxes and soil development in complex semiarid rangelands, Hydrologic Processes, 20, 3289–3316.

Hancock, G.S., R.S. Anderson, O.A. Chadwick, R.C. Finkel, 1999, Dating fluvial terraces with ^{10}Be and ^{26}Al profiles: application to the Wind River, Wyoming, Geomorphology, 27: 41-60.

Harden, J.W., 1982. A quantitative index of soil development from field descriptions: examples from a chronosequence in central California. Geoderma 28, 1–28.

Hook, P.B., I.C. Burke 2000, Biogeochemistry in a shortgrass landscape: control by topography, soil texture, and microclimate, Ecology 81:2686-2703.

Hotchkiss, S., P.M. Vitousek, O.A. Chadwick, J. Price, 2000, Climate cycles, and geomorphological change, and the interpretation of soil and ecosystem development, Ecosystems.

Jenny, H., 1941, Factors of Soil Formation: A System of Quantitative Pedology. McGraw-Hill, New York, 281 p.

Jenny, H., 1980. The Soil Resource: Origin and Behaviour. Ecological Studies, Vol. 37. Springer, New York.

Johnson, E.A., and K. Miyanishi. 2008b. Testing the assumptions of chronosequences in succession. Ecology Letters 11: 419-431.

Johnson, D.L., E.A. Keller, T.K. Rockwell, 1990, Dynamic pedogenesis: New views on some key soil concepts, and a model for interpreting quaternary soils, Quaternary Research 33: 306-319.

Kaye, J.P., J.E. Barrett, and I.C. Burke. 2002. Stable carbon and nitrogen pools in grassland soils of variable texture and carbon content. Ecosystems 5: 461-471.

Lauenroth, W.K., I.C. Burke, J.A. Morgan, 2008, The shortgrass steppe, *in* Ecology of the Shortgrass steppe: a long-term perspective, 3-14p.

Loadholt, S.W., 2002, Soil Physiography of a Terrace Chronsequence in the Pawnee National Grasslands, Colorado, PhD Dissertation, Colorado State University, 198 p.

McFadden, L.D., 2013, Strongly dust-influenced soils and what they tell us about landscape dynamics in vegetated aridlands of the southwestern United States, in Brickford, M.E. ed., The Web of Geological Sciences: Advances, Impacts, and Interactions: Geological Society of America Special Paper 500, p. 501-532.

Nemes, A., W.J. Rawls, and Ya. A. Pachepsky. 2006. Use of the non-parametric nearest neighbor pproach to estimate soil hydraulic properties. Soil Science Society of America Journal 70: 327-336.

Nemes, A., R.T. Roberts, W.J. Rawls, Ua.A. Pachepsky, M.Th. van Genuchten, 2008, Software to estimate 33 and 1500 kPa soil water retention using the non-parametric k-Nearest Neighbor technique, Environmental Modeling & Software, 23, 254-255.

Noy-Meir, I., 1973, Desert ecosystems: Environment and Producers, Annual Review of Ecology and Systematics, Volume 4: 25-51.

Odening, W.R., B.R. Strain, W.C. Oechel, 1974, The effect of decreasing water potential on net CO2 exchange of intact desert shrubs, Ecology, 55: 1086-1095.

Peltzer DA, Wardle DA, Allison VJ, Baisden WT, Bardgett RD, Chadwick OA, Condron LM, Parfitt RL, Porder S, Richardson SJ, Turner BL, Vitousek PM, Walker J, Walker LR, 2010, Understanding ecosystem retrogression. Ecol Monogr 80:509–529.

Phillips, J. D., 1992, Progressive and Regressive pedogenesis and complex soil evolution, Quaternary Research 40, 169-176.

Peacock, C.R., 1994, Soil Evolution on a Terrace Chronosequence in the Wind River Basin, Wyoming, Masters Thesis, Colorado State University, 140p.

Pickett, S.T.A., 1989, Space-for-Time Substitution as an Alternative to Long-Term Studies, in Long-term Studies in Ecology (ed. G.E.Likens), pp. 110–135. Springer, New York.

Pockman, W.T., and J.S. Sperry. 2000. Vulnerability to xylem cavitation and the distribution of Sonoran Desert vegetation. Am. J. Bot. 87:1287–1299.

PRISM Climate Group, 2014, Oregon State University, http://prism.oregonstate.edu.

Riley, R.H., P.M. Vitousek, 1995, Nutrient dynamics and nitrogen trace gas flux during ecosystem development in montane rain forest, Ecology 76:292-304.

Robertson, G. P., and P. M. Vitousek. 1981. Nitrification in primary and secondary succession. Ecology 62:376-386.

Romano, N., and A. Santini. 2002. Water retention and storage: Field. p. 721–738. In J.H. Dane and G.C. Topp (ed.) Methods of soil analysis. Part 4. SSSA Book Ser. 5. SSSA, Madison, WI.

Reheis, M.C., 1980, Loess sources and loessial soil changes on a downwind transect, Boulder-Lafayette area, Colorado, Mountain Geologist, v17,n1, p. 7-12.

Sala, O.E., W.J. Parton, L.A. Joyce, W.K. Lauenroth, 1988, Primary production of the central grassland region of the United States, Ecology, 69: 40-45.

Scott, G.R, 1982, Paleovalley and Geologic Map of Northeastern Colorado, USGS Miscellaneous Investigation Series, Map I-1378.

Schlesinger, W. H., J. F. Reynolds, G. L. Cunningham, L. F. Huenneke, W. M. Jarrell, R. A. Virginia, and W. G. Whitford. 1990. Biological feedbacks in global desertification. Science 247(4946):1043–1048.

Schwinning, S., O.E. Sala, 2004, Responses to resource pulses in arid and semi-arid ecosystems. Oecologia 141: 211-220.

Selmants, P.C., S.C. Hart, 2008, Substrate age and tree islands influence carbon and nitrogen dynamics across a retrogressive semiarid chronosequence, Global Biogeochemical Cycles, 22.

Senock, R.S., D.L. Devine, W.B. Sisson, and G.B. Donart. 1994, Ecophysiology of three C4 perennial grasses in the northern Chihuahuan Desert. Southwest. Nat. 39:122–127.

Soil Survey Staff, 1996, Soil Survey Laboratory Methods Manual, Soil Survey Investigation Report No. 42.

Soil Survey Staff, 1997, National Characterization Data, Soil Survey Laboratory, National Soil Survey Center, and Natural Resources Conservation Service, Lincoln, NE.

Soil Survey Staff, 2009, Soil Survey Field and Laboratory Methods Manual. Soil Survey Investigations Report No. 51, Version 1.0. R. Burt (ed.), U.S. Department of Agriculture, Natural Resources Conservation Service.

Stevens, P. R., and T. W. Walker. “The chronosequence concept and soil formation.” Quarterly Review of Biology 45 (1970): 333-350.

Syers, J.K., Adams, J.A. and Walker, T.W., 1970a. Accumulation of organic matter in a chronosequence of soils developed on wind-blown sand in New Zealand. J. Soil Sci., 21: 146--153.

Thompson, CH, 1981, Podzol chronosequences on coastal sand dunes of eastern Australia, Nature 291:59–61.

Vitousek, P. M. 2004. Nutrient cycling and limitation. Hawaii as a model system. Princeton University Press. Princeton, New Jersey, USA.

Vreeken, W.G., 1975, Principle kinds of chronosequences and their significance in soil history, Journal Soil Science, 26:4, 378-393

Walker, T W, J K. Syers, 1976, The fate of phosphorus during pedogenesis, Geoderma 15:1–19.

Walker, LR and R del Moral 2003 Primary succession and ecosystem rehabilitation Cambridge University Press, Cambridge UK.

Wardle et al 2004, Ecological Linkages Between Aboveground and Belowground Biota, Science

Waring, R. H., and R. W. Running (2007), Forest Ecosystems: Analysis at Multiple Scales, 3rd ed., Academic, San Diego, Calif.

Wilcox, B. P., Breshears, D. D., & Allen, C. D. (2003). Ecohydrology of a resource-conserving semiarid woodland: Effects of scale and disturbance. Ecological Monographs, 73, 223–239.

CHAPTER 3

ROLE OF BIOLOGICALLY AVAILABLE WATER IN ECOSYSTEM DEVELOPMENT[2]

3.1 Summary

Changes in soil properties due to long-term pedogenesis are well documented regarding the trajectory of nutrients, cycling rates, and soil morphology. It is also established that soil moisture is the major control of ecosystem structure and function, and it is considered the most limiting resource to biological activity globally in ecosystems. Over timescales of thousands to millions of years, long-term changes in soil bulk density, volumetric dilation and collapse, mineral composition, and organic matter ultimately drive shifts in the ecosystem's plant available water capacity. Yet studies of soil pedogenesis and ecosystem development have largely ignored long-term alterations of plant available water in soils. To test predictions of ecosystem development on biologically available water, I modeled soil hydrologic properties for nine established soil-chronosequences across semi-arid to arid climates of North America. Results show that initial stages of soil formation generally increase water holding capacity, and absent of rejuvenating disturbance, pedogenesis ultimately leads to a total reduction in biologically available water on longtime scales. We suggest that long-term development of ecosystems imparts a predictable pattern regarding changes in biologically available water and propose that soil hydraulic properties ultimately play an important role in the evolution of ecosystems, providing an important biophysical feedback to soil forming processes.

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3.2 Introduction

Understanding the role of soil development in long-term ecological research is necessary in assessing the vulnerability of ecosystems to global change drivers. Integrating pedology into studies of the biogeochemical and ecohydrologic processes across multiple ecosystems allows for a further quantification of long-term ecosystem processes that condition net primary productivity, organic matter decomposition, and rates of nutrient cycling. For example, in long-term studies (thousands to millions of years) of ecological systems it has been reported that most are initially limited by nitrogen (Vitousek and Howarth 1991) and ultimately, over longer timescales, limited by phosphorus (Walker and Syers 1976). While the magnitude of this pattern varies with climatic regimes and geologic substrate, systematic patterns of key ecosystem properties at the ecosystem level emerge over millennial time scales, including: consequences of primary productivity, microbial biomass, total soil organic carbon, total soil nitrogen, and available soil phosphorus (Peltzer et al. 2010).

Soils are ultimately conditioned by the amounts, intensity, and residence time of water that flows through ecosystems. Water is the main agent as which solids and soluble ions are transported within soils, and are a key regulator of the balance between the major drivers of plant productivity and decomposition. Furthermore, soil water controls ecosystem structure and function, and it is recognized as the greatest limiting resource to biological activity (Noy-Meir 1973). Yet studies of soil and ecosystem development tend to focus on the biogeochemical dynamics while often overlooking the soil's capacity to govern long-term changes in the availability of water for biological activity. Some key properties of soil have been related to long-term water dynamics. For example, certain types of soil organic matter can hold up to 20 times their weight in water (Reicosky 2005). Thus during ecosystem development, if organic matter content is at a maximum, soil water retention should peak relative to other stages of development. Despite these apparent changes over the long term, there has been no attempt to quantify how plant available water changes as a function of soil development within ecosystems. In addition, little is known about the other soil properties that may regulate systematic variations of soil moisture storage. These include, but are not limited to: 1) texture, 2) soil structure, 3) micro-porosity, 4) bulk density, 5) soil depth, and 6) clay content and mineralogy. All these properties individually and collectively interact with hydrologic inflows and outflows to augment soil moisture that is portioned for biological activity in ecosystems.

Because the timing and quantity of seasonal to annual water fluxes, including precipitation and evapotranspiration, strongly determine the rate and trajectory of soil development (Chadwick and Chorover 2001), ecosystem development is strongly conditioned by climate. In the absence of any rejuvenating processes, a general trend with increasing soil age appears in both humid and semi-arid ecosystems (Vitousek 2004; Selmants and Hart 2008) with progressive increases in organic matter, nitrogen, and biomass followed by a decline in these properties in older and more highly weathered soils. Vitousek and others (1997) proposed that these developmental stages could be portioned into broad functional pedologic phases (building, sustaining, and degrading phases) of ecosystem development. Although progress has been made in understanding the fate of ecosystem processes and properties, such as primary productivity, decomposition, and rates of nutrient cycling, across the long-term time scales necessary to understand ecosystem development, no studies have attempted to quantify the characteristics that control the behavior of biologically available soil water at these time scales.

In general, studies that investigate ecosystem development over long-term time scales (hundreds to millions of years) attempt to translate the spatial differences between soils into temporal differences by substituting space for time (Jenny 1980, Pickett 1989). These soil-chronosequences represent “genetically related” groups of soils evolving with similar vegetation, topography, parent material, and climate (Harden 1982). Several studies of soil-chronosequences have yielded significant insights into pedogenesis and long-term ecosystem development (Stevens and Walker 1970; Walker and Syers 1976; Thompson 1981; Vitousek 2004; Wardel et al. 2004). This approach has been criticized since it is impossible to assume that all sites have been conditioned by “time” (Picket 1989; Philips 1993; Johnson and Miyanishi 2008); however, consistent robust patterns and properties do emerge when looking at a broad range (e.g., climatically) of chronosequences.

In this study, we assess how water retention characteristics of soils in the form of biologically available water vary with degrees of soil development in ecosystems. To determine patterns of biologically available water we conducted an analysis of soil hydraulic properties from soil chronosequences across biomes in western North America. Utilizing a soil geophysical model (Walker and Syers 1976) and the phase soil development model (Vitousek et al. 1997), we hypothesize that water limitation for biological processes can be predicted as a function of soil age within ecosystems and is an important indicator of ecosystem vulnerability and decline. To test these predictions we modeled soil biologically available water (BAW) from physical (sand, silt, clay, and bulk density) and chemical (carbon) properties of soil pedons from established soil chronosequences in the Western United States in order to propose a new conceptual model linking the degree of soil development to climate for water-limited ecosystems.

3.3 Soil-Chronosequences: Background and Locations

This study models soil hydraulic properties from fifteen established soil chronosequences in the western United States of semi-arid, arid, continental-humid and Mediterranean climates (Table 3-1, Figure 3-1). These sequences span several different types of parent materials, have vegetation respective of the regional climate, and span varying ranges of soil-age gradients. Based on a state factor approach (Jenny 1941, 1980) each chronosequence represents a series of soils that can be ascribed to the influence of time alone as differences in other important soil forming factors (e.g., parent material, climate, biota, relief, and land use) were minimized across each age gradient.

The Horsetail Creek sequence in northeast Colorado (Loadholt 2002), Wind River sequence in central Wyoming (Peacock 1994), and Kane fans of northern Wyoming (Rejeis 1987a) are the semi-arid sequences. Arid sequences from the desert southwest include Jornada in southern New Mexico (Gile and Grossman 1979, Gile et al. 2003), a Mojave desert sequence in California (McDonald 1994), and a sequence in southern Nevada near Yucca Mountain (Taylor 1980). Three humid continental climate sequence were used from the Rendija plateau in New Mexico (McDonald 1996) and two sites along Rock Creek in Montana (Reheis 1987b), a lower elevation "basin" sequence and a "mountain front" sequence. Ages of soils within each chronosequence are reported based upon cited material and approximated when literature reports ranges of soil age.

3.4 Materials and Methods

Soil hydraulic properties for horizons from each soil surface were compiled from the seventeen chronosequences, and modeled using the PTF model "k- Nearest Neighbor" Version 1.00.02 (Nemes et al. 2008) at water contents at the -0.033 mPa and at -1.5 MPa matric potentials. This non-parametric technique used US-NRCS-SCS Soil Characterization Database (Soil Survey Staff 1997) as reference soils data for pattern recognition based on sand, silt, and clay, as well as soil bulk density (BD) and soil organic matter (OM) contents (Nemes et al. 2006). Input data included texture, BD, and OM variables, and was ran using bootstrap techniques to perform multiple subset sections on the reference dataset. For many sequences, BD and/or OM were unavailable from the cited literature and those soil horizons were run with only soil texture. Water retention differences are reported for both sites as the volume fraction (cm^3 cm^{-3}) of water retained in the soil between -0.033 MPa and -1.5 MPa suction.

Biologically available water holding capacity (BAW) values were calculated for each soil pedon using weighted mean percent profile data indices (Birkland 1999) by summing each horizon across the pedon to maximum rooting depth to find total available water holding capacity. Available water capacity is the capacity of the soil to hold water for use by most plants, calculated as:

$$BAW = \sum_{i=1}^{n} WRD_{\ i} \times z_i \tag{1}$$

where BAW is the total available water capacity for the pedon up to a standardized depth of 100 cm, *I* is the soil horizon (1, 2, 3…*n*), WRD is the calculated water retention difference in the *i*th horizon, and z_i is the horizon thickness of the *i*th horizon (cm, although it is commonly expressed in mm). If multiple pedons of the same age existed within a soil age gradient, an average of those values was used to represent BAW. A t=0 from each soil chronosequence representing initiation of soil formation were based upon modeling the C horizon of the youngest soil in the age gradient. Hydraulic properties of this horizon were then multiplied by a depth of 100 cm to get t=0 of the sequence. Error for each soil's BAW was calculated by square rooting the sum of squares (RSS) for standard deviations reported from the "k- Nearest Neighbor" PTF model. This was done for each horizon, multiplied by the horizons thickness, and RSS for the soil profile. Weighted mean percent clay content was calculated in a similar manner and reported as percent soil profile to depth of one meter.

The soil leaching intensity index (V/V_o) was calculated using the model developed by Chadwick and others (2003). The available pore volume (V_o) of the top meter is calculated by subtracting the bound water (<−1500 kPa water) from the total water-holding pore volume. V is the average annual depth of water penetration, and is calculated using a monthly water availability model described by Arkley (1963) and Birkeland (1999) with climate for each soil-age gradient approximated using PRISM (2014) climate data, and potential evapotranspiration values input into the water balance were derived using the Thornthwaite equation. Each chronosequence was further tested against a 2nd order polynomial model in ranked order against available water to test a non-linear trend with a concave function. This test was applied to all chronosequences reported in this chapter to see if the early and late aged soil surfaces contained less plant available water than the mid-stages of each soil chronosequence.

3.5 Results

We first present the results of biologically available water from the three contrasting climatic regimes: continental, semi-arid steppe, desert southwest. Each soil-age gradient is presented with respect to their changes in plant available water, profile weighted mean clay percentage, and leaching intensity. We then focus on the general timing of changes in biologically available water patterns observed related to soil texture and ranges in parent material age.

Continental: Parent material from the soils in humid continental climates all show high sandy textures with the Rendija Plateau having 95-98% sand, upper Rock Creek 80-95% sand, and lower Rock Creek 85-95% sand. Modeled BAW values for those parent materials likewise are generally low, ranging between 96.8mm to 106.3mm plant available water to 100 cm (figure 3-1; table 3-2).

The Rendija soil chronosequences show the greatest increase among the three sites with BAW increasing by almost 10 cm of water up to 195.2 mm at its peak in the 75.8 ky site. Over that period it also shows a 14% increase in clay content and gradual decreases in leaching intensity from the parent material, decreasing from 2.2 to 0.3. Between the 157.7 ky soil and the 225 ky soil, BAW shows a 10% decrease dropping from 19.5 cm to 17.7 cm.

The mountain soil chronosequence along Rock Creek, Montana exhibits the least variation of BAW with only a 13.6 cm increase between the parent material and the 120ky soil. The next oldest soil at 415ky shows a decline of BAW back to similar volumes as the parent material and the younger soil sites. In general, the leaching intensity is also higher in these soils (figure 3-1 e), with average annual depth of water penetration (8-18cm) being higher due to the wetter climate, colder temperatures, and lower PET. In this sequence, when BAW is at its highest at the 120ky, leaching intensity is at its nadir (table 3-1) seen by both lower water penetration and higher available pore volume. A successive doubling of V/V_o is then seen so that leaching intensity increase from 1.1 to 2.3. Minor declines in profile clay percent was also seen at the oldest soil in the sequence. Rock Creek's basin (figure 3-1c, 3-1f) soils show declines in BAW similar to the mountain sequence, with a decrease in available water of 4cm water potential between the 120ky and 600ky soil. Leaching potential along this gradient varies between 0.5 and 1.4 with some variation.

Semi-arid Steppe: The three semi-arid gradients all exhibit lower BAW in their older soils than with the younger soils (figure 3-2; table 3-2). The Wind River sequence has a sandier parent material (approximately 95% sand) than Horsetail Creek and the Kane alluvial fans which have higher silt percentages (with 20%-50% silt C horizons on the Kane sites and 15%-30% silt C horizons on the Horsetail Creek sites). Soil hydraulic properties of Horsetail Creek and Wind River are described in greater detail in Chapter 2.

The Kane gradient shows a decline of 3.6 cm BAW between 6ky soil and the 65ky soil, and further declines to a low of 8.8cm BAW at the 315ky soil. The leaching intensity across all ages are low (0.2-0.4), and a general increase in clay content is measured. The Kane alluvial fans are notable because of their high gypsum contents (Reheis 1987a). It is unclear the role gypsum has on texture and water characteristics; furthermore, this sequence may lack the temporal resolution to detect nuanced patterns in soil hydraulic properties.

Horsetail Creek soils exhibits a general decrease in BAW from its parent material and youngest soil to the oldest soil. At the oldest soil (600ky), the Horsetail Creek sequence displays its lowest BAW of 7.75 cm (a 51% decrease). This soil also has the lowest clay percentages across the gradient and demonstrates an increase of V/V_o from 0.8 to 1.9. When the soil profile indexes are extend to 150 cm, an increase in BAW is observed (see chapter 2) with a peak in available water around the 17ky soil.

Soils on terraces above the Wind River do not see the same variation in BAW as the Kane and Horsetail Creek gradients. An increase from 11cm to 13cm available water occurs between the youngest to the soils at 870ky and 1090ky, followed by a decline 3.2cm to the 1110ky soil. Mean weighted profile clay percentage gradually decline within the same time frame (from 15% to 10% clay), and V/V_o is relatively minimal (0.5-0.9) with the exception of the parent material (1.9) which has low porosity (table 3-3).

Desert Southwest: The three soil chronosequences from the desert southwest climate have some of the lowest BAW values in the study (figure 3-3; table 3-3). Parent materials from these gradients are typically sandy (85% to 95% sand) with the Jornada gradient having the lowest available water (7.5cm) and the Yucca and Mojave sequences having 11.2cm and 11.6cm BAW. Overall leaching intensity values are similar to the semi-arid soils (table 3-2), however both average annual depth of water penetration and available pore volume values are much lower.

The Jornada sequence has a higher temporal resolution then other sites in this study and it shows increases in both BAW and Clay in the early to middle stages of pedogenesis (figure 3-3a, 3-3d) from 7.3cm to 11.6cm available water from the youngest soil to the 75ky soil. Clay percentages continue to increase until a maximum is reached at 231ky at 28% flowed by declines in both BAW and profile clay. During the peak of clay content and BAW, leaching intensity reaches a minimum of 0.5cm, a decrease from parent material's V/V_o of 1.3cm. V/V_o and then sees higher values above 1.0 in the oldest soil profiles where it has lower available pore volume and slightly higher water-penetrating depths (table 3-3).

The Mojave Desert soil sequence generally has lower BAW values from its parent material (figure 3-3b). Between 10ky and 30ky soils, the BAW drops from 96.4mm to 78.2mm and then starts to see increases in available water to 111.9mm. With progressive soil age, this sequence also sees a general increase in clay percentage (from 4% to 20.4%) which then drops to 7.9% at the oldest site (from 70ky to 135ky). Leaching intensities across the different soils are relatively uniform with values ranging between 0.5 and 1.2.

Soils near Yucca Mountain prominently show variable changes in clay content across the age gradient (Figure 3-3f). Values of V/V_o are low across the soil age gradient (0.3-0.5). A slight increase in BAW from the youngest soil (1.1ky) to the 10ky soil of 3.04cm available water is noticed, and after that soil age each older soil remains relatively uniform (around 114mm) despite the 10% to 20% changes in total clay.

3.6 Discussion

While there are generally established trends of the conditioning of soil properties and morphology associated with pedogenesis, scarcely any studies have examined patterns of soil hydraulic properties during long-term ecosystem development. Our analysis of nine established soil-chronosequences across semi-arid to arid climates in western North America show that the degree of soil development results in predictable patterns in biologically available water over time. Results show that initial stages of soil formation generally increase water holding capacity, and absent of rejuvenating disturbance, pedogenesis ultimately leads to a total reduction in biologically available water on longtime scales.

As recognized in previous studies of soil development in ecosystems, a general trend with pedogenesis appears with progressive increases in organic matter, nitrogen, and biomass followed by a decline in these properties in older and more highly weathered soils (Vitousek 2004). Established from patterns of long-term biogeochemical cycles, Vitousek and others (1997) proposed that these developmental stages could be portioned into broad functional pedologic phases (building, sustaining, and degrading phases) of ecosystem development. While their model of soil development focuses solely on biogeochemistry, our results concentrated on the soil-hydrologic properties, and through our analysis long-term trends of biologically available water similarly conform into distinct functional pedologic phases.

Mass balance analysis of soil formation has shown that in the early stages of soil development, a volumetric dilatation of the soil occurs due to biological and physical processes (Brimhall et al. 1992). Biotic controls that primarily drive soil dilation include accumulation of organic matter; expansion and mixing through growing roots; and the development of soil structure, aggregation, and porosity. Organic matter primarily adds an influx of carbon to the soil which in turn adds mass relative to the soil's parent material. This increases organic matter accumulation via soil aggregation and adsorption on mineral surfaces enhances soil structure, water holding capacity, and increases infiltration. Plant roots as well as other below-ground mechanisms expand the soil, create porosity, and generally assist in the mixing and expansion of the soil material.

This building phase is characterized by the loss of parent material/rock structure and a reduction of bulk density. In this building phase of soil development, our data show the amount of biologically available water increases with soil age, coinciding with the development of soil porosity and reductions in bulk density. The intensity of leaching is variable across the sequences in this study with some gradients seeing a rapid decline from the parent material suggesting quick formation of porosity and reductions of soil bulk density. In most cases, our sites may be deficient in the temporal resolution in the earliest stages of soil development to detect the initial conditioning of the soil hydraulic properties. Yet increases in BAW are observed across the soil-chronosequences in this study.

The middle stage of pedogenesis is in a quasi-equilibrium state where progressive soil forming processes are balanced with regressive soil destroying processes. The initial volumetric expansion caused by additions of carbon declines quickly while other progressive pedogenic processes persist (for example: horizonation, developmental upbuilding, and soil deepening). Depending on the drivers of the ecosystem, the duration of this stage can vary greatly. These soils are characterized by the onset of weathering product accumulation such as iron oxides, silicate clays, and carbonates and loses various weathering products. Our analysis suggests that while clay content impacts BAW, there is a poor connection between the timing of maximum clay percentage with maximum available water when indexed to 100cm. Leaching intensity decreases in these soils with high BAW mostly due to larger available pore volume. Within this stage, our results show that BAW reaches its greatest potential.

In the late stages of pedogenesis, absent of any rejuvenating disturbances, soil formation is characterized by volumetric collapse due to chemical losses of major elements which given sufficient time, result in nutrient limitation or senility of the landscape. The integrated effects of mineral weathering and leaching of elements (such as silicon, calcium, magnesium, sodium, and potassium) begin to become perceptible and over long-time scales eliminates the expansion leading to volumetric collapse. These retrogressive soils shift from primary mineral to secondary iron and aluminum oxides and become devoid of primary mineral phosphorus. Plant productivity and carbon inputs to these soils begin to decline due to fewer plant essential elements (i.e., phosphorus and calcium) which subsequently lead to slow declines in carbon and nitrogen.

Research from humid ecosystems propose that declines in soil organic matter quality is due to reduction in substrate quality (loss of phosphorus) and this feeds back to reduce decomposition and plant productivity (Wardle et al 2004; Williamson et al 2005). Recent research proposed that declines in organic matter in semi-arid ecosystems were not due to substrate quality, but to declines in the fluxes of carbon, nitrogen, and microbial biomass (Selmants and Hart 2008). Our results show that slowdown in organic matter is linked to changes in soil porosity. Reductions in organic matter are undoubtedly caused by reduced inputs of carbon and energy to the soil subsystem; however, we suggest that lower quantities in plant available water ultimately condition the decline in soil organic matter.

Few of the Earth’s surfaces exhibit the geomorphic stability leading to retrogressive landscapes, and moreover atmospheric inputs compensate globally for the chemical weathering losses (Chadwick et al. 1999) where ecosystems subsist on the steady slow flux of atmospherically derived elements. Our analysis shows that biologically available water ultimately declines compared to BAW peak in the middle stages of soil formation. Many of the age gradients we studied also showed declines in available water below the parent material and the youngest soils. As BAW declines the leaching intensity of soils generally increase due to deeper water penetration and less available pore volume. However, further analysis is needed to discriminate the mechanisms controlling magnitude soil hydraulic changes, especially in those instances where BAW on the oldest surfaces drop below water holding capacity of the soil parent material.

Timescales and the precise mechanisms driving patterns of progressive and retrogressive pedogenesis vary greatly over the contrasting climate regimes, geologic strata, and vegetation types. Perhaps the greatest limitation of our analysis is that each soil chronosequence is conditioned by wildly different parent materials and climatic conditions. However, a common observation can be made from our analysis: that climates with greater leaching intensities tend to condition a decline in BAW faster than systems with less available moisture. The duration of the sustaining phase in which soil BAW maintains its largest quantity of available water capacity tends to persist longer in dryer climates due to slower conditioning rates of the soil system. Perhaps of greater interest are the gradients with high magnitude of rate change in BAW properties, for example the semi-arid sequence Horsetail Creek sees a 58% reduction in available water (18.9 cm to 7.8 cm) from the peak to oldest soil.

3.7 Conclusion

Studies of ecosystem development have generally been deficient in linking biological processes and plant-soil feedbacks relative to physical and climatic processes controlling long-term ecosystem dynamics. Most studies linking above ground and below ground ecosystems to long-term developmental change have focused solely on pool size and fluxes of nutrients. Our biophysical model of biologically available water conditioned through long-term soil development provides insight to the development of soils. We suggest that with development of soils in ecosystems, a pattern arises regarding changes in biologically available water. Changes in soil bulk density, volumetric dilation and collapse, mineral composition, and organic matter drive shifts in the ecosystem's plant available water. Our data show that the initial stages of soil formation generally increase ecosystem available water, and absent of rejuvenating disturbance, pedogenesis untimely drives a total reduction in biologically available water on long-time scales. We conclude that the soil hydraulic properties ultimately play an important role in the development of ecosystems, and provide an important biophysical feedback to other processes of soil formation.

3.5: TABLES AND FIGURES

Table 3-1. Published long-term chronosequences in the Western U.S. for which hydraulic properties where modeled from soil morphologic data used in this study. Chronosequences are divided into three like climates. Geographic location estimated from approximate geographic middle of soil chronosequence. Mean annual precipitation and temperature estimated from PRISM (PRISM 2008). Climate was based upon the observed Koppen-Geiger classification (Kottek et al. 2006), and soil landform age and vegetation is contained within cited references found in section 3-7.

Table 3-1:

Chronosequences		Location	Latitude, Longitude		Parent Material	Age Gradients (ky)	Vegetation	Climate (Köppen)	Climate	Mean Annual T °C	Mean Annual P (mm)	References
Semiarid Steppe												
HTC	Horsetail Creek	Northeast Colorado	40.75	-103.67	Fluvial Terrace	2-600	Shortgrass	BSk	Semiarid Grassland Steppe	9.4	386.7	Loadholt (2002)
WR	Wind River	Central Wyoming	43.17	-109.04	Glacial Fluvial Terrace	20-1320	Sagebrush	BSk	Semiarid Shrub Steppe	5.6	253.5	Peacock (1994)
KANE	Kane	Northern Wyoming	44.87	-108.13	Gypsic Fans	6-585	Sagebrush/ grassland	BSk / Dfb	Humid Continental Wet all year	7.8	207.6	Reheis (1987a)
Continental												
RKCU	Rock Creek: Mountain	South Central Montana	45.17	-109.25	Glacial Fluvial	7-2000	Conifer, Aspen, Grass	Dfc/ Dfa	Subarctic Cool summer	6.0	561.9	Reheis (1987b)
RKCL	Rock Creek: Basin	South Central Montana	45.54	-108.87	Fluvial Terrace	7-2000	Sagebrush, grass	Dfa	Continental Hot Summer	8.2	341.0	Reheis (1987b)
REN	Rendija Plateau	Northern New Mexico	35.91	-106.28	Alluvial Terrace	1.1-225	Ponderosa pine	Dfb	Humid Continental Wet all year	9.8	508.3	McDonald (1996)
Desert Southwest												
JER	Jornada	Southern New Mexico	32.63	-106.74	Alluvial Terraces	1-2500	Mesquite shrubland	BSk/ BWk	Mid-latitude Desert	15.4	263.5	Gile & Grossman (1979) Gile et al (2003)
MOJ	Mojave	Southeast California	34.92	-115.63	Fluvial	1-135	Shrubland	Bwh	Subtropical Desert	18.1	151.7	McDonald (1994)
YUC	Yucca	Southern Nevada	36.87	-116.42	Fluvial	1.1-270	Shrubland	BWk	Mid-latitude Desert	15.5	153.0	Taylor (1980)

Table 3-2: Indexed values for Continental soil chronosequences on river terraces of the Rendija (REN), upper Rock Creek (RKCU), and lower Rock Creek (RKCL). Soil age is in thousands of years; biologically available water (BAW) is indexed to 100cm for each soil; standard deviation of BAW is reported based upon the root sum of squares from the k-Nearest PTF; clay percent is the weighted mean percent clay indexed to 100cm; organic matter percent (OM) is indexed to 10cm; V is the average annual depth of water penetration; V_o is the available pore volume in the top meter; and V/V_o is an index of leaching intensity.

	Age (ky)	BAW	BAW SD	Clay (%)	OM (%)	V	V_o	V/V_o
RKCL	PM	96.8	7.1	4.4	0.12	7.9	3.0	2.6
	7	125.5	4.1	12.4	3.16	5.0	5.8	0.9
	20	111.8	3.0	14.7	4.44	4.1	5.2	0.8
	120	140.9	3.0	21.6	12.26	5.0	10.1	0.5
	415	114.1	2.7	13.2	5.13	4.1	6.3	0.6
	600	100.8	2.7	17.1	3.87	7.2	5.0	1.4
	945	107.3	3.2	14.7	4.41	5.0	6.0	0.8
	2000	115.0	2.8	12.0	10.49	5.6	5.3	1.0
RKCU	PM	106.3	6.9	2.3	0.26	14.6	3.4	4.3
	7	103.5	3.3	1.6	4.67	18.5	5.1	3.7
	20	105.1	4.7	2.6	13.53	8.0	5.5	1.5
	120	119.9	4.9	8.2	9.24	8.6	8.2	1.1
	415	104.6	2.4	13.3	8.18	10.3	5.4	1.9
	600	105.8	3.4	14.6	5.74	14.4	6.4	2.3
	2000	99.9	3.6	9.9	9.86	11.8	5.2	2.3
REN	PM	100.4	9.9	1.0	-	8.9	4.1	2.2
	1.1	110.2	5.4	2.4	-	4.3	6.2	0.7
	3.2	140.2	8.3	3.9	-	3.7	7.0	0.5
	5.6	146.5	18.0	8.1	-	3.5	8.1	0.4
	75.8	195.2	4.1	13.6	-	3.3	9.6	0.3
	157.7	193.8	6.8	14.9	-	4.8	9.7	0.5
	225	176.8	8.0	14.0	-	3.9	8.2	0.5

Table 3-3: Indexed values for semi-arid soil chronosequences at Horsetail Creek (HTC), Wind River (WR), and the Kane alluvial fans, (KANE). Soil age is in thousands of years; biologically available water (BAW) is indexed to 100cm for each soil; standard deviation of BAW is reported based upon the root sum of squares from the k-Nearest PTF; clay percent is the weighted mean percent clay indexed to 100cm; organic matter percent (OM) is indexed to 10cm; V is the average annual depth of water penetration; V_o is the available pore volume in the top meter; and V/V_o is an index of leaching intensity.

	Age (ky)	BAW	BAW SD	Clay (%)	OM-10	V	Vo	V/Vo
HTC	PM	158.7	12.1	24.4	0.28	4.1	8.6	0.5
	2	151.0	4.8	22.0	2.53	4.4	8.0	0.6
	10	129.6	5.6	24.3	1.86	4.5	6.2	0.7
	17	148.4	5.3	19.0	2.01	4.2	7.4	0.6
	130	125.3	4.1	21.6	2.03	5.0	5.6	0.9
	240	117.8	3.4	34.1	3.65	4.7	5.8	0.8
	600	77.5	3.9	6.9	2.07	6.0	3.2	1.9
WRR	PM	115.2	5.4	1.1	0.12	8.2	3.0	2.8
	20	116.2	4.1	8.6	1.57	3.8	9.2	0.4
	37	127.7	4.5	8.2	1.08	3.7	9.7	0.4
	139	121.9	4.4	14.2	1.32	4.8	6.0	0.8
	870	133.9	6.5	15.0	1.52	3.8	6.9	0.6
	1090	132.2	5.3	15.4	1.21	3.8	6.4	0.6
	1110	100.0	4.3	13.2	1.63	4.0	5.0	0.8
	1320	102.0	3.2	10.0	1.40	3.7	6.1	0.6
KANE	PM	165.1	5.7	6.7	0.22	2.7	15.9	0.2
	6	168.4	3.0	12.6	1.06	2.8	16.1	0.2
	65	131.9	4.1	14.8	0.80	2.8	15.3	0.2
	100	124.9	2.5	19.6	0.71	2.7	11.8	0.2
	315	88.0	3.2	37.1	1.18	3.1	8.0	0.4
	410	108.0	3.1	30.5	0.97	3.0	9.9	0.3
	505	110.9	4.2	28.4	0.65	3.4	10.0	0.3
	585	108.0	3.8	33.1	1.00	3.3	9.9	0.3

Table 3-4: Indexed values for desert southwest soil chronosequences at Jornada (JER), the Mojave Desert (MOJ), and Yucca Mountain (YUC). Soil age is in thousands of years; biologically available water (BAW) is indexed to 100cm for each soil; standard deviation of BAW is reported based upon the root sum of squares from the k-Nearest PTF; clay percent is the weighted mean percent clay indexed to 100cm; organic matter percent (OM) is indexed to 10cm; V is the average annual depth of water penetration; V_o is the available pore volume in the top meter; and V/V_o is an index of leaching intensity.

	Age (ky)	BAW	BAW SD	Clay (%)	OM (%)	V	Vo	V/Vo
JER	PM	74.9	7.4	3.7	0.09	4.6	3.5	1.3
	7	73.6	3.9	6.2	0.29	4.2	3.5	1.2
	11	109.7	3.7	7.4	0.49	3.1	4.8	0.6
	50	109.2	3.6	10.2	0.21	2.7	5.0	0.5
	75	115.7	3.3	21.9	0.36	2.7	5.2	0.5
	184	99.7	4.6	29.4	0.17	3.7	3.9	0.9
	231	95.6	2.4	28.5	0.30	2.9	3.5	0.8
	625	101.9	5.4	20.0	0.97	2.6	4.1	0.6
	780	82.1	2.9	14.3	0.33	4.5	3.2	1.4
	1600	95.8	2.9	12.1	0.43	4.1	3.8	1.1
	2250	85.5	2.2	17.2	0.33	3.1	2.9	1.1
MOJ	PM	73.1	1.2	4.0	-	4.0	3.2	1.2
	1	86.4	4.1	4.5	-	2.9	3.5	0.8
	3.6	86.3	4.1	5.2	-	2.2	3.1	0.7
	10	96.4	3.4	8.2	-	2.9	3.5	0.8
	30	78.2	2.5	10.4	-	2.6	2.9	0.9
	70	86.4	3.1	20.4	-	3.1	2.6	1.2
	135	111.9	5.5	7.9	-	2.1	3.9	0.5
YUC	PM	112.4	7.3	2.3	-	1.3	3.3	0.4
	1.1	100.0	3.2	20.2	-	1.3	4.1	0.3
	8.3	115.4	10.4	4.9	1.35	1.7	5.4	0.3
	10	130.4	4.7	8.8	0.99	1.6	5.3	0.3
	30	114.4	3.3	27.5	-	1.4	5.1	0.3
	140	114.8	8.6	1.5	0.33	2.5	5.6	0.4
	145	113.8	3.1	13.0	0.45	2.2	4.0	0.5
	270	114.1	4.1	8.6	0.28	1.2	3.8	0.3

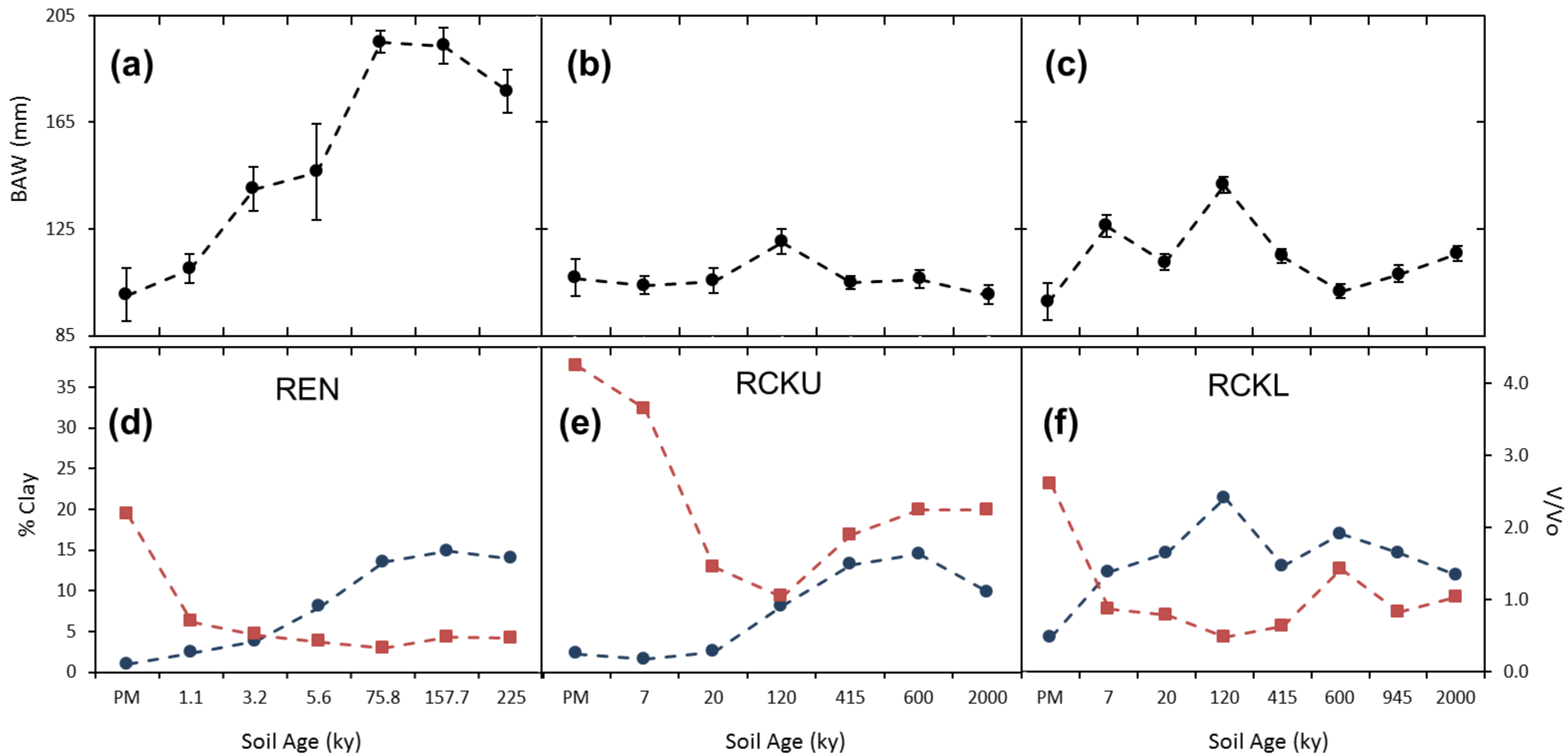

Figure 3-1: Biologically available water (upper), weighted mean percent clay (lower circle), and leaching intensity (lower square) for the top meter of the soil profile on humid continental soil-chronosequences on river terraces of the Rendija (REN), upper Rock Creek (RKCU), and lower Rock Creek (RKCL).

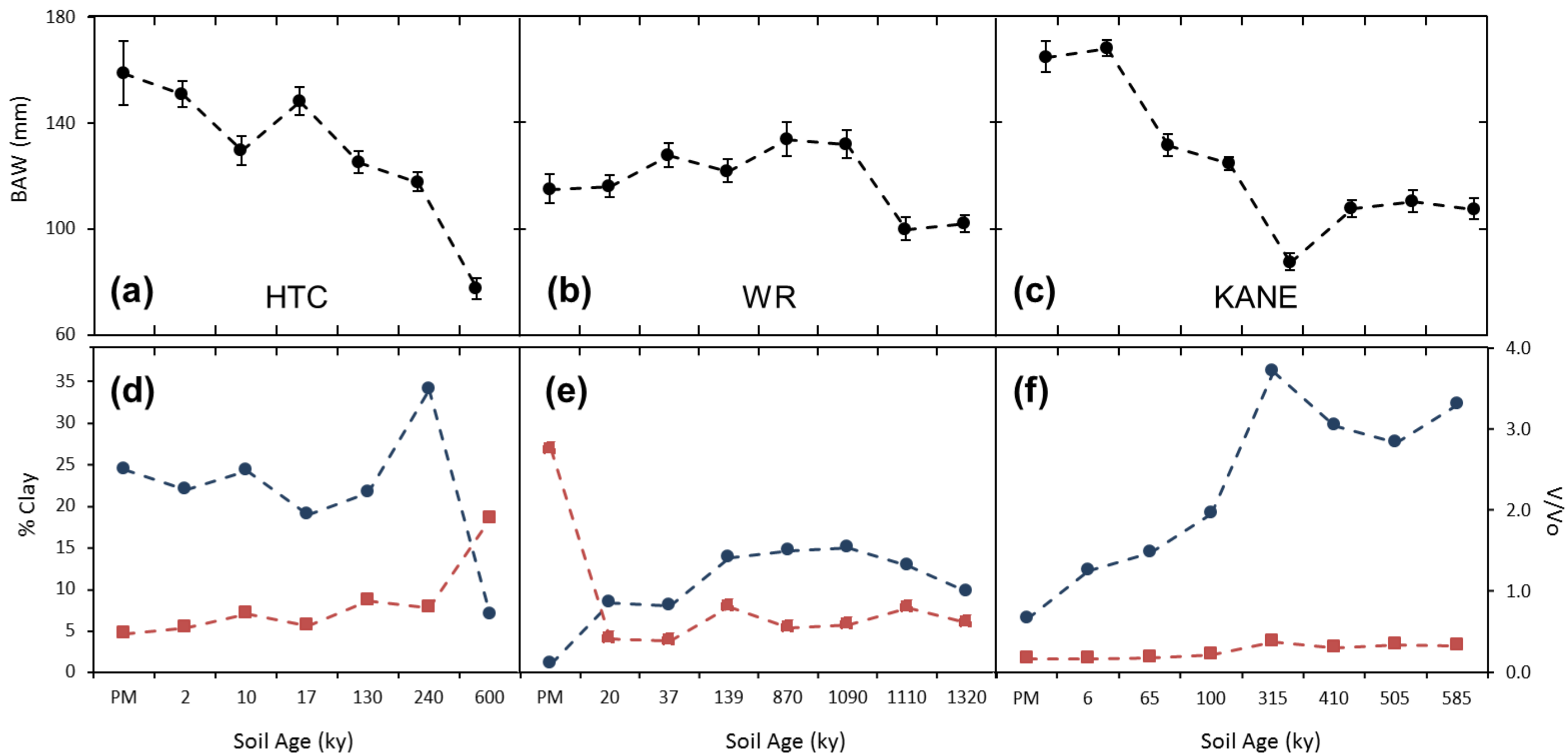

Figure 3-2: Biologically available water (upper), weighted mean percent clay (lower circle), and leaching intensity (lower square) for the top meter of the soil profile on semi-arid soil-chronosequences at Horsetail Creek (HTC), Wind River (WR), and the Kane alluvial fans, (KANE).

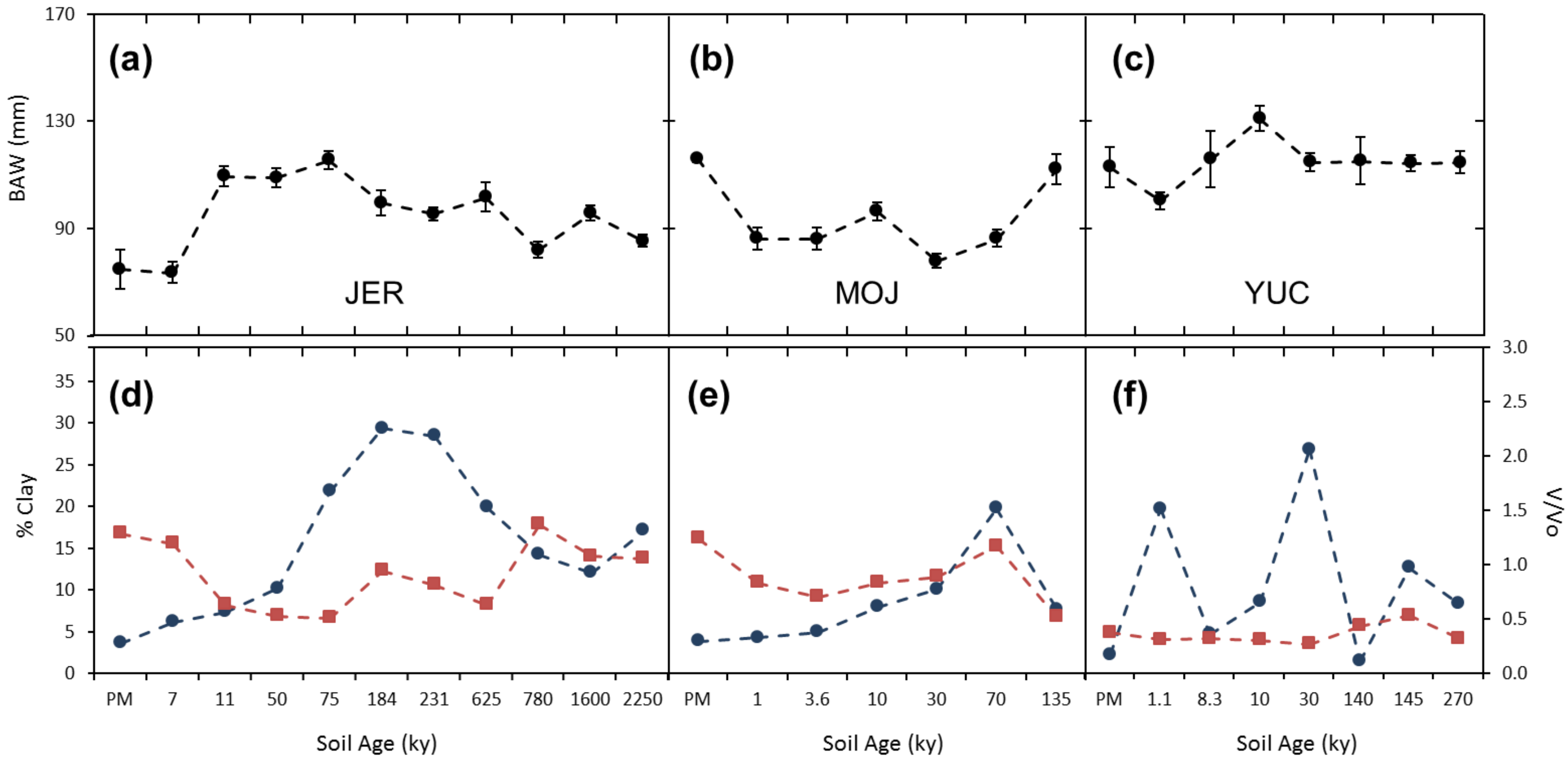

Figure 3-3: Biologically available water (upper), weighted mean percent clay (lower blue circle), and leaching intensity (lower red square) for the top meter of the soil profile on desert south west soil-chronosequences at Jornada (JER), the Mojave Desert (MOJ), and Yucca Mountain (YUC).

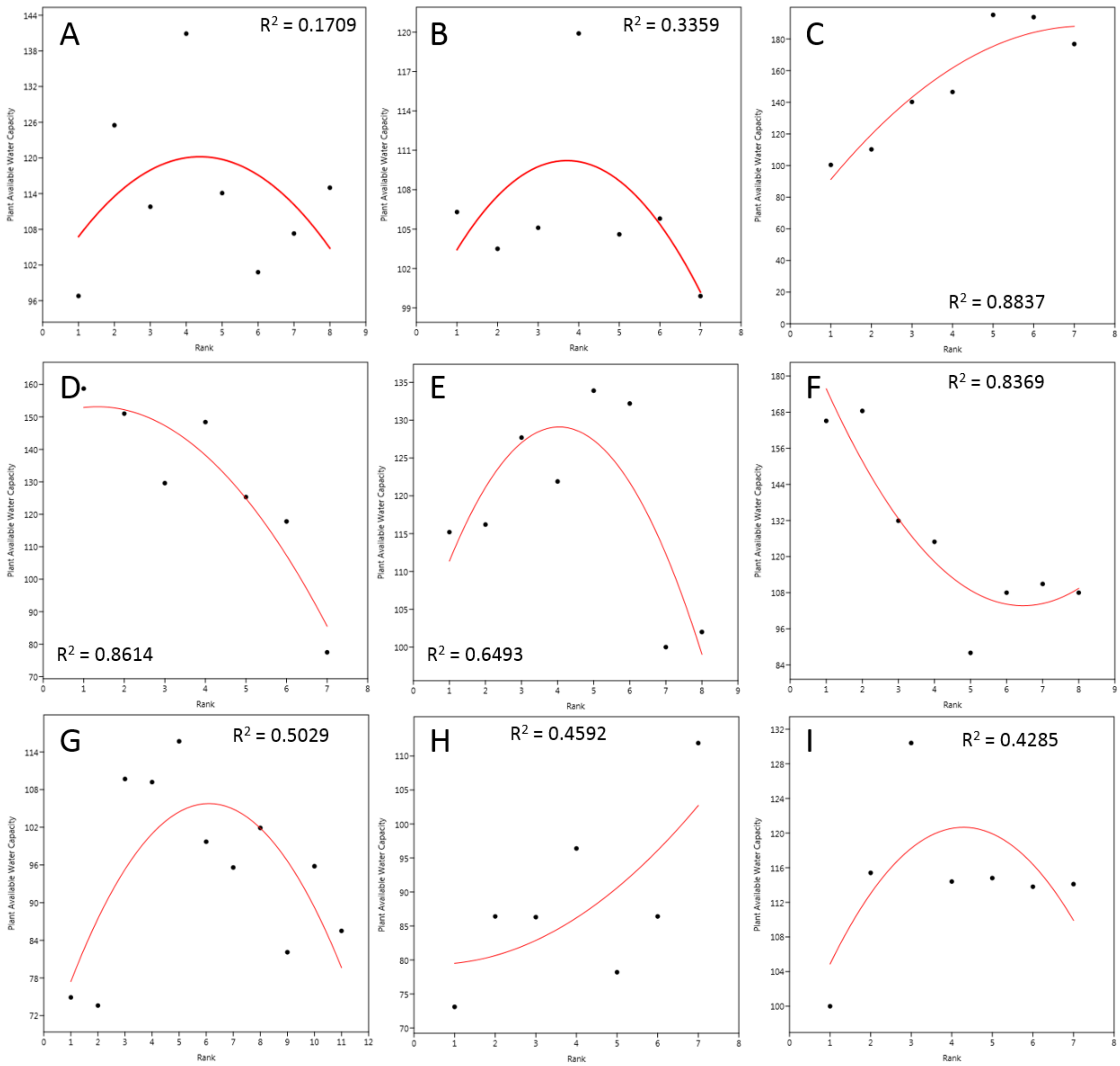

Figure 3-4: Soil-chronosequences to Plant Available Water within the top 100 cm. Each plot is modeled using ranked ordered second-order polynomial regressions of (A) Rock Creek Lower, (B) Rock Creek Upper, (C) Rendija, (d) Horsetail Creek, (E) Wind River Range, (F) Kane, (G) Jornada, (H) Mojave, (I) Yucca.

3: REFERENCES

Birkland, P.W., 1999, Soils and Geomorphology, 3rd ed., Oxford University Press, New York, NY, 430 p.

Chadwick, O.A. L.A. Derry, P.M. Vitousek, B.J. Hubert, L.O. Hedin, Changing sources of nutrients during four million years of ecosystem development, Nature 397: 491-497.Chadwick, O.A., J. Chorover, 2001, The chemistry of pedogenic thresholds, Geoderma, 100: 321-353.

Chadwick, O.A., R.T. Gavenda, E.F. Kelly, K Zielgler, C.G. Olson, W.C. Elliott, D.M. Hendricks, 2003, The impact of climate on the biogeochemical functioning of volcanic soils, Chemical Geology 2002: 195-223.

Dethier, DP, 1988, The soil chronosequence along the Cowlitz River, Washington, USGS Bulletin 1590-F, 47 p.

Gile LH, RB Grossman, 1979, The Desert Project Soil Monograph, U.S. Gov. Printing Office, Washington, D.C.

Gile LH, RJ Ahrens, SP Anderson, 2003, Supplement to the Desert Project Soil Monograph, Volume 3, National Soil Survey Center, Lincoln, NE.

Harden, JW, 1987, Soils developed in granitic alluvium near Merced, California, USGS Bulletin 1590-A, 65 p.

Harden JW, 1987, Forward to USGS Bulletin 1590.

Jenny, H., 1941, Factors of Soil Formation: A System of Quantitative Pedology. McGraw-Hill, New York, 281 p.

Jenny, H., 1980. The Soil Resource: Origin and Behaviour. Ecological Studies, Vol. 37. Springer, New York.

Johnson, E.A., K. Miyanishi, 2008, Testing the assumptions of chronosequences in succession, Ecology Letters, 11:05 419-431.

Kottek, M., J. Grieser, C. Beck, B. Rudolf, and F. Rubel, 2006: World Map of the Köppen-Geiger climate classification updated. Meteorol. Z., 15, 259-263.

Loadholt, S.W., 2002, Soil Physiography of a Terrace Chronosequence in the Pawnee National Grasslands, Colorado, PhD Dissertation, Colorado State University, 198 p.

McDonald, EV, 1994, The Relative Influences of climate change, desert dust, and lithologic control on soil-geomorphic processes and soil hydrology of calcic soils formed on quaternary alluvial fan deposits in the Mojave Desert, California, PhD Dissertation, University of New Mexico, 383 p.

McDonald, EV, SL Reneau, JN Gardner, 1996, Soil-Forming Processes on the Pajarito Plateau: Investigation of a Soil Chronosequence in Rendija Canyon, New Mexico Geological Society Guidebook, 47th Field Conference, Jemez Mountains Region, pg 367-374.

McFadden LD, RJ Weldon II, 1987, Rates and processes of soil development on quaternary terraces in Cajon Pass, California, GSA Bulletin, V98:28--293.

Merrits, DJ, 1987, Geomorphic response to late Quaternary tectonism: Coastal northern California, Mendocino triple junction region, PhD Dissertation, The University of Arizona, 190 p.

Nemes, A., W.J. Rawls, and Ya. A. Pachepsky. 2006. Use of the non-parametric nearest neighbor pproach to estimate soil hydraulic properties. Soil Science Society of America Journal 70: 327-336.Peacock, C.R., 1994, Soil Evolution on a Terrace Chronosequence in the Wind River Basin, Wyoming, Master's Thesis, Colorado State University, 140 p.

Noy-Meir, I., 1973, Desert ecosystems: Environment and Producers, Annual Review of Ecology and Systematics, Volume 4: 25-51.

Peltzer DA, DA Wardle, VJ Allison, WT Baisden, RD Bardgett, OA Chadwick, LM Condron, RL Parfitt, S Porder, SJ Richardson, BL Turner, PM Vitousek, J Walkker, LR Walker, 2010, Understanding ecosystem retrogression, Ecological Monographs 80(4) 509-529.

Phillips, J. D., 1992, Progressive and Regressive pedogenesis and complex soil evolution, Quaternary Research 40, 169-176.

Pickett, S.T.A., 1989, Space-for-Time Substitution as an Alternative to Long-Term Studies, in Long-term Studies in Ecology (ed. G.E.Likens), pp. 110–135. Springer, New York.

Pinney, C, J Aniku, R Burke, J Harden, M Singer, J Munster, 2002, Soil chemistry and mineralogy of the Santa Cruz Coastal Terraces, USGS Open-File Report, 02-277, 20 p.

Reheis MC, 1987a, Gypsic Soils on the Kane Alluvial Fans, Big Horn County, Wyoming, USGS Bulletin 1590-C, 39 p.

Reheis MC, 1987b, Soils in Granitic Alluvium in Humid and Semiarid climates along Rock Creek, Carbon County, Montana, USGS Bulletin 1590-D, 71 p.

Selmants, P.C., S.C. Hart, 2008, Substrate age and tree islands influence carbon and nitrogen dynamics across a retrogressive semiarid chronosequence, Global Biogeochemical Cycles, 22.

Soil Survey Staff, 1997, National Characterization Data, Soil Survey Laboratory, National Soil Survey Center, and Natural Resources Conservation Service, Lincoln, NE.

Stevens, P. R., and T. W. Walker. "The chronosequence concept and soil formation." Quarterly Review of Biology 45 (1970): 333-350.

Taylor, EM, 1980, Impact of Time and Climate on Quaternary Soils in the Yucca Mountain Area of the Nevada Test Site, Master's Thesis, University of California Berkley, 217 p.

Thompson, CH, 1981, Podzol chronosequences on coastal sand dunes of eastern Australia, Nature 291:59–61.

Vitousek, P.M., R.W. Howarth, 1991, Nitrogen limitation on land and in the sea: How can it occur?, Biogeochemistry 13: 87-115.

Vitousek, P.M., O.A. Chadwick, T.E. Crews, J.H. Fownes, D.M. Hendricks, D. Herbert, 1997, Soil and ecosystem development across the Hawaiian Islands, GSA Today, 7:1-8.

Vitousek, P. M. 2004. Nutrient cycling and limitation. Hawaii as a model system. Princeton University Press. Princeton, New Jersey, USA.

Walker, T W, J K. Syers, 1976, The fate of phosphorus during pedogenesis, Geoderma 15:1–19.

Wardel, D.A., R.D. Bardgett, J.N. Klironomos, H. Setala, W.H. van der Putten, D.H.Wall, 2004, Ecological linkages between aboveground and belowground biota, Science 304:5677 p 1629-1633.

CHAPTER 4

VULNERABILITY OF LANDCAPSES TO CLIMATE CHANGE: ASSESSING A 118-YEAR CHRONOLOGY OF SOIL MOISTURE DYNAMICS IN THE GREAT PLAINS OF NORTH AMERICA, USA[3]

4.1 Summary

The meteorologic history of the Great Plains of North America includes substantial spatial and temporal variability in the magnitude and intensity of precipitation inputs. Sustained drought coupled with significant shifts in land-use patterns as happened during the 19th century has fundamentally transformed large portions of the Great Plains. Climate change scenarios suggest the Great Plains will be warmer and precipitation will become more variable with more frequent climate extremes such as more frequent drought (IPCC 2014). To fully evaluate ecosystem responses to these forecasts there is a need to quantify how soils may either dampen or amplify climate changes. The hydrologic functioning of soils is dependent on intrinsic edaphic properties and the variability imparted by landscapes. We used a soil pedohydrologic template to evaluate soil climate vulnerability and gain a better understanding of the response of soil ecosystems to global change drivers (e.g., climate). This research builds on a functional soil phase model and incorporates short-term extreme climate perturbations and management of soil resources into a soil hydrologic framework. We have quantified the historic range of soil moisture and hydrologic conditioning of soils to climatic variables in the Great Plains and utilized simulation models to project changes in soil moisture that are the result of changes in radiative forcing for the years

[3] This manuscript is being prepared for publication-journal yet to be determined: with co-authors P.H. Martin, A.K. Knapp, and E.F. Kelly.

2080-2099. This analysis serves as a baseline for addressing the complexity of soil moisture responses to global climate change.

4.2 Introduction

The response of soil moisture to changing climate in the North American Great Plains (30°-50°N, 105°-95°W) is of great interest given the history of severe drought during the 1930s, 1950s, and over the last decade. The extreme drought conditions occurring in many portions of the Great Plains in the past decade have focused research on the consequence of more frequent summer droughts and increases in temperature within the domain. This area shows signs of becoming increasingly vulnerable to drought due to the increase in the cultivation of marginal lands and the depletion of groundwater from the Ogallala Aquifer (Little 2009; Steward et al 2013). Climate models vary regarding projections in annual precipitation amounts (IPCC 2014; Zhang et al 2007), but there is a general consensus that the hydrological cycle is intensifying with increasing extreme events forecasts along with large inter-annual variation of precipitation (e.g., wetter-wet and dryer-dry years; IPPC 2014).

Soil moisture plays a key role in a number of important land-surface processes – affecting weather, vegetation, and global biogeochemical cycles. Insufficient soil water is considered the factor most limiting to plant growth, with soil controlling ecosystem structure and function (Noy-Meir 1973). Quantifying the effects of drought intensity and frequency is therefore important for sustaining ecosystem function and mitigating the effects of climate change. Currently, drought severity is quantified using the Palmer drought severity index (PDSI; Palmer 1965) or the Standardized Precipitation Index (SPI; McKee et al. 1993). However, when these indices are used to evaluate historic droughts, they lack the integration of soil edaphic properties to assess the variability of droughts at multiple temporal and spatial scales.

Based on extensive studies of soil development in the western portions of the Great Plains the capacity of soil to store and supply water for biological use can be characterized by three broad phases (functional phases) of soil development. *Phase I*, the *Aggrading or Building Stages* in which soil development begins with a new substrate deposited by Alluvial and Aeolian processes or exposed sedimentary rocks. These soils are generally weakly developed and although mineral transformations have occurred the soil is genetically simple; for example, there soils have little horizon differentiation with surface and subsurface materials being pedogenically and hydrologically similar. The total potential of biologically available water or the amount of water available to plants (soil water held in soil tension between field capacity and permanent wilting point) is relatively low and is closely related to the edaphic properties of soil parent material. Soils that would be in this stage include ustipsamments, toriorthents, haplocambids, or haplustepts—mostly depending on parent material and climate.

The *Phase II*, the *Intermediate or Equilibrium Stage,* is found in soils where the formation and clay and $CaCO_3$ become dominant features in the soil profiles and there is significant pedological and hydrological differentiation between surface and subsurface horizons. Clay particles that form as secondary minerals within the soil also retain relatively mobile ions such as Ca, Mg, and K by ion exchange. In this stage biologically available water (BAW) reaches its maximum capacity due to high soil porosity, low bulk density, greater soil organic carbon (SOC) content, and improved soil structure. Soils found in this stage across the Great Plains include argiustols, calciustols, haplocalcids, or calciusterts.

Phase III, the *Degrading or Declining Stage* occurs when the soils have been transformed to the point that they experience losses of clay, buffering capacity, and/or SOC (relative to soils in the intermediate stages of development) due to erosion or other soil forming processes. Pedological and hydrological differentiation is at a maximum and the system's capacity to store water is likely diminished due to reductions in SOC, clay content, and soil structure. Classification of these soils is quite variable due to the multiple conditions driving decline or degradation in soil properties, but some soils in this phase could include haplustalfs, haplargids, or haplustolls. Understanding the distribution and the role these three phases of soil development play in constraining or amplifying ecological responses to future climate changes is a key challenge for the global change researchers and research programs.

The basis of the soil-age based functional phase model was in part based upon biogeochemical functioning of the soil landscapes (Chadwick et al 1999; Vitousek 2004). The pedohydrologic template developed in chapters 2 and 3 of this dissertation were thus realized through conditioning by the long-term weathering of soils over geologic time. Since this model is primarily based on the biophysical soil properties, a soil similar pedo-hydrologic template could be adapted to explain soil degradation from the effects of land use and climate change.

Our research incorporates a deterministic soil water model into a geospatial dataset that allows us to compare patterns of soil moisture regimes within the contemporary record (1895-2012) and utilize scenarios from projections of future climate for the Great Plains (IPCC 2014). The goal of this research is to (1) categorize the variability and frequency of soil moisture regimes within the historical record, (2) identify regions of the Great Plains that are more vulnerable/susceptible to climate change based on an integration of climatic and edaphic properties, and (3) through simulation, forecast future soil climate from projected climate change scenarios across the Great Plains.

4.3 Data and Methods

Soil Hydrologic Model: The Newhall Simulation Model (NSM) is a deterministic soil water balance model designed to integrate monthly atmospheric climate data into information relevant to soil classification categories by simulating soil moisture and temperature data on a daily basis (Van Wambeke 1982, 1986; Smith 1986; Newhall and Berdanier 1996; Jeutong et al. 2000; Yamoah et al. 2003). The NSM can be compared to similar process models, but it retains features uniquely suited to taxonomic classification of soil climate which is used in local, regional, and national resource inventory programs for displaying spatial distribution of soil properties. While other models generate inferences of soil moisture and temperature parameters from climate records—such as the field scale models EPIC and CENTURY (Costantini et al. 2002; Williams et al. 1989)—the NSM couples water balance calculations more directly with available water holding capacity of soils and gives output of predicted "*soil climatic regimes*" or "*soil moisture regimes*" used for taxonomic classification. Also, because the Newhall model represents a longer time window (growing season of 6 to 9 months) then other drought indexes, NSM can provide an improved view of historical drought events during the growing season. Soil moisture regimes were previously proposed as a drought risk indicator as part of a drought decision support system (Waltman et al. 2003). Their work used simulated soil climate to provide historic context supporting drought interpretations for vulnerability mapping and mitigation. They believed when NSM was mapped at multiple scales moisture regimes can be used to identify ecological regions with higher probabilities of drought events. The NSM generates a mesoscale approximation of soil climate that is applicable to soil survey and taxonomic classification (Smith 1986). It assumes all precipitation excess exits the soil as runoff or as deep percolation, thus resulting soil moisture estimates are only valid for well-drained soils associated with relatively level landscapes. NSM lacks a runoff/ponding subroutine and functions on a calendar year rather than hydrological year with no carryover from the previous year, and NSM does not account for snowmelt and also lacks a mechanism for accounting for antecedent moisture conditions. In spite of these limitations, it is widely believed that in most cases the NSM provides a reasonable approximation of soil moisture (number of days moist, days dry) and temperature (number of days <5°C to >8°C) at a monthly interval (Waltman et al. 2003).

Climate Model Simulations: We modeled annual soil moisture status for the Great Plains of North America based on monthly climate data from 1895 through 2012. Although originally developed to calculate soil moisture regimes for classifying soils from climatic data (Newhall and Berdanier 1996), we used NSM to assess annual and seasonal water balance patterns (e.g. soil moisture content) (Zobeck and Daugherty 1982; SSS 2010). Monthly calendars of soil moisture from NSM are based on potential evapotranspiration calculations using the Thornthwaite equation (Thornthwaite 1948). An updated NSM, java version 1.6.0 (jNSM) added soil temperature calculations that model thermal lag behavior, along with annual and summer water balances (USDA-NRCS 2012; Waltman and Others 2011). Winzeler and Others (2012, 2013) first adapted the model into a spatially explicit output into a raster based product, termed the grid element Newhall simulation model (GEN). Geographic models incorporating the jNSM include multiple inputs of monthly precipitation means, monthly air temperature means, root-zone available water holding capacity (AWHC), elevation, geographic coordinates, and the mean soil-air temperature offset. Spatially explicit datasets were created of the soil moisture budget incorporating the GEN simulation model methodology (Winzeler et al. 2013) where each cell of the monthly climate grids provided a reference area in which data from all the input GIS layers was queried. The dataset for each grid cell element was then populated with values for each of the inputs, including: 12 monthly temperature values, 12 montly precipitation values, AWHC, latitude, longitude, and elevation. The Newhall simulation was then run individually for each grid cell using the default 2.5° C mean annual soil-to-air offset at 50 cm below the surface. Model outputs were then aggregated and classed for thematic maps using the jNSM associated software *xml2csv-version* 1.2.0 (USDA-NRCS 2012) including potential evapotranspiration, annual and seasonal water budget, soil moisture regime, soil temperature regimes, and biological window maps.

Coupling of Climate and Soil Data: Monthly raster temperature and precipitation datasets representing one month of one year used in conjunction with the jNSM were obtained from the Parameter Regression on Independent Slopes Model (PRISM) climate data. Gridded PRISM data of monthly precipitation, monthly maximum temperature, and monthly minimum temperature were acquired for the entire period of record from 1895 to 2012 (PRISM 2012). Maximum and minimum temperature grids for each month were averaged to acquire mean monthly temperature. Root-zone available water capacity (AWC) was derived from the continuous United States (CONUS) USDA-NRCS digital general soil map of the (STATSGO2) soil database (Miller and White 1998) and converted to 2.5-arcmin (4 km) resolution, to match the monthly precipitation and air temperature gridded monthly data. Projections of soil climate change on the Great Plains were based on the NCAR Community Climate System Model (CCSM4) using CMIP5 protocol from the fifth assessment of climate change. Downscaled climate datasets were obtained from available from CGIAR Research Program on Climate Change (Ramirez and Jarvis 2008). Soil moisture simulations of downscaled CCSM4 used the representative concentration pathways scenarios RCP2.6, RCP4.5, RCP6.0, and RCP8.5 for the 2080-2099 year range. In general, RCP’s represent increasing climatic forcing of anthropogenic greenhouse gases, with RCP2.6 yielding an increase of about 1°C and RCP8.5 representing an increase in global temperatures of approximately 5°C (see IPCC 2014 for more detail on the different concentration pathways).

4.4 Results and Discussion

Regional Water Balance: Water balance from the growing season and calendar year were averaged over the western portion of the Great Plains for the period 1895-2012 (figure 4-1) and the three-year running average (red line). Statistical comparisons between annual water balance and growing season water balance using spearman rank correlation coefficient yielded an insignificant relationship (r_s = 0.1224) and the two data sets produced poor linear correlation (p=0.281, R^2= 0.009). Poor correlation between the growing seasonal water balance and annual water balance suggests that single large precipitation events outside the growing season are enough and sufficient to alter the annual soil hydrologic regime. From the annual water balance data, five years were more than two standard deviations (SD) lower than the western Great Plains mean (1910, 1934, 1954, 1956, and 2012), and only four years from the seasonal water balance were more than two SD below the mean (1950, 1954, 1957, and 1976). Regarding the duration of droughts, our data indicates that there were five periods where the annual water balance was lower than the mean for four or more consecutive years (1901-1904, 1907-1911, 1933-1940, 1952-1956, and 2000-2003). Similarly, there were six periods where the seasonal water balance remained lower than the mean for four or more consecutive years (1906-1910, 1925-1930, 1933-1938, 1950-1957, 1967-1970, and 1974-1977).

Regional Patterns of Soil Moisture: Soil moisture regimes were calculated for the whole Great Plains from the historical data (1895-2012). Based upon the aforementioned analysis of annual water balance, specific years were examined to identify regional geographic trends calculated from soil properties and climate extremes. Three years (1934, 1956, and 2012) were found to have the greatest extent of aridic soil moisture regimes (with soil moisture regime sub-modifiers as weak, typic, and extreme) across the western margins of the Great Plains (figure 4-2). Within these years, soils in western Kansas, Oklahoma, and Texas showed extreme drought conditions seen by the typic-aridic and extreme-aridic soil moisture classes. Soils in the region were highly eroded or degraded during the period of time known as the “Dust Bowl” era.

Climatic research of the drought occurring during 2012 has shown that conditions rivaled the combined rainfall deficits and high temperatures observed during 1930’s (Hoerling et al 2014). Our analysis shows similar soil water conditions geographically between the two periods (figure 4-2); however, due to large erosion and loss of fertility that occurred as a result from the farming practices and periodic drought from the 1930s to 1960s, these soil landscapes are fundamentally different today (Lal et al. 2007). This change in soil condition mimics the functional soil phases (described in Chapter 3) seen as a result of soil development, where instead of long-term weathering conditioning the soil into a degraded or declining phase, it is instead the combined effect of short-term climate and land management practices that shifted the soil from *Phase II* into *Phase III*.

Comparable to intense droughts, some years saw relatively large precipitation during the period of study. For example, our results show that during 1941 precipitation patterns in portions of the Chihuahuan desert had rain patterns more familiar to Iowa then New Mexico. In part due to the vegetation types more common to the arid southernmost region of the Great Plains, much of this precipitation did not infiltrate the soil, but contributed to over land flow, increasing erosion potential, and surging into the river flow regime (Yuan et al. 2007). Thus, certain soil landscapes won’t always be able to adjust to the wettest years, suggesting that wet years could be just as harmful from a soil hydrologic perspective as some of the driest years.

Forecasting Soil Moisture Status*:* Simulations of soil moisture using the NCAR CCSM4 projections of the four representative concentration pathways from the IPCC5 report are reported in figure 4-3. Although the different representative concentration pathways signify increasing levels anthropogenic radiative forcing of greenhouse gasses, in general, RCP2.6 signifies an average global temperature increase of 0.2°C to 1.6°C, RCP4.5 a 1.0°C to 2.5°C, RCP6.0 a 1.4°C to 3.1°C, and RCP 8.5 a 2.6°C to 4.8°C increase. Our results of simulated soil climate show that with each progressive increase in temperatures, much of the Great Plains will shift toward a more thermic or hyperthermic soil temperature classification, with the mean annual soil temperature between 15°C and 22°C (thermic) or greater than 22°C (hyperthermic) at a depth of 50 cm. The CCSM4 model is less certain regarding the expected changes in precipitation with the various concentration pathways. Soil moisture regimes show a general trend toward dryer conditions, but patterns in moisture regimes are not as pronounced as seen in the temperature simulations. As shown in figure 4-3, increasing radiative forcing generally shifts drier soil moisture regimes eastward; however, aridic soil moisture regimes do not generally increase in total area across the Great Plains. Integration of forecasts of the annual variability for the 2081 to 2099 period into these climatic averages could provide further insight into the expected year-to-year drought conditions.

Vulnerability of Landscapes: Considering the coupling of BAW to changes in the short-term climate, our results show that some landscapes with higher water holding capacities are buffered from drought conditions. For example, our model shows that areas within our study along the Canadian and Red River in the panhandle of Texas maintain wetter soil moisture regimes when most of the regions around those rivers are simulated to be classified as much drier moisture regimes; however, because the NSM does not carryover moisture status from year-to-year, these results may not reflect the sustained drought conditions experienced in the region. We suggest that soils with higher BAW which are conditioned with higher soil fertility, improved land management, and drought adaptation strategies have the potential to dampen the effect of global change. Similarly, soils in portions of southwest Kansas, western Oklahoma, and far northwest Texas which were subject to intense wind erosion during the 1930s are classified to have extreme-aridic soil moisture conditions during the driest years from the record (figure 4-2). Through this sort of soil degradation, soils have thus lost their inherent fertility through either mass erosion of the soil profile.

The functional soil phase model described in chapters 2 and 3 explains the evolution of BAW in soil ecosystems conditioned by soil age over the full life of the soil: from the initial stages of soil formation to old and highly weathered soil landforms (absent of rejuvenating disturbances). We propose that these functional phases could also be conditioned by short term climate perturbations and management decisions. An extreme example of soil transitioning from a phase II to a phase III soil pedo-hydrologic would be converting natural prairie rangeland into cultivated cropland.

4.5 Conclusions and Further Analysis

The framework presented in this chapter to study regional soil moisture patterns demonstrates potential of using soil taxonomic classes of soil moisture regime to gain perspectives of historic droughts. Because the NSM simulates the total number of calendar day in which soils are either wet or dry, it couples water balance calculations more directly with available water-holding capacity. Also, the Newhall model represents a longer time window (growing season 6 to 9 months) than other drought indexes, such as PDSI, allowing it to describe different parameters of extreme drought events, although other indexes such as the Standardized Precipitation-Evapotranspiration Index (SPEI), can be any length of time. Using the dataset and methodology presented in this chapter, future research is needed through geospatial modeling to identify specific landscapes that are vulnerable to climate change – specifically due to changes in soil moisture regimes or changes in precipitation.

Future work should specifically address the properties of root-zone available water capacity or the BAW in the soil. Combined with analysis of climatic forcing, BAW should be used to ascertain if landforms across the Great Plains of North America are particularly vulnerable to global change. The working hypothesis is that soil landscapes with >200mm BAW will not be as constrained to climatic shifts, soil landscapes with 150mm to 200mm BAW would show marginal responses to changes in climate, and soils <150mm will reflect the greatest response to climate with those soil landscapes displaying edaphic droughts with even small decreases in precipitation or increases in temperatures.

The Conclusions from this chapter are:

(1) Poor correlation between the seasonal water balance and annual water balance suggests that single, large precipitation events outside the growing season are sufficient enough to alter the annual soil hydrologic regime,

(2) Analysis of trends in annual and seasonal soil water balance have delineated the major drought periods from the historic record in the western portion of the Great Plains, providing years to further analyze patterns of soil moisture dynamics,

(3) Regional modeling of taxonomic classes of soil moisture regimes on an annual basis have helped provide perspectives of historic droughts across the Great Plains,

(4) Simulated soil moisture projections of soil climate for 2080-2099 suggest that soil moisture classification yields a general drying of the Great Plains, however projections do not show a general increase in total area of aridic soil classification.

(5) Soil temperature classification shows a consistent warming of the Great Plains toward thermic and hyperthermic soil temperature classification.

The soil classes described in this chapter represent a soil-hydrologic template in which landforms are stratified to understand soil moisture's response to global change. This work proposes three functional phases that can be determined by soil age (Chapter 3), climatic change or variability, human induced disturbance, geologic substrate, and /or landscape position. This research builds on the functional soil phase model described in chapters 2 and 3 to incorporate short-term extreme climate perturbations and management of soil resources into a soil hydrologic template.

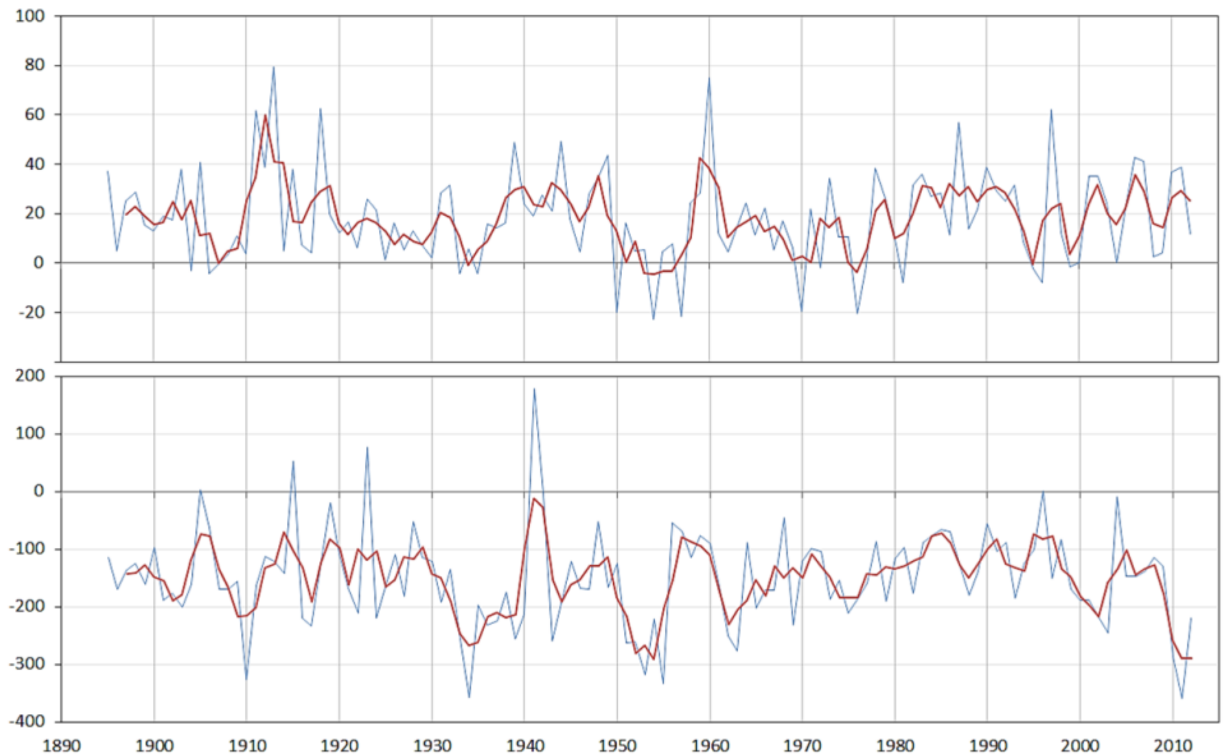

Figure 4-1: Seasonal (top) and annual (bottom) water balance derived from the Newhall simulation model for the driest portion of the Great Plains biome (blue line). Red line is a smoothed 3 year average.

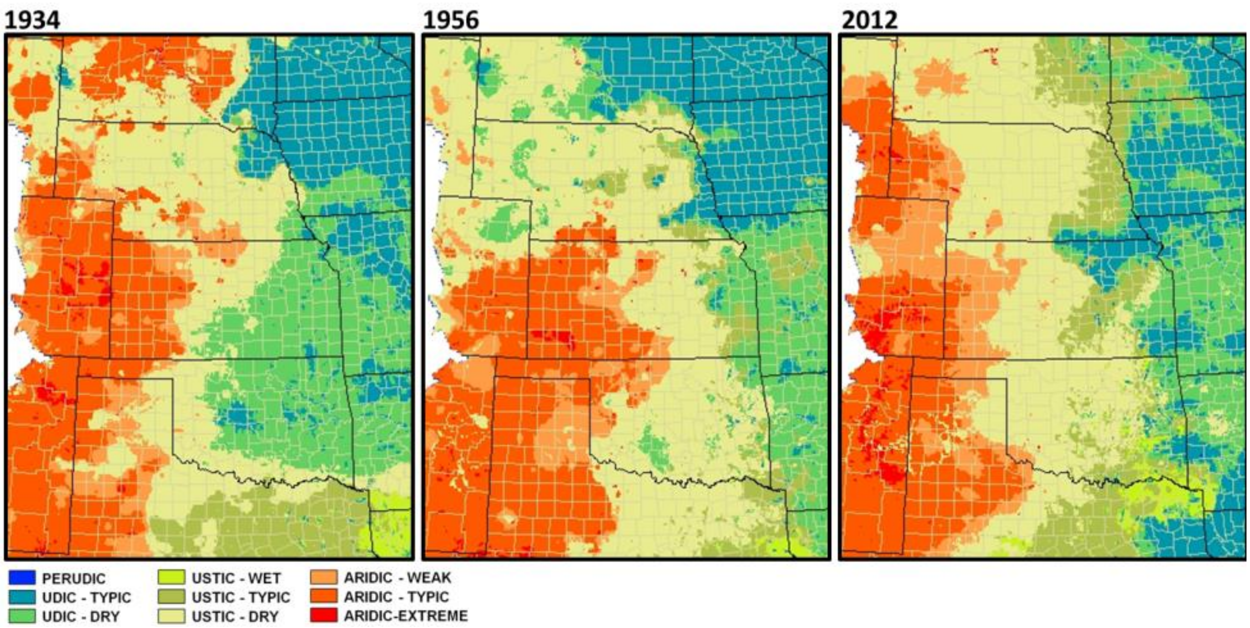

Figure 4-2: The most challenging years to farm or ranch in western Great Plains from the historic record (1895-2013) where drought conditions were at their greatest extent. Geospatial data derived from Newhall simulation runs.

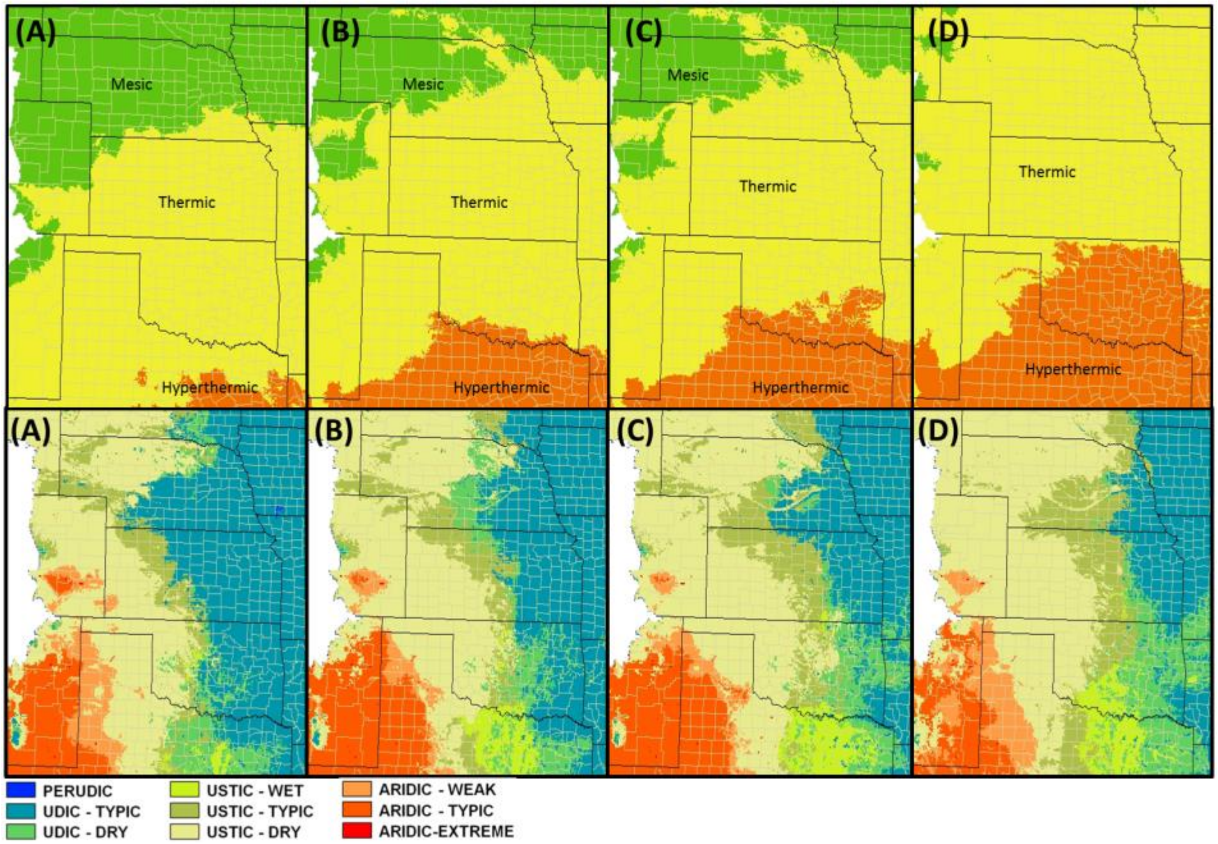

Figure 4-3: Soil climate simulation using the NCAR climate model 2080-2099 for soil temperature regimes (top) and soil moisture regimes (bottom). (A) Uses the RCP2.6 scenario, (B) uses the RCP4.5 scenario, (C) uses the RCP6.0 scenario, and (D) uses the RCP8.5 scenario.

4: REFERENCES

Baldwin M, Kellogg CE, Thorp J. 1938. Soil Classification. In Soils and Men: Yearbook of Agriculture 1938. United States Department of Agriculture. United States Government Printing Office, Washington, D.C., USA. Pg 979-992.

Chadwick, O.A. L.A. Derry, P.M. Vitousek, B.J. Hubert, L.O. Hedin, Changing sources of nutrients during four million years of ecosystem development, Nature 397: 491-497.

Costantini, E.A.C., F. Castelli, S. Raimondi, and P. Lorenzoni. 2002. Assessing soil moisture regimes with traditional and new methods. Soil Sci. Soc. Am. J. 66:1889–1896.

Danels D, Johnson G. 2001. Development of a New Precipitation-Effectiveness Index. ESRI User Conference Proceedings.

Field, C. B., and Coauthors, 2012: Managing the Risks of Extreme Events and Disasters to Advance Climate Change Adaptation. Cambridge University Press, 582 pp.

Hoerling, M, J Eischeid, A Kumar, R Leung, A Mariotti, K Mo, S Schubert, R Seager, 2014,Causes and predictability of the 2012 Great Plains Drought, Bulletin American Meteorological Society, 95:269-282.

IPCC, 2014: Summary for Policymakers. In: Climate Change 2014: Impacts, Adaptation, and Vulnerability. Part A: Global and Sectoral Aspects. Contribution of Working Group II to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change [Field, C.B., et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp. 1-32.

Jeutong, F., K.M. Eskridge, W.J. Waltman, and O.S. Smith. 2000. Comparison of bioclimatic indices for prediction of maize yields. Crop Sci. 40:1612–1617.

Lal, R., D.C. Reicosky, J.D. Hanson, 2007, Evolution of the plow over 10,000 years and the rationale for no-till farming, Soil and Tillage Research, 93:1, pg 1-12.

Little, J. 2009. "The Ogallala Aquifer: Saving a Vital U.S. Water Source." Scientific American, March.

Machette, MN. 1985. Calcic soils of the southwestern. United States Special Paper 203 Pg 1-21.

Marbut, CF. 1935. Soils of the United States in "Atlas of American Agriculture" (E. 0. Baker, Ed.), part 3, Advance Sheets No. 8. USDA.

McKee, T. B., N. J. Doesken, and J. Kleist, 1993: The relationship of drought frequency and duration of time scales. Eighth Conference on Applied Climatology, American Meteorological Society, Jan17-23, 1993, Anaheim CA, pp.179-186.

Miller DA, White RA. 1998. A Conterminous United States Multi-Layer Soil Characteristics Data Set for Regional Climate and Hydrology Modeling. Earth Interactions, 2. Available from: http://EarthInteractions.org.

Newhall F, Berdanier CR. 1996. Calculation of soil moisture regimes from the climatic record. Soil Survey Invest. Rep. 46. Natural Resources Conservation Services, Lincoln, NE.

Nichols JD, Stone JF. 1970. Evaluation of Soil Moisture Measurements in Oklahoma as Soil Characteristics for Classification. Soil Sci Soc Am Journal 34: 4: pg638-641.

Noy-Meir, I., 1973, Desert ecosystems: Environment and Producers, Annual Review of Ecology and Systematics, Volume 4: 25-51.

Palmer, W.C., 1965: Meteorological drought. Research Paper No. 45. U.S. Weather Bureau. [NOAA Library and Information Services Division, Washington, D.C. 20852]

Powell, JW. 1879. Report on the lands of the arid region of the United States. United States Government Printing Office.

PRISM Climate Group. 2012. Gridded climate data for the contiguous USA. [cited 2012 Dec 15]. Available from: http://prism.oregonstate.edu.

Ramirez, J.; Jarvis, A. 2008. High Resolution Statistically Downscaled Future Climate Surfaces. International Center for Tropical Agriculture (CIAT); CGIAR Research Program on Climate Change, Agriculture and Food Security (CCAFS). Cali, Colombia.

Smith, GD. 1986. The Guy Smith Interview: Rational for Concepts in Soil Taxonomy. SMSS Technical Monograph # 11. Soil Conservation Service, Washington, DC.

[SSS] Soil Survey Staff. 2010. Keys to Soil Taxonomy, 11th ed. USDA-Natural Resources Conservation Service, Washington, DC.

Steward, DR, PJ Bruss, U Xiaoying, SA Staggenborg, SM Welch, MD Apley, 2013, Tapping unstainable groundwater stores for agricultural production in the High Plain Aquifer of Kansas, projections to 2110.

Smith, GD. 1986. The Guy Smith Interview: Rational for Concepts in Soil Taxonomy. SMSS Technical Monograph # 11. Soil Conservation Service, Washington, DC.

Thornthwaite CW. 1948, An approach towards a rational classification of climate, Geographical Review. 38:55-94.

USDA-NRCS. 2011. JAVA newhall simulation model—Update to a traditional soil climate simulation model, Abstract. Soil Sci Soc Am Int. Annual Meeting. San Antonio, TX. 18 Oct.

USDA-NRCS. 2012. jNSM 1.6.0 - Java Newhall Simulation Model. [cited 2013 Feb 1]. Available from: http://soils.usda.gov/technical/classification/jNSM/.

Van Wambeke, A, 1982, Calculated moisture and temperature regimes of Africa. Soil Management Support Services. SCS, USDA. U.S. Gov. Print. Office, Washington, DC.

Van Wambeke, A. 1986, Newhall Simulation Model, a BASIC program for the IBM PC (DOS 2.0 or later). Cornell University, Ithaca, NY.

Vitousek, P. M. 2004. Nutrient cycling and limitation. Hawaii as a model system. Princeton University Press. Princeton, New Jersey, USA.

Waltman SW, Miller DA, Bills B, Waltman WJ. 2011. JAVA Newhall simulation model—Update to a traditional soil climate simulation model. Presented at the Abstract Soil Sci. Soc. Am. Int. Annual Meeting, San Antonio, TX, 18 Oct.

Waltman, W.J., Goddard, S., Reichenbach, S.E., Svoboda, M.D., Hayes, M.J., Peake, J.S., 2003. Patterns and trends of soil climate regimes and drought events in the northern great plains. In: Proceedings of the Applied Geography Conference, Colorado Springs, Colorado.

Williams, J.R., C.A. Jones, J.R. Kiniry, and D.A. Spanel. 1989. The EPIC crop growth model. Trans. ASAE 32:497–511

Winzeler HE, Owens PR, Waltman SW, Libohova Z, Waltman WJ. 2012. Soil Climate indicators from the Geographically Explicit Newhall Simulation Model (GEN) as potential environmental covariates in digital soil mapping applications. In: Minasny, Malone, McBatney, editors. Digital Soil Assessments and Beyond: Proceedings of the 5th Global Workshop on Digital Soil Mapping 2012, p 167-71.

Winzeler HE, Owens PR, Waltman SW, Waltman WJ, Libohova Z, Beaudette D. 2013. A Methodology for Examining Changes in Soil Climate Geography through time: U.S. Soil Moisure Regimes for the Period of 1971-2000. Soil Sci. Soc. Am. J. 77: 213-25.

Yamoah, C.F., A. Bationo, T.J. Wyatt, B. Shapiro, and S. Koala. 2003. Simulated weather variables effects on millet fertilized with phosphate rock sand in the Sahel. Nutr. Cycling Agroecosystems 67:167–176.

Yuan, F, S. Miyamoto, S. Anand, 2007, Changes in major element hydrochemistry of the Pecos River in the American southwest since 1935, Applied Geochemistry, 22:1798-1813.

Zobeck T M, Daugherty LA. 1982. Calculating the depths of soil moisture control sections from laboratory data. Soil Sci. Soc. Am. J. 46:792-5.

CHAPTER 5

A LONG-TERM PERSPECTIVE OF THE DRY BOUNDARY FOR THE GREAT PLAINS OF NORTH AMERICA[4]

5.1 Summary

The boundary between the humid eastern and the arid western United States is of great economic interest and historic intrigue. Areas to the east of this boundary have historically enjoyed favorable rainfall, fertile soil, and reliable surface water permitting conventionalal agriculture over surprising large areas. The expansion of population and agriculture during the nineteenth century across the Great Plains of North America tested the extent to which agriculture could be successful in areas where year-to-year rainfall was unreliable. In this chapter, we have quantified the historic range of soil moisture and hydrologic conditioning to climatic variables in the Great Plains by showing regions that historically demonstrate reliable precipitation, identifying a boundary for the arid regions of the United States. We compare our results to historic delineations of the arid west, including the those from past director of the U.S. Geological Survey, John Wesley Powell and of the Soil Survey Division, Curtis Marbut. This analysis serves as a baseline in understanding the fate and distribution of soil moisture's response to droughts and global climate change.

[4] This manuscript is being prepared for publication to SSSA Horizons with co-authors RO Sleezer, and EF Kelly.

5.2 Introduction

In his 1878 report on the arid lands of the Western United States, John Wesley Powell described that midway across the Great Plains there begins a region that is “so arid that agriculture is not successful without irrigation” (Powell 1879). He deemed that this boundary between the humid region to the east and the arid region to the west occurs at 20 inches (50.8 cm) of precipitation annually, assuming that rainfall is evenly distributed throughout the year. This boundary, falling approximately along the 100th Meridian, marks the normal reach of moist air from the Gulf of Mexico. Settlement of the Great Plains soon began after the American Civil War with agricultural lands encroaching on the 100th Meridian beginning in the 1870s; however, since Powell’s report on the western lands, agroecosystems have expanded considerably beyond the 20 inch isohyetal line of seasonal precipitation, resulting in sometimes drastic effects on the historic native landscape. This time of increased exploitation of the pedosphere by human activity marks a turning point in our history where agricultural civilization has had a significant impact on the Earth’s ecosystems. It is thusly important to recognize the complex soil-human interactions within the critical zone during this shift.

Soil scientists have long recognized that the Great Plains pose an exceptional natural experiment with climate as a soil-forming factor. While the climate expresses great variability across the region, the variations of other soil formation factors are relatively modest (Jenny 1941), mainly due to the soil’s postglacial age (younger than 14ka), uniform loess parent material, rolling to flat topography, and resilient grassy plant communities. With this natural arrangement, the Great Plains can generally be interpreted with increasing temperatures from north-to-south and increasing moisture from west-to-east allowing us to apply a coupled hydrologic and climatic model to define characteristic regions of soil moisture from the contemporary record.

One of the first attempts to classify soils in the United State divided them at the highest level into pedocals and pedalfers (Marbut 1935). The Pedocal-Pedalfer soil boundary line is a zero point where the mean annual precipitation and evapotranspiration are equal (Jenny 1941). Pedocal soils were distinguished by accumulation of calcium and magnesium in arid or semi-arid regions in the form of pedogenic carbonates, and pedalfer soils were identified by the absence of carbonates and were enriched in aluminum and iron sesquioxides in humid regions. The now antiquated terms of pedocal and pedalfer are still used in quaternary geology and soil science to distinguish the sub-humid climatic zone. The boundary between the semi-arid and humid climate regime fundamentally still exists in soil taxonomy at the suborder level, and this basic wet-dry categorization evolved into the modern soil moisture regime's conceptual framework (Perudic, Udic, Xeric, Ustic, and Aridic). However, soil moisture classes were originally based on the agricultural usage (Smith 1986). Where Aridic soils are defined as typically too dry to support crops without irrigation, Ustic soils support crops that are drought tolerant, and Udic soils do not require irrigation to successfully grow crops.

Long-term climate imprints pedogenic properties on the soil, allowing scientists to recognize moisture regimes through evaluation of the soil pedon. However, soil moisture classes are typically derived with climatic data to describe the ecological boundaries of crops and classify growing season environments. Deterministic soil water balance models using evapotranspiration formulas such as the Newhall Simulation Model were developed by the United State Department of Agriculture in the 1970s to derive soil moisture regimes regionally. Using these same soil climate models on an annual time-step basis with monthly meteorological data soil moisture models, the same soil moisture regimes can be used as proxies to assess intensity of drought events, climatic shifts, and inherent climate variability.

In this chapter, we use a soil water balance simulation model to determine the variation of the boundary for the “arid-” Western United States. Specifically, we have created geospatial datasets of (1) the annual Pedocal-Pedalfer boundary determined where the annual precipitation equals annual evapotranspiration, and (2) the Udic-Ustic soil moisture boundary simulated from monthly meteorologic data using a pedologic soil water balance model. We then have mapped the long-term average and variation of the boundary from 1895 to 2012 as well as the location of that margin during periods of intense drought on the Great Plains of North America.

5.3 Methods

Simulations for soil hydrologic properties are based on monthly climate data from 1895 to 2012. The cross-continental gradient of semi-arid to humid climates across the historical prairie, steppe, and grassland regions were mapped to understand climatic extremes and regional patterns of drought. Key soil variables such as soil moisture, soil temperature, and windows of biological activity were analyzed to create a temporal record of soil moisture for the study area. Mapping tasks were performed using ArcGIS 10 software (ESRI 2012). Grid-point population tasks were performed in WGS72 spatial projection, and all aerial estimates were made using Albers equal area projection parameters. Higher resolution raster and vector data were resampled and rasterized to a common target of 2.5 arc minute of geographic degree, or approximately a 4 km resolution.

The Newhall Simulation Model (NSM) was developed by Franklin Newhall and Guy Smith in 1973 to compute soil moisture regimes according to Soil Taxonomy (Newhall and Berdanier 1996). The NSM was used to simulate annual and seasonal water balance patterns for soil moisture in the calculated soil moisture control section and temperature defining taxonomic windows of soil climate regimes. Soil moisture control section concepts are further described by Zobeck and Daugherty (1982) and *The Keys to Soil Taxonomy* (SSS 2010). Monthly calendars of soil moisture from NSM are based on potential evapotranspiration (PET) calculations using the Thornthwaite equation (Thornthwaite 1948).

The updated NSM, java version 1.6.0 (jNSM) added soil temperature calculations that model thermal lag behavior, along with annual and summer water balances (USDA-NRCS 2012; Waltman and Others 2011). Winzeler and others (2012, 2013) first adapted the model runs input from the jNSM into a spatially explicit output based upon each grid cell of raster, termed the grid element Newhall simulation model (GEN). Geographic models incorporating the jNSM include multiple inputs of monthly precipitation means, monthly air temperature means, root-zone available water holding capacity (AWHC), elevation, geographic coordinates, and the mean soil-air temperature offset.

Simulations estimate the cumulative number of days soil moisture is both moist and greater than 5° C, and the highest number of consecutive days moisture is both moist and greater than 8° C. Both of these estimates are defined as "biological windows" of plant and microbial activity. The concept of biological windows serves as useful bioclimatic indicators of inherent soil quality, since it integrates soil moisture and temperature, as well as the window of time available for root and microbial activity. The biological window calculation relates to soil processes such as the mineralization of organic matter, soil carbon storage, herbicide degradation, and nitrification. Seasonal water balance patterns in the calculated soil moisture control section are simulated in three categories for soil water (moist, partly moist and partly dry, and dry) and in three categories for soil temperature (number of days <5° C, 5° to 8° C, and >8° C).

To create spatially explicit maps of the soil moisture budget, we incorporated the grid element Newhall simulation model methodology (GEN) described by Winzeler and Others (2013). Each cell of the monthly climate grids provided a reference area in which data from all the input

GIS layers was queried. The dataset for each grid cell element was then populated with values for each of the inputs, including 12-monthly temperature values, 12-montly precipitation values, AWHC, latitude, longitude, and elevation. The Newhall simulation was then run individually for each grid cell using the default 2.5° C mean annual soil-to-air offset at 50cm above the surface. Model outputs were then aggregated and classed for thematic maps using the jNSM associated software xml2csv-version 1.2.0 (USDA-NRCS 2012) including potential evapotranspiration, annual and seasonal water budget, soil moisture regime, soil temperature regimes, and biological window maps.

The monthly raster temperature and precipitation datasets representing one month of one year used in conjunction with the jNSM were the Parameter Regression on Independent Slopes Model (PRISM). Gridded PRISM data of monthly precipitation, monthly maximum temperature, and monthly minimum temperature were downloaded for the entire period of record from 1895 to 2012 (PRISM 2012). Maximum and minimum temperature grids for each month were averaged to acquire mean monthly temperature. Root-zone available water capacity (AWC) was derived from the continuous United States (CONUS) USDA-NRCS digital general soil map of the (STATSGO2) soil database (Miller and White 1998) and converted to 2.5-arcmin (4 km) resolution, to match the monthly precipitation and air temperature gridded monthly data. The PRISM digital elevation terrain model used as the elevation input for all simulations (PRISM 2012).

The annual water balance output from the jNSM is based upon precipitation inputs and calculated potential evapotranspiration. Where net water balance is zero, we have digitized an isohyetal line from north to south across the Great Plains. Similarly, where the NSM predicts the border between Udic and Ustic moisture regimes, we created binomial raster grids (1, 0) where

Ustic classification had a value of one and Udic regimes had a value of zero. Then through averaging those raters using a raster calculator, a distribution of typical years was established. creating the basis for our analysis of the temporal scale of the soil pedocal/soil pedalfer line as it changes from 1895 to 2013. All raster products were projected to Albers equal area projection parameters. Similar analysis was done between the Aridic and Ustic moisture regimes. Total land areas of each output were calculated in ArcGIS.

5.4 Results

Long-term soil moisture's determination of the "arid west" closely mimics the mean climate isohyetal where precipitation equals potential evapotranspiration (figure 5-1). The soil boundary occurs further to the west in areas where there is larger root-zone available water capacities, and the largest discrepancy between these boundaries is across Kansas in which deep loamy soils occur and have some of the greatest soil water holding capacities on the Great Plains. Locations where the soil boundary occurs east of the climate boundary is in regions within the sand hills of Nebraska and across some of the shallow soil regions of Texas. The first half of the record (1895–1953) was found to be slightly drier then the last half (1954–2014) with more departure from the normal occurring in portions in the southern United States with Texas having the greater amount of deviation (figure 5-2a). These records closely mimic the long term average with trend stretching from the northwest to the southeast.

The ten driest and wettest years based upon total area are presented in figure 5-2b. Notably the average of the ten driest years, a more north-south oriented boundary are modeled that follows the sixth principle meridian (97.38° Latitude) from central Oklahoma north to the Canadian border, while the maximum extent of dry soil moisture determination extended into Minnesota, Iowa, Missouri, and Illinois (figure 5-2b). The wettest ten years do not deviate from the long-term averages to the same degree as the dry years north of Oklahoma. This trend does deviate within the Texas Panhandle where the region appears to have higher soil moisture contents during wet years than the region just to the north (i.e., southeast Colorado and southwest Kansas). The large rain-shadow effect is likewise seen to the east of the Black Hills of South Dakota. Looking at patterns during the two most significant droughts for the Great Plains (1931–1939 and 1952–1956) geographic trends (figure 5-2c) mimic the driest 5% line (orange line in figure 5-2b). However, we see a pattern in Montana and North Dakota in which the whole northern region of the Great Plains has higher soil moisture contents during these dry years. Also between these two prolonged drought periods, the central portion of the Great Plains (Kansas and Oklahoma) is affected differently, where although the drought in the 1950s was shorter in duration, it appears to be equally intense. Aggregating all years, the central tendency of the Pedocal-pedalfer soil isohyetal (middle 50th percentile) migrates within a more traditional view of the Western Great Plains (figure 2d). Dry soil moisture extents have an infrequent recurrence, yet occur with greater frequency within common decades: specifically the 1930s and the 1950s. Frequencies of favorable years are more widely spaced with some decades being “wetter” such as the 1910’s, 1940’s, and the 1980’s.

5.5 Discussion

The goal of this study was to quantifiably define the eastern edge of the Great Plains. Comparing our analysis of the Arid West with Powell’s and Marbut’s border we have recognized, as others have (SSS 2010; Winzeler 2013), that areas in the southern extents of the study area tended to be to a greater extent drier and the leaching boundary occurring further to the east. This is explained by a negative summer water balance associated with higher evapotranspiration and lower summer precipitation. A similar result is observed near the Canadian border where the region is modeled to have wetter conditions. Lower temperatures have caused a noticeable positive water balance due to lower annual evapotranspiration values.

Newhall simulations also identified—quite unexpectedly—xeric soil moisture regimes with varying frequency in the northern portion of the study area, soils that pedologists normally would classify as ustic. Xeric moisture regimes are typically characterized by wet winters with soil moisture surpluses and dry summers with large moisture deficits. Normally the xeric soil moisture condition is associated with much warmer Mediterranean climates such as in California and eastern Oregon, yet the Newhall model predicts a xeric condition in the Northern Great Plains as a product of cold winters producing large winter moisture surplus and subsequently large summer deficits when temperatures and evapotranspiration increase. Winzeler and others (2013) conceded that these areas are classified as xeric, however these cold xeric profiles behave differently than the xeric moisture regimes of warmer Mediterranean climates and different biologically than the ustic soils of the western Great Plains— for example deep percolation of soil water will occur before seasonal increases in actual evapotranspiration.

Annual soil moisture patterns on the North American Great Plains are extremely variable both spatially and temporally. Configurations of soil moisture variability in the Great Plains similarly follow the climatological concept of continentality, where the sum of climate features are determined by the influence of large land masses and the remoteness of the land area from the direct impact of oceans. We also found that some soil landscapes have a tendency to be more resilient and some less resilient to drought due to the soil's edaphic properties to store moisture within the profile. Recognizing that stored soil moisture is depleted when actual evapotranspiration is greater than precipitation is not a contemporary observation; however, the spatial representation of this phenomenon shows that some landscapes on the Great Plains tend to be less vulnerable to drought. The exception to this is when drought occurs in multiple consecutive years, such as in the mid-1930's, early 1950's, and early 2010's, when soil moisture storage would not have been replenished for plant availability in the following year.

While trying to define where the arid west begins based upon the contemporary record, we recognized that the extreme temporal and spatial variability that occurs on the Great Plains restricts a precise delineation. Our analysis could lead to multiple peripheral boundaries such as (1) the median boundary in figure 5-1 (blue line) for the contemporary record, (2) the more recent median boundary in figure 5-2a (blue line), (3) the gradational zone seen in figure 5-2d, or (4) even possibly the more arid boundary seen in figure 5-2c. Using the more arid interpretation seen as the yellow line in figure 2b would be plausible since this closely follows the 1990s NRCS Ustic-Udic boundary. The original impetus of defining moisture regimes for soil taxonomy was to capture the conditions that occur "most of the time," thus soil scientists pragmatically defined moisture state conditions from occurrence of 6 out of 10 (Smith 1986).

Our analysis of soil moisture conditions for the contemporary record, regarding the isoedaphic boundary of leaching and non-leaching soils, vary greatly from Marbut's line falling further eastern. Both the current National Soil Survey's and Marbut's 1935 line are defined by soil properties and not climatic data. Occurrence of the pedogenic-based line further east then our climatic variable line suggests that carbonate precipitation is driven either by the current sporadic drought frequency or from a more arid past hydrologic regime. This leads us to believe that the return occurrence of drought within geologic time is sufficiently frequent that the pedo-hydrologic conditions of these soils reflect drought conditions more positively then wetter conditions. When years are averaged from the driest 5th percentile (figure 5-2b), a closer approximation is reached to the established pedogenic leaching boundary.

Comparing average isoedaphic boundaries from the first half of the record (1895–1953) to the last half (1954–2013) we have noted that regions of Oklahoma and Texas are actually ”wetter,” or that the boundary for leaching occurs further west—as much as 300 km west along the 32 parallel north. We believe this moisture bias is due to the lower root-zone water holding capacities of those regions exhibiting shallow soil conditions allowing larger apparent shifts in the boundary then if there was a more uniform water holding capacity. Consequently such a shift was not observed in western Kansas in a region with more uniform parent material of Pleistocene and Holocene loess having higher root-zone water holding capacities. In this case we have identified specific regions and edaphic properties in the North America Great Plains that are most sensitive to global climate change drivers. The climatic drivers that have caused this shift are more complicated and include the complex patterns of atmospheric and oceanic currents and possibly warmer sea temperatures in the Gulf of Mexico producing a higher incidence of precipitation in the far southeast portion of our study area.

Pedo-hydrologic properties of soil that limit root-zone water capacity include sandy soils, shallow soils, and soils with a root restrictive layer. As mentioned above, soil surfaces with lower root-zone available water capacities tend to be most sensitive to shifts in climate. Understanding the pedogenesis of these landscapes with low water capacities will help understand the pedosphere’s response to climate change. We also recognize, although it was not modeled in this project, that tilled soils from agroecosystems are known to decrease water holding capacity by reducing in soil structure, increasing evaporative demand, and altering how water is distributed after precipitation. As improved estimates of past annual climates within the contemporary record become available, land use should be integrated within this model framework.

5.6 Conclusions

We have identified for the contemporary record a generalized western limit in the United States in which agroecosytems would have to be either drought tolerant or irrigated to grow successfully and predictably. Our results support the suggestion that soil landscapes with coarse textural soils and low water holding capacities are most sensitive to changing climate drivers. The work presented here also suggests that the extreme inter-annual variability of climate on the western portions of the North America Great Plains limits reliable forecasting of soil moisture as we note that soil moisture response is highly dependent on the intensity and timing of precipitation. Informed management decisions regarding climate change mitigation for western lands require a clear understanding of soil-hydrologic functions of landscapes, as we conclude that the greatest degree of vulnerability for an ecosystem to global climate change is reflected in the soil's hydrologic regime.

5.6: FIGURES

Figure 5-1: Arid boundaries of the Great Plains. Great Plains of North America within the shaded area and region modeled for this project. Green and Black lines are the Powell line (1878) and the Marbut Line (1935), respectively. for how they defined the boundary between the humid east and the arid west. The red line is the long-term average climatic boundary from the historic record (1895-2013) where precipitation equals evapotranspiration and the blue line is the long-term soil moisture boundary from the same time period.

Figure 5-2: Soil moisture simulations of the 119-year record and the pedocal/pedalfer boundary of: **(a)** differences in average soil moisture conditions from 1895-1953 (green) and 1954-2013 (blue), **(b)** the most arid 6 years (5^{th} percentile in orange), the maximum extent of arid soils from whole record (red), the wettest 6 years (green), and the minimum extent of pedocal/pedalfer boundary (blue), **(c)** average conditions during the drought in the 1930s and 1950s, and **(d)** the central tendency of the 25^{th} percentiles where red is wet line and blue is the drier 25^{th} percentile of the total record.

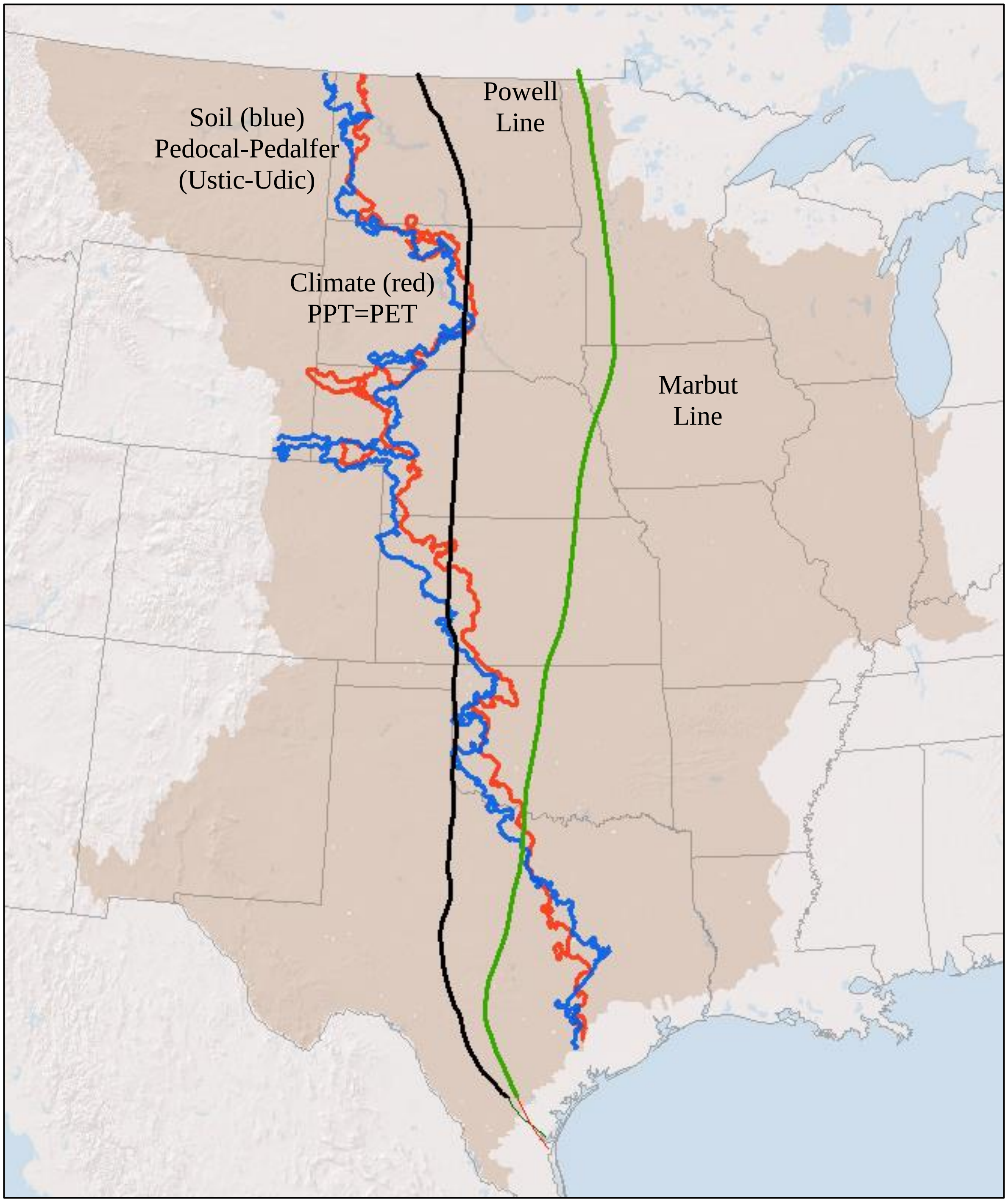
Soil (blue)
Pedocal-Pedalfer
(Ustic-Udic)
Powell
Line
Climate (red)
PPT=PET
Marbut
Line

Figure 5-1

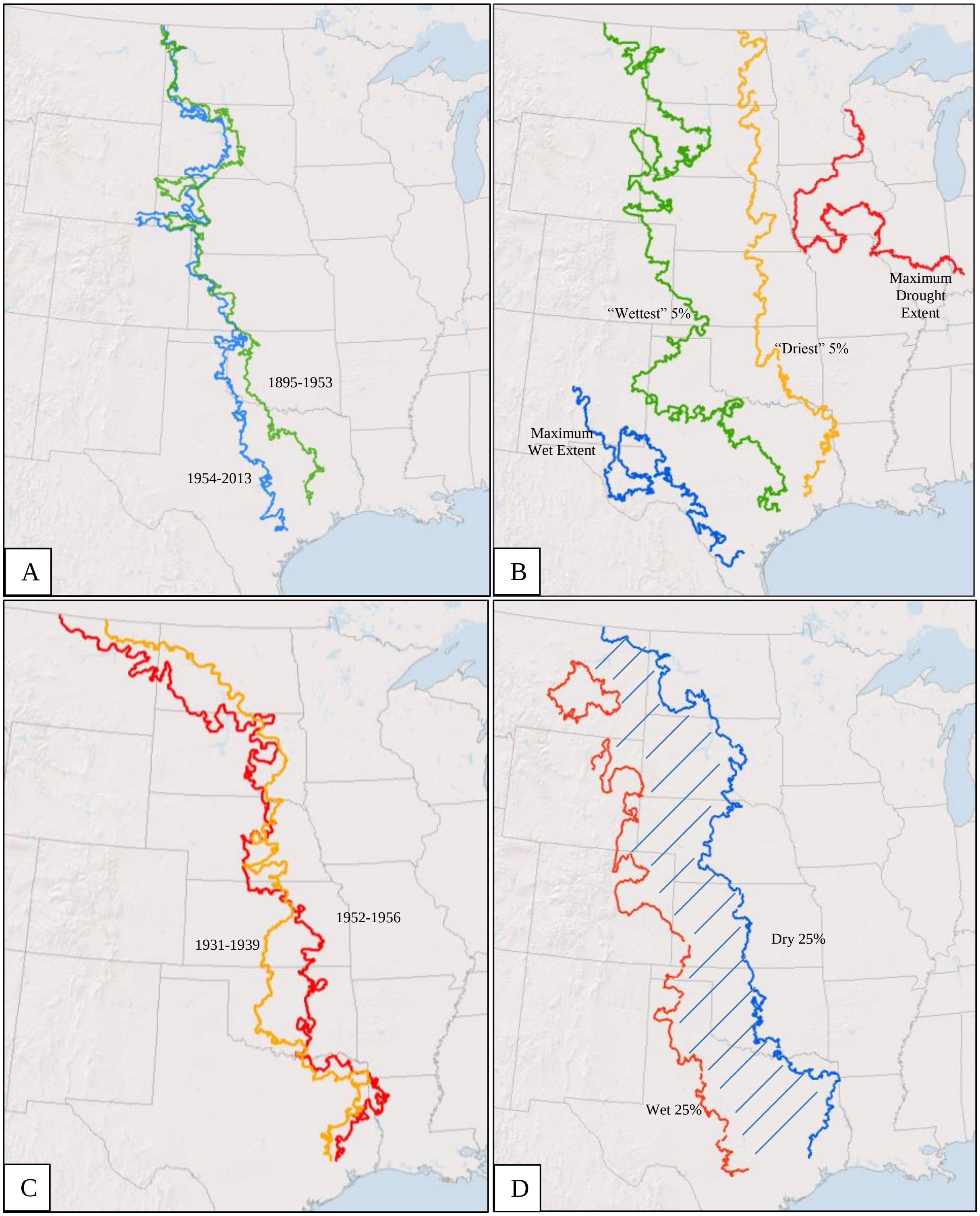
1895-1953
1954-2013
A
"Wettest" 5%
"Driest" 5%
Maximum
Drought
Extent
Maximum
Wet Extent
B
1952-1956
1931-1939
C
Dry 25%
Wet 25%
D

Figure 5-2

5: REFERENCES

Danels D, Johnson G. 2001. Development of a New Precipitation-Effectiveness Index. ESRI User Conference Proceedings.

Machette, MN. 1985. Calcic soils of the southwestern. United States Special Paper 203 Pg 1-21.

Marbut, CF. 1935. Soils of the United States in “Atlas of American Agriculture” (E. 0. Baker, Ed.), part 3, Advance Sheets No. 8. USDA.

Baldwin M, Kellogg CE, Thorp J. 1938. Soil Classification. In Soils and Men: Yearbook of Agriculture 1938. United States Department of Agriculture. United States Government Printing Office, Washington, D.C., USA. Pg 979-992

Miller DA, White RA. 1998. A Conterminous United States Multi-Layer Soil Characteristics Data Set for Regional Climate and Hydrology Modeling. Earth Interactions, 2. Available from: http://EarthInteractions.org.

Newhall F, Berdanier CR. 1996. Calculation of soil moisture regimes from the climatic record. Soil Survey Invest. Rep. 46. Natural Resources Conservation Services, Lincoln, NE.

Nichols JD, Stone JF. 1970. Evaluation of Soil Moisture Measurements in Oklahoma as Soil Characteristics for Classification. Soil Sci Soc Am Journal 34: 4: pg638-641.

Powell, JW. 1879. Report on the lands of the arid region of the United States. United States Government Printing Office.

PRISM Climate Group. 2012. Gridded climate data for the contiguous USA. [cited 2012 Dec 15]. Available from: http://prism.oregonstate.edu

Smith, GD. 1986. The Guy Smith Interview: Rational for Concepts in Soil Taxonomy. SMSS Technical Monograph # 11. Soil Conservation Service, Washington, DC.

[SSS] Soil Survey Staff. 2010. Keys to Soil Taxonomy, 11th ed. USDA-Natural Resources Conservation Service, Washington, DC.

Thornthwaite CW. 1948, An approach towards a rational classification of climate, Geographical Review. 38:55-94.

Waltman SW, Miller DA, Bills B, Waltman WJ. 2011. JAVA Newhall simulation model—Update to a traditional soil climate simulation model. Presented at the Abstract Soil Sci. Soc. Am. Int. Annual Meeting, San Antonio, TX, 18 Oct.

Winzeler HE, Owens PR, Waltman SW, Libohova Z, Waltman WJ. 2012. Soil Climate indicators from the Geographically Explicit Newhall Simulation Model (GEN) as potential environmental covariates in digital soil mapping applications. In: Minasny, Malone, McBatney, editors. Digital Soil Assessments and Beyond: Proceedings of the 5th Global Workshop on Digital Soil Mapping 2012, p 167-71

Winzeler HE, Owens PR, Waltman SW, Waltman WJ, Libohova Z, Beaudette D. 2013. A Methodology for Examining Changes in Soil Climate Geography through time: U.S. Soil Moisure Regimes for the Period of 1971-2000. Soil Sci. Soc. Am. J. 77: 213-25.

USDA-NRCS. 2011. JAVA newhall simulation model—Update to a traditional soil climate simulation model, Abstract. Soil Sci Soc Am Int. Annual Meeting. San Antonio, TX. 18 Oct.

USDA-NRCS. 2012. jNSM 1.6.0 - Java Newhall Simulation Model. [cited 2013 Feb 1]. Available from: http://soils.usda.gov/technical/classification/jNSM/

Zobeck T M, Daugherty LA. 1982. Calculating the depths of soil moisture control sections from laboratory data. Soil Sci. Soc. Am. J. 46:792-5.

CHAPTER 6

SUMMARY AND CONCLUSIONS

6.1 Summary and Conclusions

The primary objective of this work was to elucidate the change in water holding characteristics of soils through pedogenesis and to quantify how global change will impact soil moisture in the U.S. Great Plains. Specifically, this primary objective allowed me to develop a pedohydrologic model of soil profile development as a function of time and define unique pedogenic phases, grouping soil into distinct hydrologic units (e.g, soil hydrologic regimes) to identify soil properties and landscape elements that are either resilient or vulnerable to global change drivers. A secondary objective of this research was to create a chronicle of soil moisture and soil climate for the North American Great Plains as a function of monthly temperature, monthly precipitation, and soil properties. This secondary objective allowed me to quantify the dynamic behavior of soil moisture regionally for the contemporary record and to forecast soil climate regimes across soil landscapes as they are expected to respond with global climate change.

To address these objectives, I analyzed two semi-arid chronosequences (Chapter 2), as well as thirteen additional established soil-chronosequences in the western United States (Chapter 3) to assess biologically available water as a function of soil development. I further used a deterministic soil water balance model to analyze historic patterns and simulate future scenarios of soil moisture (Chapter 4). Results from the soil water balance model allowed me to also assess the long-term dynamics of the Pedocal-Pedalfer in Great Plains of North America (Chapter 5). The following chapter summaries address each objective as they are described above.

6.2 Chapter 2

The results from this chapter show that biologically available water in semi-arid ecosystems exhibit predictable changes in water as a function of soil development. Through analysis of two semi-arid chronosequences I showed that distinct pedohydrologic phases occur across soil age gradients. Results show that in semi-arid ecosystems, broad stages of soil development exist and are linked to landscape ages that are ecologically and biogeochemically significant: an aggrading, an intermediate phase where plant available water capacity reaches a maximum, and a regressive phase where long-term weathering has altered soil properties (e.g., structure, texture, and chemical composition) where biologically available water declines.

Soil texture was a strong determinant in available water holding capacity with the Wind River's sandy soils holding less available water then the loamy soil gradient in the Colorado Piedmont. I also found that in intermediate soil ages where clay and carbonates continue to increase, biologically available water begins to decrease. At the oldest soil age stages of soil development, water retention as well as other soil properties decreases, including organic matter, clays, and carbonates. Results from this study indicate that the quantity of biologically available water is directly tied to soil age and this work helps improve the understanding of long-term ecosystem biophysical feedback through soil moisture retention characteristics.

6.3 Chapter 3

To assess our model of conditioning of biologically available water as a function of soil development I modeled soil hydrologic properties for nine established soil-chronosequences across various semi-arid and arid climates of North America. Changes in soil bulk density, volumetric dilation and collapse, mineral composition, and organic matter ultimately drive shifts in the ecosystem's biologically available water. Results showed that initial stages of soil formation generally increase water holding capacity, and absent of rejuvenating disturbance, pedogenesis ultimately leads to a total reduction in biologically available water on longtime scales. My results suggest that long-term development of ecosystems imparts a predictable pattern regarding changes in biologically available water and propose that soil hydraulic properties ultimately play an important role in the evolution of ecosystems, providing an important biophysical feedback to soil forming processes.

6.4 Chapter 4

By simulating the contemporary record of soil moisture, we are able to capture the range and variability of soil moisture conditions in the Great Plains of North America. Results from Chapter 4 show the variability of soil climate in response to annual precipitation and temperature patterns. Analysis of annual and seasonal soil water balances in the dry western portion of the Great Plains confirm that soil moisture is subject to short-term precipitation events. These results support the suggestion that soil landscapes with course textural soils and low water holding capacities could be more sensitive to changing climate drivers. Projections of soil moisture response to climate change are less dependable due to uncertainty in precipitation patterns; however, patterns of soil temperature regimes tied to increasing global warming seem to be more consistent. Soil classes described in this chapter are a part of a framework for a soil-hydrologic template in which landforms are stratified to understand soil moisture's response to global change. This chapter proposes three functional phases that can be determined by soil age (chapter 3), climatic change or variability, human induced disturbance, geologic substrate, and / or landscape position. This research builds on the functional soil phase model described in chapters 2 and 3 to incorporate short-term extreme climate perturbations and management of soil resources into a soil hydrologic template.

6.4 Chapter 5

Geospatial data derived from the Newhall Simulation Model in Chapter 5 has identified a generalized western limit in the United States in which agroecosytems would have to be either drought tolerant or irrigated to grow successfully and predictably. The work presented here also suggests that the extreme inter-annual variability of climate on the western portions of the North America Great Plains limits reliable forecasting of soil moisture as we note that soil moisture response is highly dependent on the intensity and timing of precipitation. Informed management decisions regarding climate change mitigation for western lands require a clear understanding of pedohydrologic functions of landscapes, as we conclude that the greatest degree of vulnerability for an ecosystem to global climate change is reflected in the soil’s hydrologic regime.