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Genesis of argillic horizons in soils of the Charkiln Series, Spring Mountains, Clark County, Nevada

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GENESIS OF ARGILLIC HORIZONS IN SOILS OF THE CHARKILN SERIES,

SPRING MOUNTAINS, CLARK COUNTY, NEVADA

by

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ABSTRACT

Genesis of Argillic Horizons in Soils of the Charkiln Series, Spring Mountains, Clark County, Nevada

by

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Understanding the pedogenesis of argillic horizons in atypical parent materials and climates provides a valuable context for understanding traditional pedogenic interpretations. In this study I examine the genesis of argillic horizons in soils forming in alluvium dominated by quartzite (an atypical parent material for argillic horizons) in an arid to semi-arid climate (usually insufficient moisture to translocate clays). Using soil physical, chemical, and mineralogical analyses with field soil mapping I examined whether the argillic horizons are currently forming and their potential source materials. Results suggest that argillic horizons formed from quartzite, limestone, and eolian dust. While mineralogy suggests soil weathering is minimal, A horizons are clay-depleted and B horizons contain actively forming channel argillans and clay accumulation; a lack of lithologic discontinuities also suggests the argillic horizons are not products of a past climate. This study's results provide new insight into argillic-horizon development in atypical parent materials and climates.

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CHAPTER 1

INTRODUCTION

An argillic horizon is a subsurface soil horizon that exhibits evidence of clay illuviation (accumulation), has a higher percentage of phyllosilicate clay than overlying horizons, and typically requires at least a few thousand years to form (Soil Survey Staff, 1999). Argillic-horizon formation occurs from clay eluviation/illuviation via *in situ* weathering from the breakdown of primary and secondary minerals (Soil Survey Staff, 1999). The weathering rates of minerals contributing to argillic-horizon formation within the soil profile are a function of climate and parent material (Allen and Hajek, 1989). Mineral transformation (primary to secondary mineral formation) and translocation of minerals via water are common soil-forming processes responsible for clay movement and thus the presence of clays in argillic horizons (Soil Survey Staff, 1999).

Argillic horizons can form under current climatic regimes that undergo wetting and drying cycles (Aandahl, 1982; Soil Survey Staff, 1999) which lead to clay dispersion, deposition, and translocation (Soil Survey Staff, 1999). Argillic horizons also can be preserved from a formation process that has occurred in a past climate (Nettleton et al., 1989), and thus be a feature of a paleosol (Ruhe, 1965, 1975; Birkeland, 1999).

The identification of argillic-horizon features in a soil is complicated by the interpretation of the soil as a paleosol, a currently forming soil, or a soil forming in/on a relict soil. Paleosol argillic horizons may act as a barrier to water flow and enhance the formation of current argillic horizons, or contribute to the formation of new argillic horizons, further complicating interpretation of the soils as paleosols or currently forming soils. If the argillic horizon is part of a soil that has not been buried or eroded, it may be a relict paleosol-a soil that since its initial formation has remained at land surface and has not been buried by younger sediment (Ruhe, 1965, 1975; Birkeland, 1999).

The definition of an argillic horizon is subject to revision as research progresses. For example, much of the past research on argillic horizons has occurred in soils of the eastern United States (Smeck et al., 1968; Bilzi and Ciolkosz, 1977; Franzmeier et al., 1985; Ciolkosz et al., 1989; Goenadi and Tan, 1989; Nettleton et al., 1989; Stolt and Rabenhorst, 1991), and support the Soil Survey Staff's (1999) hypothesis that argillic horizons usually require at least a few thousand years to form. Studies west of the Mississippi River from North Dakota to Texas, and as far west as California, indicate that many argillic horizons in Midwestern and western soils are relict features of the Pleistocene (Gile and Grossman, 1968; Gile and Hawley, 1968; Nettleton et al., 1975; Sobecki and Wilding, 1983; Nettleton et al., 1989) or early to middle Holocene (Gile, 1975; Parsons and Herriman, 1976; Southard and Southard, 1985; Hopkins and Franzen, 2003). The western Unites States, however, currently is dominated by an aridic soil moisture regime (NRCS, 2006c), which may not have the necessary moisture and precipitation to promote argillic-horizon formation (Gile and Grossman, 1968; Gile and Hawley, 1968; Nettleton et al., 1989). However, exceptions can exist where quicker formation occurs if environmental conditions (leading to physical and chemical weathering) favor argillic-horizon formation. An example is the research by Graham and Wood (1991) and Johnson-Maynard et al. (2004), who found that an argillic horizon formed in approximately 50 years in disturbed soil under Coulter pine (*Pinus coulteri* B. Don) in lysimeters at the San Dimas Experimental Forest in southern California. Illuviation of clays was aided by the production of organic acids, absence of burrowing macrofauna, and water available for leaching (Johnson-Maynard et al., 2004).

Purpose of study

This research focuses on a soil series with argillic horizons [Charkiln series (fine-loamy, mixed, superactive, mesic Aridic Argiustoll; NRCS, 2006b; 2006d)] in the Spring Mountains, Clark County, Nevada (Figure 1), which is forming in alluvium dominated by quartzite; quartzite is an atypical parent material for soils with argillic horizons (Ogg and Baker, 1999). Ogg and Baker (1999) and Ogg et al. (2001) studied soils with argillic horizons in the Virginia Blue Ridge that formed from quartzite parent material. Based on clay mineralogy and clay percentages, they concluded that it was unusual for the soils to have formed only from a quartzite parent material; argillic horizons are less likely to form from quartzite because it is composed essentially of quartz (Klein and Hurlbut, 1993) and therefore lacks the clays needed to form argillic horizons. Other sediments and lithologies in addition to the quartzite were responsible for the formation of the soils in Virginia (Ogg and Baker, 1999; Ogg et al., 2001). It is hypothesized in this study that the clay mineralogy for the argillic horizons is not forming from the quartzite but from upslope contributions from a limestone parent material and eolian dust. However, the quartzite could contain impurities that contribute to argillic-horizon formation, and consequently may be an important factor in the formation of the argillic horizon. Therefore, the objective of this study is to examine three pedons of the Charkiln Series, the associated climate, soil-forming processes, and clay mineralogy to determine (1) the genesis of the argillic horizons, and (2) whether the argillic horizons are relict or currently forming.

Significance

Studies on the genesis of argillic horizons in quartzite parent material are limited to the research of Ogg and Baker (1999) and Ogg et al. (2001). No previous studies have examined argillic-horizon formation in the Spring Mountains of Nevada, therefore, limited knowledge exists on their occurrence. Additionally, it is hypothesized in this study that the argillic horizons are young, and forming in the present climate which contradicts past research on argillic-horizon formation in general (Gile and Hawley, 1968; Nettleton et al., 1989). This study will help the U.S. Department of Agriculture-Natural Resources Conservation Service (USDA-NRCS) better understand the genesis of these soils and aid in future landscape interpretations where such soils occur.

Study area location

The study area is located in the northwest Spring Mountains, Clark County, Nevada, approximately 117 km northwest of Las Vegas (Figure 1). The Spring Mountains are bordered on the east by the Las Vegas valley and to the west by the Pahrump valley. The pedons discussed in this study are located in an east-west transect along Wheeler Pass Road on the west side of the mountain range northwest of Charleston Peak. The pedons range in elevation from about 2,032 to 2,424 m (USGS, 1984), and site coordinates [in North American Datum (NAD) 1983] for the pedons are as follows: pedon C1 is located at 36° 21' 59" north latitude and 115° 49' 16" west longitude; pedon C2 is located at 36° 22' 04" north latitude and 115° 48' 36" west longitude; pedon C3 is located at 36° 22' 13" north latitude and 115° 47' 57" west longitude; and the Ts pedon is located at 36° 22' 27" north latitude and 115° 46' 46" west longitude.

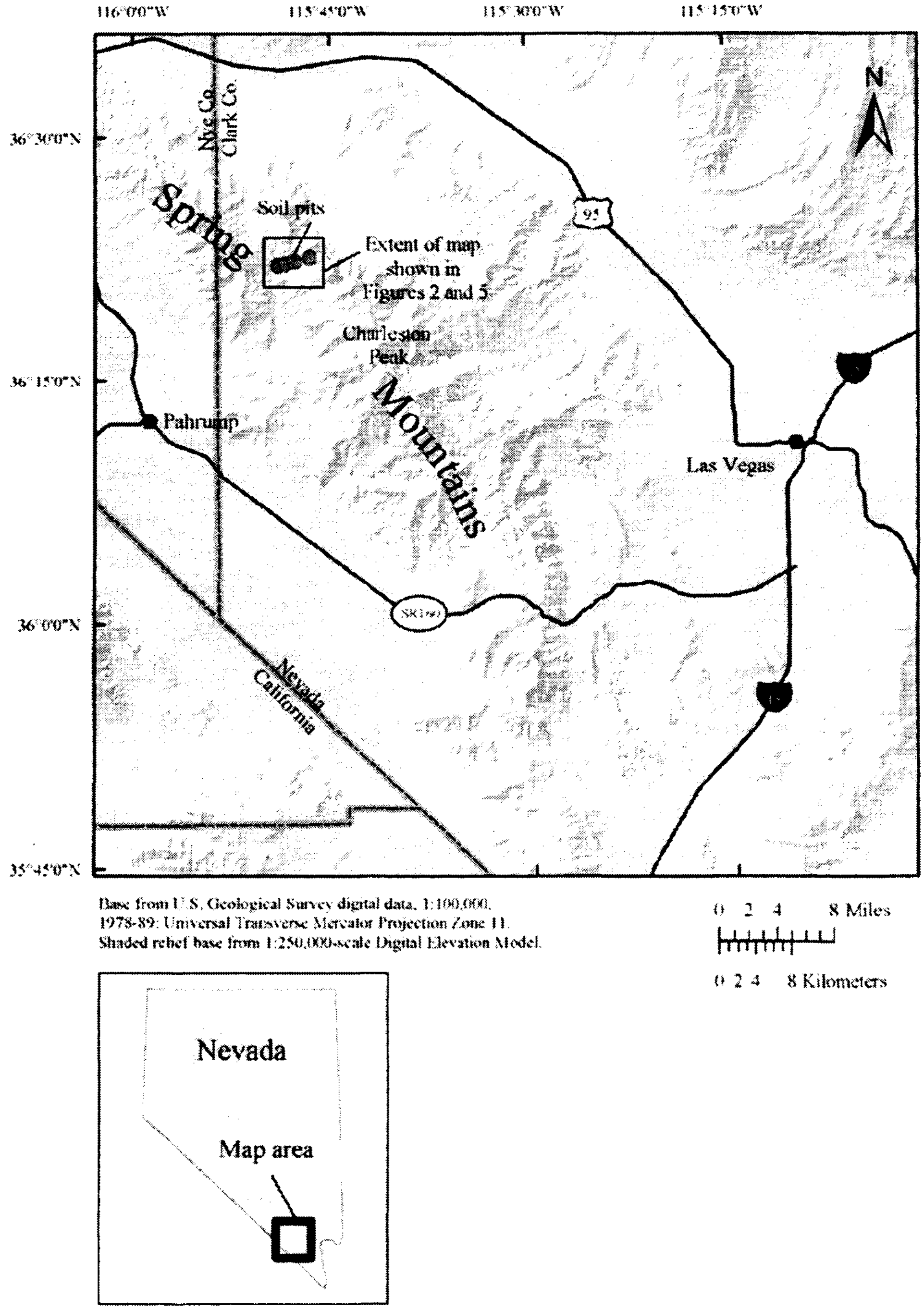
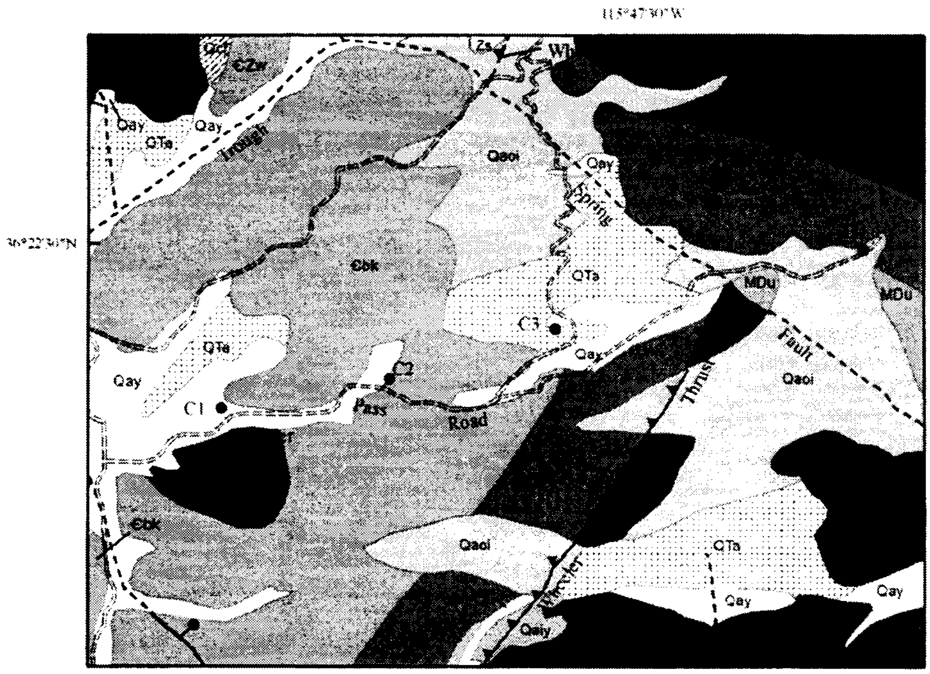


Figure 1. Study location, and geographic and physiographic features of the Spring Mountains, Clark County, Nevada

Geology

The geology of the Spring Mountains, as described by Longwell et al. (1965) and Page et al. (2005), is complex and extensively faulted in the northern half of the range (Figure 2). Lithology of the study area as summarized from Page et al. (2005) consists of Precambrian and Paleozoic quartzite; Paleozoic sedimentary rocks, including mostly limestone, dolostone, sandstone, siltstone, and shale, and some quartzite; and Cenozoic alluvium, colluvium, and fan alluvium ranging from Miocene to Holocene age. Structure of the area consists of thrust faults and abundant normal faults. Geomorphic surfaces located throughout the study area include fan remnants, inset fans (NRCS, 2006d), active to inactive washes exhibiting various stages of bar and channel topography, and many exposures of bedrock.



Legend

C1● Soil sites

Normal fault-ball and bar on downthrown side; dashed where inferred

Thrust fault-sawteeth on upper plate; dashed where inferred

==== Road

Geologic units

Qay Holocene and Pleistocene, young fan alluvium

Qcl Holocene and Pleistocene, hillslope colluvium and alluvium

Qaiy Late Pleistocene, younger intermediate fan alluvium

Qaoi Pleistocene, undivided alluvium

QTa Pleistocene to late Miocene? Gravelly basin-fill alluvium

Lower Permian to Upper Mississippian, Bird Spring Formation, undivided: limestone, dolostone, sandstone, siltstone, shale, chert, quartzite

Upper and Lower Mississippian, Monte Cristo Group: limestone, quartzite, chert

MDu Lower Mississippian and Upper and Middle Devonian, Mississippian and Devonian rocks, undivided: limestone, quartzite, dolostone

Middle Ordovician, Eureka Quartzite: quartzite

Middle Ordovician to Upper Cambrian, Pogonip Group: limestone, dolostone, shale, chert

Cbk Upper and Middle Cambrian, Bonanza King Formation: dolostone, limestone, chert

Middle and Lower Cambrian, Carrara Formation: limestone, siltstone, sandstone, shale

CZw Lower Cambrian and Late Proterozoic, Wood Canyon Formation: quartzite, sandstone, siltstone, shale

Zs Late Proterozoic, Stirling Quartzite: quartzite

Figure 2. Geology of the northern Spring Mountains, Clark County, Nevada (From Page et al., 2005).

Climate

The soil moisture regime of the study area is aridic to ustic, with a mesic temperature regime, and a mean annual precipitation rate of 38 to 41 centimeters (NRCS, 2006d). Water supply is sufficient to support a pinyon (*Pinus monophylla* Torr. & Frém.)-juniper [*Juniperus osteosperma* (Torr.)] woodland with mountain mahogany (*Cercocarpus ledifolius* Nutt. var. *intermontanus* N. Holmgren) and Gambel oak (*Quercus gambelii* Nutt.; Houghton et al., 1975). The climate of the study area, as described by Houghton et al. (1975), is characteristic of a subhumid continental climate with cold winters and moderate precipitation. The region typically experiences subfreezing temperatures in the higher elevations. Prevailing westerlies provide much of the moisture for the region. Most precipitation comes from Pacific storms, Great Basin lows, or convective rainfall during the summer monsoons. Pacific storms typically produce precipitation in the winter months from December through February. Great Basin lows produce precipitation over most of Nevada from April to June, and summer monsoons produce thunderstorms that bring the maximum rainfall to southeast Nevada during July and August. In the Spring Mountains, clay formation likely is greater during the summer monsoon season due to higher temperature and precipitation (Jenny, 1994) than during other times of the year.

Vegetation

Vegetation at the pedons of the Charkiln series ranges from a singleleaf pinyon pine (*Pinus monophylla* Torr. & Frém.), Utah juniper [*Juniperus osteosperma* (Torr.)], and Gambel oak (*Quercus gambelii* Nutt.) association, with mountain big sagebrush [*Artemisia tridentata* Nutt. ssp. *vaseyana* (Rydb.) Beetle], and prickly pear cactus

(*Opuntia* P. Mill.) in association (Figure 3; NRCS, 2006b; 2006d). Vegetation at the pedon of the Troughspring series, a nearby soil, contains singleleaf pinyon pine (*Pinus monophylla* Torr. & Frém.) with the following in association: curl-leaf mountain mahogany (*Cercocarpus ledifolius* Nutt. var. *intermontanus* N. Holmgren), Utah serviceberry (*Amelanchier utahensis* Koehne), Gambel oak (*Quercus gambelii* Nutt.), and (yellow) silktassle (*Garrya* Dougl. Ex Lindl.) (Figure 4; NRCS, 2006b; 2006d).

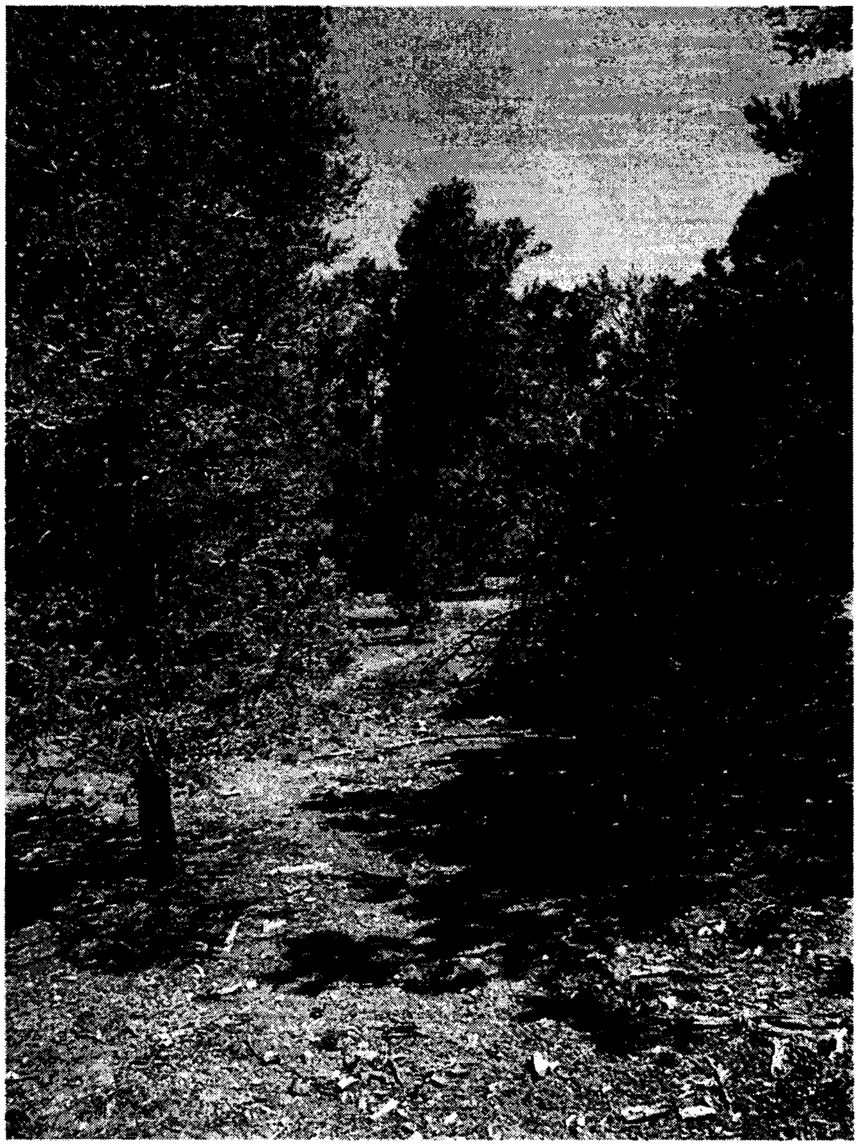


Figure 3. Vegetation at the C2 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Photograph taken by Doug Merkler, NRCS, 2004.

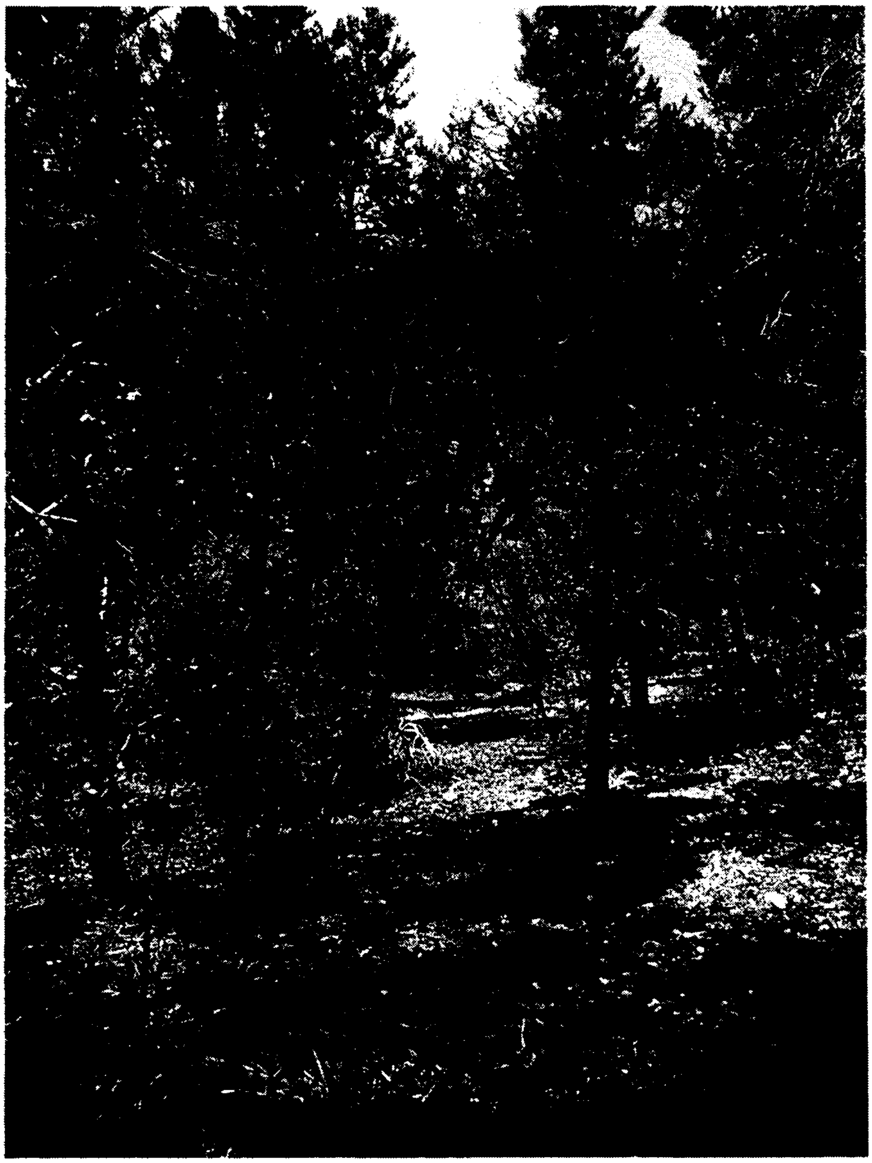


Figure 4. Vegetation at the Ts pedon, Troughspring series, Spring Mountains, Clark County, Nevada. Photograph taken by Doug Merkler, NRCS, 2004.

CHAPTER 2

METHODOLOGY

To derive the geomorphic history of the study area, aerial photograph stereoscopic pairs were viewed three dimensionally using stereoscopic glasses. The aerial photographs, from the U.S. Department of Agriculture, are color photographs at an approximate scale of 1:44,568 and were taken on September 7, 2003. Geomorphic units were mapped on Mylar overlain on the aerial photographs. The map units were transferred into ArcGIS 9 ArcMap using a U. S. Geological Survey, 1-meter resolution, digital orthophotograph quadrangle of the study area as the base layer.

Soil samples from three pedons of the Charkiln series (C1, C2, and C3) and from one pedon of the Troughspring series (Ts) were examined in this study. Soils were sampled and described by the Natural Resources Conservation Service and the University of Nevada, Las Vegas (UNLV) during a soil survey of the Clark County area (NRCS, 2006d) in the summer of 2004, using methods discussed in the "Field Book for Describing and Sampling Soils" by Schoeneberger et al. (2002). Data collected from the soil pits included soil texture, structure, color, horizon depths, thickness, rooting depth, pores, percent rock fragments and composition, salt content, chemical properties (pH, percent $CaCO_3$, effervescence), clay films, type of boundaries, and field observed macrostructures. Pedon locations and the elevation of each soil pit were recorded using a Global Positioning System (GPS). Slope and aspect of the landscape were determined via a clinometer and compass. Approximately 4 kg of soil sample were collected from each horizon and stored in plastic storage bags. Samples from the C2 and Ts pedons were analyzed by the NRCS Soil Laboratory in Lincoln, Nebraska and published as part of a soil survey of Clark County area, Nevada; pedons C2 and Ts are the type locations for the study area (NRCS, 2006d). Samples from the two remaining pedons, C1 and C3, were analyzed at the UNLV Soil Laboratory, unless otherwise noted. Sub-samples of the C2 and Ts pedons also were analyzed at UNLV as a crosscheck of laboratory procedures.

Percent sand, silt, and clay were obtained using the hydrometer method described by Gee and Bauder (1986). Silt-to-clay ratios were calculated by dividing the percent silt by the percent clay. Moisture content and pH measurements were completed in the UNLV Soil Laboratory following methods described by Burt (2004). Soil pH was estimated in a 0.01 M $CaCl_2$ solution (Burt, 2004). Samples for phosphorous analysis were prepared in the UNLV Soil Laboratory following the Olsen sodium bicarbonate extraction method (Burt, 2004). A Thermo Spectronic Helios Gamma spectrophotometer, in the UNLV Biology Department, was used to measure the phosphorus concentration of each sample mixture. Samples were prepared for base-cation analysis using a mechanical vacuum extractor and the ammonium-acetate method as described by Burt (2004). The samples were analyzed using a Perkin Elmer AAnalyst 400 atomic absorption spectrometer. For cation exchange capacity (CEC), air-dried soil samples were sent to the NRCS Soil Laboratory in Lincoln, Nebraska and processed using the ammonium-acetate method as described by Burt (2004). Carbon and nitrogen were measured using an Exeter CHN analyzer by the UNLV Chemistry Department.

Mineralogy

Relative amounts of minerals were identified using x-ray diffraction (XRD) according to methods described by Thurman et al. (1994) and Burt (2004). Slides were analyzed on a PANalytical X'Pert PRO X-ray diffractometer in the UNLV XRF/XRD Laboratory using the following settings: 40 kV, 30 mA, from 3 to 30 degrees, 2θ, with a step size of 0.0167, time per step was 25 seconds, scan speed of 0.086, with ¼ degree divergence slit and ½ degree anti-scatter slit.

Dithionite-citrate extraction was used on the <2mm fraction of all genetic horizons to determine the pedogenically active free iron oxide, and thus the degree of weathering attributed to pedogenic iron in the soil (Burt, 2004; Weisenborn and Schaetzl, 2005). Dithionite-citrate (CD) iron to acid ammonium oxalate (AAO) iron ratios provide an estimate of the relative age of the soils. Younger soils have less free iron oxides (Shaw et al., 2003; Burt, 2004, p. 304; Weisenborn and Schaetzl, 2005). The soil samples were analyzed using the CD extraction method described by Burt (2004). The sample extracts and calibration standards for Fe and Mn were analyzed using a Perkin Elmer AAnalyst 400 atomic absorption spectrometer.

Thin sections were made from clods collected from the C1 and Ts pedons, and rock fragments from the C1 and C3 pedons representative of each B horizon, using methods described by Burt (2004). A staple was placed on the top of each clod to indicate orientation, the clods were wrapped in netting and dipped in Saran to coat the clods and prevent them from breaking apart. The Saran-coated clods were shipped to Spectrum Petrographics, Inc., in Vancouver, Washington where they made thin sections from the clods. Thin sections of rock fragments were made in the UNLV Rock Laboratory. A petrographic microscope was used for mineral identification on all of the thin sections, to search for evidence of clay illuviation, and for point counts on the rock-fragment thin sections. Thin-section analysis was done based on methods described by Cady et al. (1986) and Brewer (1976). Clays indicative of translocation were identified as distinct, brownish linings along pores (channel argillans) and grains (grain argillans). A JEOL scanning electron microscope (SEM), model JSM-5610 located in the UNLV Electron Microanalysis & Imaging Laboratory, was used to cross check mineral identification. The NRCS made thin sections of the sand-silt mineralogy (2.0-0.002 mm) from the Bt1 horizon of the C2 pedon, and conducted optical grain counts of mineralogy.

CHAPTER 3

RESULTS

Geomorphic analysis

Geomorphic units throughout the study area consist of fans (NRCS, 2006d), inactive and active washes, and bedrock (Figure 5). Due to the scale at which the geomorphology was mapped (approximately 1:44,568), only limited units are shown. These units include fan, fan apron, and active wash. For this study, the geomorphic unit fan includes fan remnants and inset fans (NRCS, 2006d) which have been combined for simplicity. The geomorphic unit fan represents those landforms that consist of alluvium and sediment that have filled intramontane basins, and includes fans that are relict to currently forming (Peterson, 1981). Fans occur throughout the study area, and soils of the Charkiln and Troughspring series formed on fan remnants (NRCS, 2006b; 2006d). The geomorphic unit fan apron is a component landform of a fan (Peterson, 1981), but for this study it has been excluded from the unit fan because of its distinct characteristics within the study area. The fan apron consists of alluvium derived from gullies and inset fans upslope of the fan apron (Peterson, 1981). In the study area, the unit fan apron is older than the active washes and younger than the fans, exhibits remnants of bar and channel topography, and topographically is situated above, below, and adjacent to active washes (Doug Merkler, NRCS, written commun., 2006). Active washes are incised into their associated floodplains, consist of modern alluvium, void of vegetation (Peterson, 1981), and have distinct bar and channel topography.

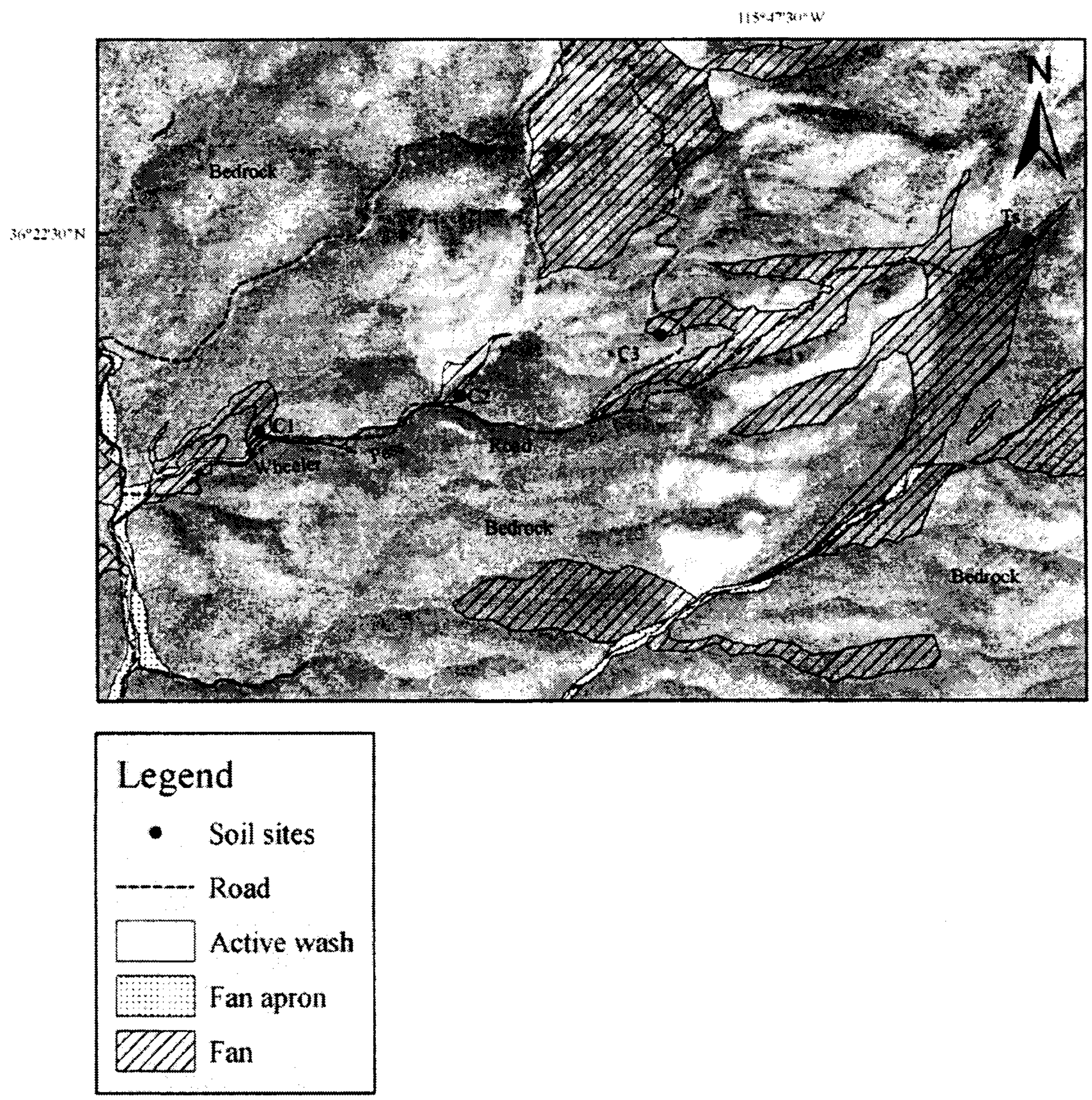


Figure 5. Geomorphic map of the northern Spring Mountains, Clark County, Nevada. Base layer is a U.S. Geological Survey 1-meter resolution, digital orthophotograph quadrangle of the study area.

Physical analyses

Soils of the Charkiln series are fine-loamy, mixed, superactive, mesic, Aridic Argiustolls, and soils of the Troughspring series are loamy-skeletal, carbonatic, mesic, Petrocalcic Paleustolls (NRCS, 2006b; 2006d). Both soil series contain mollic epipedons, soils of the Charkiln series contain argillic horizons (Bt; Figure 6; Tables 1-4), and soils of the Troughspring series contain calcic (Btk, Bk1) and petrocalcic (2Bk2 and 2Bkm) horizons (Figure 7). In general, color moist and dry hues in the soils of the Charkiln series range from 10YR in the upper O, A, and AB horizons, to 7.5YR hues in the Bt horizons. The Bt3 horizons in the C1 and C2 pedons, and all of the horizons in the Ts pedon have a moist and dry color in the 10YR hue. Texture classes in the soils of the Charkiln series ranged from fine-sandy to sandy loams in the O, A, and AB horizons, to loams, sandy-clay loams, and clay loams in the Bt horizons. Texture in the Ts pedon consisted of silt loams. Structure of the Oi and A horizons of all four soils ranged from single grain to platy or subangular blocky. B horizons were angular to subangular blocky, and the deepest B horizons in the soils of the Charkiln and Troughspring series were massive. Consistence of the soils increased with depth as did moisture content which affects consistency (Schoeneberger et al., 2002). Roots were present in all A and B horizons in the soils of the Charkiln and Troughspring series except in the 2Bk2 and 2Bkm horizons in the Ts pedon. Roots and pores ranged from few to many. Roots were very fine to coarse in size, and pores were very fine to medium in size. Pores were tubular to interstitial. Soils of the Charkiln series were noneffervescent, whereas all of the B horizons in the Ts pedon ranged from strongly to violently effervescent due to the presence of calcium carbonate.

Figure 6. The C2 pedon soil pit, Charkiln series, Spring Mountains, Clark County, Nevada (Photograph taken by Doug Merkler, NRCS, 2004).

Figure 7. The Ts pedon soil pit, Troughspring series, Spring Mountains, Clark County, Nevada (Photograph taken by Doug Merkler, NRCS, 2004).

Sand (0.05-2.0 mm) was the dominant particle size in each of the three pedons of the Charkiln series (C1, C2, and C3), followed by silt (0.002-0.05 mm), and clay (Tables 1-3). Although sand was dominant, net decreases in sand occurred with depth in all three soils (the term "net" is used here to indicate an increase or decrease occurred from the top most horizon to the bottom most horizon regardless of what changes occurred in between, that is, after all the changes the net result was an increase/decrease). Net increases in clay with depth also occurred in all three soils. In the C1 pedon sand was highest in the A horizon at 74.8% (Table 1). Net increases in both silt and clay occurred with depth in the C1 pedon. Silt was highest in the Bt3 horizon at 31.0%, whereas clay was highest in the Bt2 horizon at 21.6%. In the C2 pedon, sand was highest in the A and ABt horizons at 56.7%, whereas silt was lowest in the ABt horizon and clay was lowest in the A horizon. Silt and clay were highest in the Bt2 (35.1%) and Bt3 (22.8%) horizons, respectively (Table 2; Soil Survey Staff, 2006). In the C3 pedon, sand was highest in the A1 horizon at 66.7%, silt was highest in the A2 horizon at 30.4%, and clay was highest in the Bt1 horizon at 35.2% (Table 3). Silt-to-clay ratios in the soils of the Charkiln series showed net decreases with depth in all three pedons.

Silt was the dominant particle size in the soil of the Troughspring series, with about an even distribution of both sand and clay (Table 4; Soil Survey Staff, 2006). Silt was highest in the 2Bk2 horizon at 54.8%, sand was highest in the 2Bkm horizon at 28.6%, and clay was highest in the Btk horizon at 25.7%. Silt to clay ratios in the Ts pedon indicated a net increase with depth.

Table 1. Soil characteristics of the C1 pedon, Charkiln series, Spring Mountains, Nevada. Data from Doug Merkler, NRCS, written commun., (2006), unless otherwise noted.

Horizon	Depth (cm)	Color moist	Color dry	Sand[1]	Silt[1]	Clay[1]	USDA texture class[1]
				----------	%	----------	
Oi	0-3	10YR2/2	10YR 3/2	--	--	--	--
A	3-12	10YR 3/2	10YR 4/3	74.8	20.2	5.0	sandy loam
AB	12-24	10YR 3/2	10YR 4/3	67.7	24.7	7.6	sandy loam
Bt1	24-36	7.5YR 3/3	7.5YR 5/3	51.6	30.6	17.8	loam
Bt2	36-118	7.5YR 3/4	7.5YR 4/4	51.0	27.4	21.6	sandy clay loam
Bt3	118-166	10YR 4/4	10YR 5/4	49.7	31.0	19.3	loam

	Structure	Consistence Dry/ moist	Stickiness/ plasticity	Roots	Pores	Boun-dary	Effer-vescence
Oi	sbk,2,f-m[2]	s/vfr	so/po	--	--	AW	--
A	pl,2,m	s/vfr	so/po	3vf-f	2vf,i,1vf,v	CS	NE
AB	sbk,3,m	sh/fr	ss/sp	1c,2f-m	2vf,i,1vf,t	CW	NE
Bt1	abk,3,f-m	vh/efi	vs/p	3m-c,2f-vf	1vf,i,2m-f,t	CW	NE
Bt2	abk,3,m-co	vh/efi	s/p	2m,1f-c	1vf-f,t	CW	NE
Bt3	m	mh/fi	s/p	1f-m	2vf,i,1f,t	--	NE

	Sand fraction[1] % Very coarse	Coarse	Medium	Fine	Very fine	CEC cmol(+) kg^{-1}	pH[1]
Oi	--	--	--	--	--	--	--
A	1.8	2.7	2.7	4.5	10.4	9.7	7.2
AB	2.2	3.4	3.6	5.1	9.6	10.9	7.4
Bt1	1.4	2.3	3.1	5.0	9.8	14.1	7.6
Bt2	1.3	2.1	2.7	4.4	8.5	13.0	6.4
Bt3	3.4	3.6	3.0	3.5	6.7	13.8	7.2

	%C[3]	%N[3]	P[1] (mg kg-1)	Clay films
Oi	3.03	0.14	--	--
A	1.86	0.18	27.68	--
AB	0.89	0.03	22.19	f,f,clf,pf
Bt1	0.87	0.02	27.31	m,d,7.5YR 5/6,brf,pf
Bt2	0.43	0.01	14.65	m,d,7.5YR 5/6,clf,brf,apf
Bt3	0.25	0.05	13.58	f,f,10YR 4/6,clf,mat

[1]- Analyzed in the University of Nevada Las Vegas (UNLV) Soil Laboratory

[2]- See Schoeneberger et al. (2002) for abbreviation definitions

[3]- Analyzed at the UNLV Chemistry Department

Table 2. Soil characteristics of the C2 pedon, Charkiln series, Spring Mountains, Nevada. Data from NRCS (2006b; 2006d); Soil Survey Staff (2006).

Horizon	Depth (cm)	Color moist	Color dry	Sand	Silt	Clay	USDA texture class
				---%	---%	---%	
Oi	0-3	10YR 3/2	10YR 3/2	--	--	--	--
A	3-13	10YR 3/2	10YR 4/3	56.7	34.8	8.5	fine sandy loam
ABt	13-23	10YR 4/2	10YR 4/3	56.7	29.5	13.8	fine sandy loam
Bt1	23-36	7.5YR 3/3	7.5YR 5/3	47.3	31.4	21.3	loam
Bt2	36-117	7.5YR 3/4	7.5YR 4/4	42.4	35.1	22.5	loam
Bt3	117-165	10YR 4/4	10YR 5/4	43.9	33.3	22.8	loam

	Structure	Consistence Dry/ moist	Stickiness/ plasticity	Roots	Pores	Boun- dary	Effer- vescence
Oi	sbk,2,f-m[1]	s/vfr	so/po	--	--	AW	NE
A	pl,2,m	s/vfr	so/po	3vf-f	2vf,t,3vf,i	CS	NE
ABt	sbk,3,m	sh/fr	ss/ps	2f,m,3co	2vf,t,3vf,i	CW	NE
Bt1	abk,3,f-m	vh/efi	s/p	3vf-co	3f,m,t, 2vf,i	CW	NE
Bt2	abk,3,m-co	vh/efi	s/p	2f,co,3m	2vf-f,t	CW	NE
Bt3	m	mh/fi	s/p	2f,m	2f,t,3vf,i	--	NE

	Sand fraction % Very coarse	Coarse	Medium	Fine	Very fine	CEC cmol(+) kg^{-1}	pH
Oi	--	--	--	--	--	--	6.4
A	4.6	9.6	9.1	12.7	20.7	11.3	7.8
ABt	4.7	8.1	9.4	14.8	19.7	11.8	8.2
Bt1	2.6	6.3	8.7	11.5	18.2	14.4	8.2
Bt2	2.9	5.8	7.6	12.0	14.1	13.6	8.2
Bt3	3.2	11.1	7.8	8.2	13.6	13.7	8.2

	%C	%N	Clay films
Oi	17.10	0.654	--
A	1.64	0.100	--
ABt	1.01	0.083	vf,f,clf,pf
Bt1	0.69	0.081	c,d,7.5YR 5/6,clf,brf,apf
Bt2	0.32	0.057	m,d,7.5YR 5/6,clf,brf,pf
Bt3	0.30	0.050	vf,f,10YR 4/6,clf,rf,mat

[1]- See Schoeneberger et al. (2002) for abbreviation definitions

Table 3. Soil characteristics of the C3 pedon, Charkiln series, Spring Mountains, Nevada. Data from Doug Merkler, NRCS, written commun., (2006), unless otherwise noted.

Horizon	Depth (cm)	Color moist	Color dry	Sand[1]	Silt[1]	Clay[1]	USDA texture class[1]
				---------------%---------------			
A1	0-3	10YR 3/2	10YR 5/2	66.7	26.9	6.4	sandy loam
A2	3-12	10YR 3/2	10YR 5/2	54.4	30.4	15.2	sandy loam
Bt1	12-50	7.5YR 4/4	7.5YR 4/4	44.1	20.7	35.2	clay loam
Bt2	50-110	7.5YR 4/6	7.5YR 4/6	61.6	17.9	20.5	sandy clay loam

		Consistence					
	Structure	Dry/ moist	Stickiness/ plasticity	Roots	Pores	Boun-dary	Effer-vescence
A1	sg[2]	l/l	so/po	--	--	CS	NE
A2	pl,3,th	sh/vfr	ss/sp	2vf	3vf,1f,i	CW	NE
Bt1	abk,3,m-co	mh/fi	s/p	3f,m	1f,m,t	GS	NE
Bt2	m	h/fi	s/p	2f,m	1f,m,t	--	NE

	Sand fraction[1] %					CEC cmol(+) kg^{-1}	pH
	Very coarse	Coarse	Medium	Fine	Very fine		
A1	4.9	5.5	5.0	6.4	6.8	10.5	7.2
A2	3.3	3.8	3.7	5.1	7.2	13.4	7.7
Bt1	2.2	3.6	3.3	4.3	4.8	20.0	6.3
Bt2	3.5	5.4	5.2	5.1	3.9	14.1	6.1

	%C[3]	%N[3]	P[1] (mg kg^{-1})	Clay films
A1	1.89	0.16	21.51	--
A2	1.23	0.06	5.27	--
Bt1	0.87	0.11	1.31	m,d,7.5YR 5/6,clf,brf,pf
Bt2	0.23	0.01	0.39	m,d,7.5YR 4/6,clf,brf,apf

[1] - Analyzed in the University of Nevada Las Vegas (UNLV) Soil Laboratory

[2]- See Schoeneberger et al. (2002) for abbreviation definitions

[3]- Analyzed at the UNLV Chemistry Department

Table 4. Soil characteristics of the Ts pedon, Troughspring series, Spring Mountains, Nevada. Data from NRCS (2006b; 2006d); Soil Survey Staff (2006) unless otherwise noted.

Horizon	Depth (cm)	Color moist	Color dry	Sand	Silt	Clay	USDA texture class
				--------	%	--------	
Oi	0-5	10YR 2/1	10YR 3/2	--	--	--	--
A	5-23	10YR 3/2	10YR 4/2	26.6	49.8	23.6	silt loam
Btk	23-36	10YR 4/3	10YR 5/3	23.8	50.5	25.7	silt loam
Bk1	36-61	10YR 4/3	10YR 6/3	23.3	52.5	24.2	silt loam
2Bk2	61-74	10YR 3/2	10YR 8/2	21.7	54.8	23.5	silt loam
2Bkm	74-160	10YR 3/2	10YR 8/2	28.6	50.3	21.1	--

	Structure	Consistence Dry/ moist	Stickiness/ plasticity	Roots	Pores	Boun-dary	Effer-vescence
Oi	--	vs/vfr[1]	so/po	--	--	CS	--
A	sbk,3,m	mh/fr	ss/ps	2vf-f	2f,t, 2vf-f,i	CW	NE
Btk	abk,3,m	mh/fr	s/p	2vf-co	2vf-f,i,t	CW	ST
Bk1	sbk,2,m-co	mh/fr	s/p	2vf-f,co	2vf,i, 2f,t	CW	VE
2Bk2	sbk,1,m	h/vfi	b	--	--	AI	VE
2Bkm	m	eh/efi	b	--	--	--	VE

	Sand fraction % Very coarse	Coarse	Medium	Fine	Very fine	CEC cmol(+) kg^{-1}	pH
Oi	--	--	--	--	--	--	--
A	3.2	3.1	2.6	4.3	13.4	27.5	7.6
Btk	3.2	3.8	3.2	3.6	10.0	25.7	8.0
Bk1	3.9	4.5	3.8	4.5	6.6	19.7	7.9
2Bk2	2.8	4.6	4.2	4.6	5.5	15.9	8.0
2Bkm	5.3	6.9	6.1	5.8	4.5	9.6	8.1

	%C	%N	P[2] (mg kg^{-1})	Clay films
Oi	38.86	1.671	--	--
A	3.08	0.171	69.71	--
Btk	3.68	0.144	38.16	vf,f,10YR 4/4 (m),clf,apf
Bk1	6.75	0.191	8.02	--
2Bk2	8.82	0.117	3.41	--
2Bkm	10.27	0.083	--	--

[1]- See Schoeneberger et al. (2002) for abbreviation definitions

[2]- Analyzed in the University of Nevada Las Vegas Soil Laboratory

Net increases in moisture content with depth were measured in all three soils of the Charkiln series and in the soil of the Troughspring series (Tables 1-4). The pH of the four soils ranged from slight acidity to moderate alkalinity (Brady and Weil, 2002) with pedon C3 having the lowest pH (6.1) in the Bt2 horizon. Pedon C2 had the highest pH (8.2) in the ABt, Bt1, Bt2, and Bt3 horizons (Soil Survey Staff, 2006).

Rock fragments from the Bt1, Bt2, and Bt3 horizons of the C1 pedon, and the Bt1 and Bt2 horizons of the C3 pedon were described in hand sample for internal and external weathering characteristics (rinds, fractures, and plucking). Rock fragments consisted predominantly of quartzite. The C1 pedon had very few rock fragments, and had the smallest rock fragments (up to 1 ½ cm by 1 cm in size). Rock fragments from the Bt1 and Bt2 horizons were angular, whereas fragments in the Bt3 horizon were angular to subangular. Patchy to continuous, brown (7.5YR 5/3-7.5YR 5/4), exterior staining was common on fragments in the Bt1 and Bt2 horizons, but only patchy, reddish-yellow exterior staining (7.5YR 6/6) was present in the Bt3 horizon. The exterior staining on a rock fragment from the Bt2 horizon was analyzed using the scanning electron microscope (SEM). Results of the SEM analysis indicated the presence of silica (70%), aluminum (20%), iron (5%), potassium (4%), and calcium (1%). In all three horizons, external fractures were not evident in hand samples, and no evidence of plucked grains could be seen as the grain sizes of the fragments were too small to see with the unaided eye. However, a rock fragment from the Bt2 horizon contained a 0.5 cm by 0.4 cm indentation with white secondary aluminum and calcium silicate coating, as determined from SEM analysis (silica 62%; aluminum 16%; calcium 15%; iron 5%; and potassium 2%). A cavity 1 cm by 1 cm in size and lined with strong brown stain (7.5YR 5/6) was found along a freshly cut surface in one of the rock fragments from the Bt3 horizon. Rocks had smooth exteriors and most lacked rinds, however, a rock fragment from the Bt1 horizon had a very thin rind with a stain that extended 1.5 cm into the rock.

Rock fragments in the C3 pedon had patchy to continuous, brown to reddish-yellow (7.5YR 5/4-7.5YR 6/6) exterior staining. The fragments were mostly angular with very few subangular clasts. No rinds were present on these fragments, however there was an occasional fracture. Fractures were easily identified only if the rocks were previously cut open or on surfaces that did not have staining. Some fractures were filled with stain. Evidence of a few plucked quartz grains were visible and the voids already were filled with stain, however, grain sizes of most of the rock fragments in the C3 pedon were too small to see with the unaided eye. Most of the fragments have smooth exteriors, with a few having rough, mottled exteriors. The C3 pedon had the largest rock fragments overall. Rock fragments in the Bt1 horizon ranged in size up to 4 cm by 2.5 cm. The Bt2 horizon had the largest rock fragments of the two pedons. The two largest rock fragments were 5 cm by 1 ½ cm, and 3 cm by 2 cm in size.

Thin sections

Thin sections provide a method of observing soil constituents in their natural, undisturbed state (Cady et al., 1986). The purpose of thin-section examination in this study was to search for evidence of illuvial clay (Southard and Southard, 1985), and general mineral identification and condition (Cady et al., 1986). Argillans along grains and voids indicate translocation of clay, whereas the lack of argillans indicates clay formed *in situ* (Bilzi and Ciolkosz, 1977; Southard and Southard, 1985 Ciolkosz et al., 1995). Differences in argillan characteristics, such as color, can indicate that they formed in separate climates (Catt, 1989). Thin-section analysis of minerals can aid in determining source of parent material, the presence of lithologic discontinuities, and degree of weathering in the soil (Cady et al., 1986).

Thin sections were made from clods collected from the A1, AB, Bt1, Bt2, and Bt3 horizons from the C1 pedon, and the A, Btk, Bk1, 2Bk2, and 2Bkm horizons from the Ts pedon by Spectrum Petrographics, Inc., Vancouver, Washington. In general, the mineralogy in the A1, AB, Bt1, and Bt2 horizons from the C1 pedon, consisted predominantly of fine- to coarse-grained quartz, with lesser amounts of quartzite fragments, feldspar, garnet, chlorite or epidote, chert, biotite, muscovite, opaques, shale, organic material, a yellow to amber mineral that may be garnet or microcrystalline quartz stained with iron oxide, a red mineral that could be garnet or iron-oxide stained quartz, and iron-oxide staining between grains (Figures A.1-A.6). The AB horizon, however, did not have shale. The type of garnet found in the soils of the study area likely is almandine, the iron-rich end member of the garnet series based on SEM analysis (Graham et al., 1989a; Klein and Hurlbut, 1993). The Bt1 and Bt2 horizons had microcline grains showing tartan twinning. The microcline grains were fresh, that is, the outlines of the grains were prominent (Figure A.3). The Bt3 horizon contained predominantly fine- to coarse-grained quartz, with lesser amounts of quartzite fragments, feldspar, chlorite or epidote, chert, biotite, opaques, a red mineral that could be garnet or iron-oxide stained quartz, and iron-oxide staining between grains. The sample from the A horizon has several impurities that could be a source for clays (Figure A.1). In all of the horizons, argillans were nonexistent to continuous on all sides of grains. About two-thirds of all grains, most of which were quartz, were coated with argillans (Figures A.1-A.2). Clay filled or coated a few voids in all horizons. Clay filled some fractures in all but the Bt3 horizon (Figure A.6). Argillans were most common in the A and AB horizons, and started decreasing in the Bt1 horizon and below. Clays in the Bt1, Bt2, and Bt3 horizons are infilling between grains or growing out from grains rather than coating the grains. While most minerals were grain supported, a matrix of microcrystalline quartz and clay was beginning to form around grains in some areas. Minerals were grain supported in the A and AB horizons (Figures A.1-A.2), whereas they were matrix supported in the Bt1, Bt2, and Bt3 horizons (Figures A.3-A.6). The amount of matrix around grains increased with depth.

Thin sections made by the NRCS for the Bt1 horizon from the C2 pedon indicate the presence of oriented clays on sand grains, a relict clay film, and possibly a channel argillan (Figures A.7-A.9). The oriented clays appear to be pressure faces rather than argillans, they extend outward from the grains, and occur in the matrix around the grains (Rebecca Burt, NRCS, written commun., 2006).

Thin sections made from clods collected from the Ts pedon indicate the presence of the following minerals. The A horizon predominantly consists of fine- to coarse-grained quartz, with lesser amounts of calcite, feldspar, biotite, muscovite, epidote or chlorite, opaques, organic material, an amber-colored mineral that may be garnet or microcrystalline quartz stained with iron oxide, a red mineral that could be garnet or iron-oxide stained quartz, and a matrix consisting of calcite with quartz grains (Figure A.10). The Btk and Bk1 horizons mostly contain quartz with lesser amounts of feldspar, calcite, biotite, muscovite, epidote or chlorite, opaques, chert, fossil remnants of mollusks and coral, organic material, iron oxide stain, and a matrix consisting of calcite cement (Figures A.11-A.12). The 2Bk2 and 2Bkm horizons predominantly consist of calcite with lesser amounts of quartz, biotite, opaques, organic material, a reddish-brown mineral that could be garnet or iron-oxide stained quartz, and fossil remnants of mollusks and coral (Figures A.13-A.14). The 2Bkm horizon also contained chert and an amber-colored mineral that may be garnet or microcrystalline quartz stained with iron oxide. The Bk1 horizon marks the point at which calcite starts breaking apart grains as shown in thin section (Figure A.12). The 2Bk2 and 2Bkm horizons exhibited the best example of calcite consuming other minerals; quartzite grains are being broken and pushed apart by the calcite and grains are matrix supported by calcite cement (Gile et al., 1965; Reheis, 1988). Argillans are nonexistent to continuous on grains in the A and Btk horizons, and nonexistent to partial in the Bk1 horizon. The argillans are mostly on quartz grains and a few opaques and voids (Figure A.10). No argillans were evident in the 2Bk2 and 2Bkm horizons. A clay matrix surrounds the grains in the A and Btk horizons, whereas a calcite cement is present in the Bk1, 2Bk2, and 2Bkm horizons.

Rock Fragments

Thin sections of rock fragments from the Bt1, Bt2, and Bt3 horizons of the C1 pedon, and the Bt1 and Bt2 horizons from the C3 pedon were analyzed for mineral content, and point counts were done on rock fragments to quantify mineralogy. The dominant rock fragment type from all horizons in the C1 and C3 pedons was quartzite. Point count data for the soils indicate that quartz is the dominant mineral in the soils of the Charkiln series, and quartz and calcite are dominant minerals in the soil of the Troughspring series (Table 5; Soil Survey Staff, 2006).

Six rock fragments from the Bt1 horizon of the C1 pedon underwent mineralogical analysis. Two of the rock fragments contained grains too small to point count, and another rock fragment was too fractured to point count accurately. Of the three remaining rock fragments, grain sizes were very fine to very coarse sand (Udden, 1898; Wentworth, 1922), and were angular to rounded. The mineralogy was predominantly quartz (61-89%) with lesser amounts of garnet (≤11%), feldspar (≤3%), chlorite or epidote (≤3%), muscovite (≤1%), opaques (≤1%), chert (≤1%), biotite (trace), (some minerals were not detected in point counts, but were identified during petrographic microscope analysis and are listed as being present in trace amounts) and an unidentified yellowish-gold mineral that could be garnet, epidote, or microcrystalline quartz stained with iron oxide (≤7%), and an occasional chert matrix (≤18%; Table 5; Figures A.15-A.17). All of the rock fragments have a brown iron-oxide staining (7.5YR 4/3) around the fragment that is visible on the thin section with the unaided eye, and an iron-oxide coating around edges of grains and on some fractures that is visible microscopically (Munsell colors were obtained where possible such as the outside of rock fragments, but not for the iron stain that was visible microscopically. A Munsell color was obtained for one iron-oxide rind visible in thin section, which is actually the outside of the rock fragment, but it was difficult to see with the unaided eye and therefore is an estimate). The iron-oxide stain is believed to be due to ferri-argillans around the edges of the grains. The stain ranged from a patchy coating on and around some grains to a continuous coating throughout the rocks. Minerals ranged from grain supported to matrix supported. Fracturing is nonexistent to continuous throughout the rock fragments.

Three rock fragments from the Bt2 horizon in the C1 pedon consisted of quartzite. All samples were matrix supported and mostly contained grains too small to be point counted. The remaining visible grains were as large as fine sand (Udden, 1898; Wentworth, 1922), and were angular to rounded. Grain composition consisted of quartz with lesser amounts of chert, muscovite, opaques, garnet, a chert matrix, and an iron-oxide stain (Table 5; Figure A.18). The iron-oxide stain ranged from patchy to continuous throughout the rocks. One rock fragment contained a fossil remnant of a brachiopod and iron-oxide stain (Figure A.19).

Seven rock fragments from the Bt3 horizon in the C1 pedon were analyzed; two contained grain sizes too small for point counts. For the remaining fragments, grain sizes ranged from very fine to medium sand (Udden, 1898; Wentworth, 1922), and were angular to rounded. The composition consisted mostly of quartz (77-89%) with lesser amounts of chert (≤23%), garnet (≤3%), feldspar (≤2%), opaques (≤2%), chlorite or epidote (≤2%), muscovite (≤1%), calcite (<1%), and an iron-oxide stain (Table 5; Figure A.20). Minerals ranged from grain supported to matrix supported. In the matrix-supported rocks, the matrix consisted of chert or amorphous silica (≤6%). Some rock fragments were fractured with minimal to prominent iron-oxide stain along the fractures. Other rocks contained iron-oxide stain along a few grain boundaries or on an occasional void; grain-supported rocks had little iron-oxide stain (Figure A.21).

Three rock fragments from the Bt1 horizon in the C3 pedon were analyzed for mineralogy (Figures A.22-A.24). The composition of the rocks consisted mostly of quartz (85%) with lesser amounts of chert (11%), garnet (1%), muscovite (1%), feldspar (1%), opaques (<1%), and an iron-oxide staining around some grains and fractures. The stain

Table 5. Thin-section point count data for the Charkiln and Troughspring series pedons, Spring Mountains, Nevada.

Site information			Grain counts (%)								
Pedon	Depth (cm)	Horizon	Quartz	Feldspar (undifferentiated)	Potassium feldspar	Garnet	Opaques	Chlorite or epidote (green)	Hornblende	Chert	Muscovite
C1	24-36	Bt1	86	3		1	trace	3		trace	1
			89	<1		<1	1	1		1	1
			61	2		11	1			trace	trace
	118-166	Bt3	88	2		1	<1	2		<1	1
			77							23	
			89	<1		3	2				
			88	2		1	1	<1		3	1
C2[1]	23-36	Bt1	55		22	trace	3		3	3	4
	36-117	Bt2	54		25	trace	3		3	3	4
C3	12-50	Bt1	85		1	1	<1			11	1
	50-110	Bt2	91	3		1	trace			3	trace
			71			4	trace			6	
			81	1		8				2	1
Ts[1]	23-36	Btk	46		22	trace	1		2	4	4
	36-61	Bk1	36		11	trace	2		3	1	1

Table 5. Thin-section point count data for the Charkiln and Troughspring series pedons, Spring Mountains, Nevada (continued).

Site information			Grain counts (%)							
Pedon	Depth (cm)	Horizon	Biotite	Epidote or garnet (yellow)	Pyroxene	Weatherable aggregates	Beryl	Iron oxides (goethite)	Plagioclase feldspar	Glass
C1	24-36	Bt1								
			trace	7						
	118-166	Bt3								
C2[1]	23-36	Bt1	7		2	trace	trace	trace	trace	trace
	36-117	Bt2	5		2	trace	trace	trace	trace	
C3	12-50	Bt1								
	50-110	Bt2	1							
			trace							
Ts[1]	23-36	Btk	2		3	trace		trace	trace	1
	36-61	Bk1	1		1		trace		trace	trace

Table 5. Thin-section point count data for the Charkiln and Troughspring series pedons, Spring Mountains, Nevada (continued).

Site information			Grain counts (%)						
Pedon	Depth (cm)	Horizon	Monazite	Tourmaline	Zircon	Rutile	Calcite	Carbonate aggregates	Matrix
C1	24-36	Bt1							6
									7
									18
	118-166	Bt3							6
									6
							<1		4
C2[1]	23-36	Bt1	trace	trace	trace				
	36-117	Bt2	trace	trace	trace	trace	trace		
C3	12-50	Bt1							1
	50-110	Bt2							1
									19
									7
Ts[1]	23-36	Btk	trace	trace	trace		13	1	
	36-61	Bk1	trace	trace	trace	trace	39	6	

[1] Data from Soil Survey Staff, 2006.

ranged from patchy to continuous throughout the rocks, and minerals were mostly grain supported. Two of the rocks were fractured throughout, and some fractures were filled with quartz.

Rock fragments from the Bt2 horizon in the C3 pedon consisted of quartzite. Grain sizes ranged from very fine sand to gravel (Udden, 1898; Wentworth, 1922), angular to well rounded, and grain supported to matrix-supported (Figures A.25-A.29). In general, mineral composition of the rock fragments was predominantly quartz (71-91%) with lesser amounts of garnet (≤8%), chert (≤6%), feldspar (≤3%), muscovite (≤1%), biotite (≤1%), chlorite or epidote (trace), opaques (trace), a clay-rich, quartz or amorphous silica matrix (≤19%), and an iron-oxide staining. The iron-oxide stain was found along the edges of grains and in voids, and ranged from patchy to continuous throughout the rocks. Some grains were fractured and one rock contained quartz veins occurring at 90-degree angles (Figure A.26).

Grain count of the sand-silt mineralogy from the Bt1 and Bt2 horizons of the C2 pedon indicate quartz content ranged from 54% to 55% with lesser amounts of feldspar (≤25%), biotite (≤7%), muscovite (4%), opaques (3%), chert (3%), hornblende (3%), and pyroxene (2%). Garnet, weatherable aggregates, beryl, iron oxides (goethite), plagioclase feldspar, glass, monazite, tourmaline, zircon, rutile, and calcite were present in trace amounts in the C2 pedon (Table 5; Soil Survey Staff, 2006). Mineral analysis of the Btk and Bkm horizons from the soil of the Troughspring series indicates quartz (36-46%) and calcite (13-39%) were the dominant minerals with lesser amounts of feldspar (≤22%), carbonate aggregates (≤6%), chert (≤4%), muscovite (≤4%), opaques (≤3%), hornblende (≤3%), biotite (≤2%), pyroxene (≤3%), and glass (≤1%). Garnet, weatherable aggregates, beryl, iron oxides (goethite), plagioclase feldspar, monazite, tourmaline, zircon, and rutile were present in trace amounts in the Ts pedon (Table 5; Soil Survey Staff, 2006).

Chemical analyses

Net decreases of phosphorous occurred with depth in the C1, C3, and Ts pedons (Tables 1, 3, and 4); the C2 pedon was not analyzed for phosphorous. Phosphorous was highest in the A and Btk horizons of the Ts pedon at 69.71 mg kg^{-1} and 38.16 mg kg^{-1}, respectively, but decreased to 3.41 mg kg^{-1} in the 2Bk2 horizon. Phosphorous ranged from 27.68 mg kg^{-1} in the A horizon to 13.58 mg kg^{-1} in the Bt3 horizon of the C1 pedon, and 21.51 mg kg^{-1} in the A1 horizon to 0.39 mg kg^{-1} in the Bt2 horizon of the C3 pedon. Carbon was highest in the Ts pedon with values ranging from 38.86% in the Oi horizon to 3.08% in the A horizon (Table 4; Soil Survey Staff, 2006). These values are up to 61% higher than the highest carbon values in similar horizons in the soils of the Charkiln series. Carbon ranged from 3.03% in the Oi horizon to 0.25% in the Bt3 horizon in the C1 pedon; 17.10% in the Oi horizon to 0.30% in the Bt3 horizon in the C2 pedon; and from 1.89% to 0.23% in the A1 and Bt2 horizons, respectively, in the C3 pedon (Tables 1-3). Net decreases in nitrogen also occurred with depth in all four pedons (Tables 1-4). Nitrogen was 7-39% higher in the Oi and A horizons in the Ts pedon than in similar horizons in the soils of the Charkiln series. Nitrogen in the Ts pedon ranged from 1.671% in the Oi horizon to 0.083% in the 2Bkm horizon (Soil Survey Staff, 2006). Nitrogen in the soils of the Charkiln series ranged from 0.06% to 0.18% in the A horizons and from 0.01% to 0.08% in the B horizons.

Calcium (Ca^{2+}) and potassium (K^+) were the most common base cations in the soils of the Charkiln and Troughspring series followed by (in order of decreasing abundance) Mg^{2+}, Mn^{2+}, Na^+, and Fe^{2+} (Table 6). Sodium was present in small amounts in the C3 pedon, and in the A and Btk horizons of the Ts pedon, but was not present in the C2 pedon. Iron was detected in the AB and A horizons of the C1 and C3 pedons, respectively. Manganese was present in all horizons of the C1 and C3 pedons except for the Bt1 horizon in the C3 pedon. Pedons C2 and Ts were not analyzed for Fe^{2+} and Mn^{2+}.

Cation exchange capacity (CEC) increased with depth in the soils of the Charkiln series, and was highest in the Bt1 horizons (Tables 1-4). This increase in CEC corresponds to increases in clay. Cation exchange capacity values in the soil of the Troughspring series, however, decreased with depth and correspond to a decrease in clay and an increase in carbonate (Soil Survey Staff, 2006). The cation exchange capacity values were higher in the A and Btk horizons in the soil of the Troughspring series than in any of the soils of the Charkiln series. The A horizon in the Ts pedon is deeper (23 cm) than the A horizons in the pedons of the Charkiln series (12-13 cm). The Btk horizon in the Ts pedon had a clay content of 18-27%, which was similar to clay contents in the Bt horizons (20-35%) of the type location soil (C2 pedon), but the CEC values in the C2 pedon were lower than those in the upper horizons of the Ts pedon (Soil Survey Staff, 2006).

X-ray diffraction results for the C1 and C3 pedons indicate that quartz was the dominant mineral, followed by kaolinite and mica. The least abundant minerals were vermiculite, gibbsite, and mixed layer chlorite and vermiculite (Table 7). In the C2 pedon, mica and kaolinite were the most dominant followed by vermiculite and quartz.

Montmorillonite, mica, and calcite were dominant minerals in the Ts pedon, with lesser amounts of kaolinite, vermiculite, and quartz (Soil Survey Staff, 2006).

Results of the dithionite-citrate (CD) extraction indicate a greater percentage of CD extractable Mn was present in the C1 and C3 pedons with lesser amounts of Fe, whereas a greater percentage of Fe was present in the C2 and Ts pedons with lesser amounts of Al and Mn (Table 8; Soil Survey Staff, 2006). Net decreases in CD extractable manganese occurred with depth in the C1, C2, and C3 pedons, whereas CD extractable Fe remained steady in the C1 pedon but increased slightly in the C2 and C3 pedons. In the C2 pedon, CD extractable Al remained constant in all horizons, and in the Ts pedon CD extractable Fe and Al decreased with depth, and CD extractable Mn was detected in only trace amounts in the A, Btk, Bk1, and 2Bk2 horizons (Soil Survey Staff, 2006). Free Fe oxides are interpreted to be noncrystalline or crystalline forms (Weisenborn and Schaetzl, 2005), and a measure of the total pedogenic Fe, such as goethite, hematite, lepidocrocite, or ferrihydrite in the soil (Burt, 2004). The presence of Mn detected using the dithionite-citrate method is considered the easily reducible form (Burt, 2004), because extensive substitution of Mn^{2+} and Mn^{3+} for Mn^{4+} occurs (Taylor et al., 1983).

Table 6. Base cations in the Charkiln and Troughspring series pedons, Spring Mountains, Nevada. Data from Soil Survey Staff (2006) unless otherwise noted.

Site information			NH$_4$OAC extractable bases [cmol(+) kg^{-1}]						
Pedon	Depth (cm)	Horizon	Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}	Fe^{2+}	Mn^{2+}	Sum of bases
C1[1]	0-3	Oi	--	--	--	--	--	--	--
	3-12	A	8.5	2.0	nd	15.2	nd	0.8	25.7
	12-24	AB	9.5	2.2	0.1	12.4	0.1	1.7	24.3
	24-36	Bt1	12.3	3.3	0.2	13.6	nd	4.2	29.4
	36-118	Bt2	11.8	3.3	0.1	8.1	nd	3.0	23.3
	118-166	Bt3	13.9	3.3	0.2	5.5	nd	2.5	23.0
C2	0-3	Oi	--	--	--	--	--	--	
	3-13	A	9.1	1.8	nd	1.1	--	--	12.0
	13-23	ABt	8.9	1.9	nd	0.9	--	--	11.7
	23-36	Bt1	10.9	2.6	nd	0.7	--	--	14.2
	36-117	Bt2	10.8	2.7	nd	0.4	--	--	13.9
	117-165	Bt3	14.2	2.6	nd	0.3	--	--	17.1
C3[1]	0-3	A1	12.2	1.4	0.2	8.6	0.1	1.3	22.4
	3-12	A2	12.6	3.1	0.2	10.8	nd	0.3	26.7
	12-50	Bt1	16.0	2.4	0.3	11.2	nd	nd	30.0
	50-110	Bt2	14.9	1.4	0.2	3.6	nd	1.2	20.1
Ts	0-5	Oi	--	--	--	--	--	--	--
	5-23	A	51.4	2.1	0.1	0.8	--	--	54.4
	23-36	Btk	62.7	2.0	trace	0.4	--	--	65.1
	36-61	Bk1	53.2	1.4	nd	0.2	--	--	54.8
	61-74	2Bk2	55.2	1.2	nd	0.2	--	--	56.6
	74-160	2Bkm	51.3	0.8	nd	0.1	--	--	52.2

--, Data not analyzed

nd, Not detected

[1]-Data analyzed in the University of Nevada Las Vegas Soils Laboratory

Table 7. XRD analysis of minerals in the Charkiln and Troughspring series pedons, Spring Mountains, Nevada [XXX, highest peak; XX, medium peak; X, small peak. Size of XRD peak indicates relative quantity of mineral[1]].

Site information			Mineral							
Pedon	Depth (cm)	Horizon	Quartz	Mica	Kaolinite	Montmorillonite	Vermiculite	Calcite	Gibbsite	Mixed layer chlorite and vermiculite
C1[1]	24-36	Bt1	XXX	XX	XX				X	X
	36-118	Bt2	XXX	XX	XX				X	X
	118-166	Bt3	XXX	XX	XX				X	X
C2[2]	23-36	Bt1	X	XXX	XX		X			
	36-117	Bt2	X	XXX	XX		XX			
C3[1]	12-50	Bt1	XXX	XX	XX				X	X
	50-110	Bt2	XXX	XX	XX				X	X
Ts[2]	23-36	Btk	X	XX	X	XXX	X	X		
	36-61	Bk1	X	XX	XX	XXX	X	XX		

[1] See Figures A.30-A.34 in the Appendix for selected XRD graphs.
[2] Data from Soil Survey Staff, 2006.

Table 8. Dithionite-citrate extraction results for the Charkiln and Troughspring series pedons, Spring Mountains, Nevada.

Site information			%		
Pedon	Depth (cm)	Horizon	Fe	Al	Mn
C1	3-12	A	0.2	--	1.1
	12-24	AB	0.2	--	1.0
	24-36	Bt1	0.2	--	0.9
	36-118	Bt2	0.2	--	0.9
	118-166	Bt3	0.2	--	0.8
C2[1]	3-13	A	1.3	0.1	0.1
	13-23	ABt	1.3	0.1	trace
	23-36	Bt1	1.4	0.1	trace
	36-117	Bt2	1.4	0.1	trace
	117-165	Bt3	1.4	0.1	trace
C3	0-3	A1	0.1	--	0.6
	3-12	A2	0.2	--	0.7
	12-50	Bt1	0.2	--	0.4
	50-110	Bt2	0.2	--	0.5
Ts[1]	5-23	A	1.1	0.2	trace
	23-36	Btk	1.0	0.1	trace
	36-61	Bk1	0.6	0.1	trace
	61-74	2Bk2	0.2	trace	trace
	74-160	2Bkm	0.1	trace	--

-- Data not analyzed.

[1] Data from Soil Survey Staff, 2006.

CHAPTER 4

INTERPRETATION

The hypotheses tested in this study were 1) the clay mineralogy for the argillic horizons is not forming from the quartzite but from upslope contributions from limestone parent material and eolian dust, and 2) the argillic horizons are currently forming rather than relict features. The results of this study indicate that quartzite, limestone, and minor amounts of eolian dust have provided source material for the formation of the soils of the Charkiln series. This conclusion is based on minerals and fossils found in the soils that can be linked to specific geologic units within the study area. The presence of channel argillans in the uppermost B horizons suggests that the argillans are actively forming and that the current climate has sufficient moisture to translocate clays. Active illuviation in the soils is further supported by clay-depleted A horizons and B horizons showing increases in clay. The current climate supports a pinyon-juniper woodland which may have enhanced clay illuviation through organic-acid weathering. Additionally, the current climate has a low leaching rate because clays have not been leached out of the soil, and minerals are not highly weathered. Consistency in pedogenic iron, base cations, texture, sand sizes, and color indicates the lack of a lithologic discontinuity. The lack of rock fragment weathering and rinds indicates the fragments have undergone little change in their environment. The lack of weathering, change in soil environment, or lithologic discontinuity suggest that the soils of the Charkiln series are currently forming.

CHAPTER 5

DISCUSSION

Data presented in this study indicate that parent material and climate are the most important factors controlling formation of argillic horizons in the soils of the Charkiln series. Biota, topography, and time have had more subtle effects on the development of the soils. The soils of the Charkiln series contain argillic horizons that are forming in alluvium dominated by quartzite lithology (NRCS, 2006b; 2006d); quartzite is an atypical parent material for soils with argillic horizons (Ogg and Baker, 1999; Ogg et al., 2001). Argillic horizons form from parent material consisting of sedimentary, igneous, and metamorphic rocks (Gile and Grossman, 1968; Smeck et al., 1968; Smith and Buol, 1968; Aandahl, 1982; Chadwick et al., 1975; Nettleton et al., 1975; Franzmeier et al., 1985; McDaniel and Nielsen, 1985; Southard and Southard, 1985; Goenadi and Tan, 1989; Graham and Buol, 1990; Stolt and Rabenhorst, 1991; Hopkins and Franzen, 2003), however, not all rocks contain sufficient material for clay and argillic-horizon formation. Argillic horizons that formed in soils in arid regions, such as the southwest, are considered relict features of a past climate (Ruhe, 1965; Nettleton et al., 2000) because the current climate does not have sufficient moisture to translocate clays (Gile and Grossman, 1968; Gile and Hawley, 1968; Nettleton et al., 1989). Contrary to current belief, the soils of the Charkiln series are forming in an aridic to ustic soil moisture regime, with a mesic temperature regime (NRCS, 2006b; 2006c; 2006d), as indicated by the presence of actively forming channel argillans in the upper B horizons (Figure A.9; Buol and Hole, 1961). The channel argillans indicate that the current climate has sufficient moisture to translocate clays. Additionally, the A horizons in the soils of the Charkiln series have been depleted of clay and the B horizons show increases in clay which further support active illuviation in the soils (Tables 1-3; Graham and Wood, 1991).

Genesis of the argillic horizons

Parent Material / Topography

While soils of the Charkiln series formed in alluvium dominated by quartzite lithology (NRCS, 2006b; 2006d), quartzite parent material is not the single source of clay for argillic-horizon formation; similar results were found by Ogg and Baker (1999) and Ogg et al. (2001) who conducted research on soils with argillic horizons that formed on alluvial fans containing quartzite rocks. Ogg et al. (2001) found up to 60% clay in the soils, and concluded that this much clay was unusual for soils containing only quartzite parent material. Clay in the argillic horizons was derived from alluvial fans consisting of sandstone and phyllite (Ogg and Baker, 1999). Phyllite contains muscovite and chlorite (Klein and Hurlbut, 1993), which are sources of clay mica and vermiculite, respectively (Allen and Hajek, 1989). These same minerals also were found in the soils of the Charkiln series. A phyllitic shale is located in the northwest Spring Mountains (Zj, Johnnie Formation; not shown on Figure 2; Page et al., 2005), however, it is located outside of the study area. Changes in geomorphology, erosion, or eolian sources may have led to deposition of material from this geologic unit within the study area contributing to soil formation. Clay contents in the Charkiln series are not as high as those in the studies by Ogg and Baker (1999) and Ogg et al. (2001; Tables 1-3), however, specific quartzite and limestone units have been identified as sources of material in the soils of the Charkiln series based on mineral assemblages and fossils found in the soils. These units include the Stirling Quartzite (Zs; Figure 2; Page et al., 2005), which contains mineral assemblages of the greenschist facies (quartz, chlorite, epidote, and muscovite; Klein and Hurlbut, 1993) that were also found in rock fragments and clods collected from soils of the Charkiln series; and the limestone units include the Bird Spring Formation, undivided (PMb), Monte Cristo Group (Mm), Mississippian and Devonian rocks, undivided (MDu), and the Pogonip Group (O€p; Figure 2; Page et al., 2005), which contain brachiopods that were occasionally found in rock fragments collected from the Bt2 horizon of the C1 pedon (Figure A.19).

The presence of quartz and mica, the dominant minerals in the soils of the Charkiln series (Table 7; Soil Survey Staff, 2006), further support the conclusion that the quartzite and limestone units listed above provided the parent material for the development of argillic horizons in the soils of the Charkiln series; quartzite and limestone are major sources of quartz and secondary clay mica, respectively (Johnson, 1970; Allen and Hajek, 1989). Other secondary minerals found in rock fragments and clods suggest clays are locally derived. For example, kaolinite is a prominent mineral in the soils of the Charkiln series, followed by lesser amounts of vermiculite, gibbsite, and mixed layer chlorite and vermiculite (Table 7; Soil Survey Staff, 2006). Kaolinite, gibbsite, and vermiculite are weathering products of biotite and/or feldspar (Johnson, 1970; Allen and Hajek, 1989; Graham et al., 1989a; Ogg and Baker, 1999). Thin-section analysis of rock fragments

(Table5) and clods indicates very low amounts of biotite, further suggesting additional sources of gibbsite are from the weathering of almandine and plagioclase (Graham et al., 1989b) while vermiculite is from chlorite alteration (Allen and Hajek, 1989).

The hypothesis that source material for the argillic horizons is locally derived is further supported by an analysis of the geomorphology of the study area, and a soil series adjacent to the Charkiln series. A dominant source of clay mica in the soils of the Charkiln series likely came from erosion of limestone (Allen and Hajek, 1989; Ciolkosz et al., 1995) upslope of the Charkiln series. The limestone parent material from which the soil of the Troughspring series formed from is a likely source. The soil of the Troughspring series, a nearby paleustoll located upslope of the Charkiln series, has calcic (Btk and Bk1) and petrocalcic (2Bk2 and 2Bkm) horizons extending to depths of 61 and 160 cm, respectively (Table 4; NRCS, 2006b; 2006d). These horizons were not found in the soils of the Charkiln series that were analyzed up to depths of 166 cm (Tables 1-3). It is hypothesized that the limestone downslope of the Troughspring series has been eroded away. This process of erosion created an intramontane basin that has partially filled with alluvial material (Peterson, 1981), and formed the fans on which the soils of the Charkiln series developed (NRCS, 2006b; 2006d). Additionally, soils of the Troughspring Series formed in alluvium derived predominantly from limestone (NRCS, 2006b; 2006d), including several of the same units that also provided source material for the formation of the Charkiln series [Bird Spring Formation, undivided (PMb), Monte Cristo Group (Mm), Mississippian and Devonian rocks, undivided (MDu), the Pogonip Group (O€p), and the Stirling Quartzite (Zs; Page et al., 2005)]. These sources also were identified through fossils and mineralogy found in clods collected from the Ts pedon.

Montmorillonite and mica, the dominant minerals in the soil of the Troughspring series (Table 7; Soil Survey Staff, 2006), likely originated from the limestone parent material and weathering of phyllosilicates found in the soils (Allen and Hajek, 1989), which further supports that soils in the study area formed from locally derived material. Topography also has influenced argillic-horizon formation in the Spring Mountains. According to Alexander (1995), argillic horizons are less likely to occur on steep slopes because soil formation is hindered due to increased erosion. Comparison of the three soils of the Charkiln series shows that the soil on the steepest slope (C3 pedon) has the thinnest argillic horizons of all three soils (Tables 1-3).

While evidence for argillic-horizon development is strongly supported from local parent material, external sources of atmospheric dust also may have had an influence on soil formation. Dust accumulation rates measured by Reheis and Kihl (1995) for southern Nevada and southeastern California range from 4.3 to 15.7 g m^{-2} y^{-1}. Eolian dust provided sediment that contributed to the pedogenesis of soils on the eastern slopes of Charleston Peak (Reheis et al., 1992; Reheis and Kihl, 1995), and the formation of alluvial and colluvial deposits within the study area on the western slopes of the mountain range (Qay, Qcf, Qaiy; Figure 2; Page et al., 2005). Kaolinite and calcium or calcareous dust in soils often originate from eolian processes (Gile et al., 1966; Gardner, 1972; Gile, 1975; Gile et al., 1981; Machette, 1985; Allen and Hajek, 1989; Reheis et al., 1992) suggesting that eolian sediment could be an additional source of the kaolinite present in the soils of the Charkiln and Troughspring series, and the calcium in the calcic and petrocalcic horizons of the Troughspring series. Calcite was not detected during XRD analysis of minerals in the soils of the Charkiln series (Table 7), but was found on the surface of a rock fragment collected from the C1 pedon. The lack of calcite in the soils of the Charkiln series suggests that the calcite from the influx of atmospheric dust could be dissolving and reprecipitating as other minerals in the soil, particularly in those horizons with acidic pH (Faure, 1991). Alternatively, the lack of calcite may indicate that atmospheric dust was not a significant source of material for soil formation on the west side of the Spring Mountains. Silt, the main constituent in eolian dust within southern Nevada and California (Reheis and Kihl, 1995), is not the dominant particle size in the soils of the Charkiln series, which further indicates eolian dust is not a significant source of material in the soils (Ciolkosz et al., 1995).

Climate

The current climate of the study area ranges from an aridic to ustic soil moisture regime, with a mesic temperature regime, and a mean annual precipitation rate of 38 to 41 centimeters (NRCS, 2006b; 2006d). The current climate has sufficient moisture to translocate clays as indicated by actively forming channel argillans in the upper B horizons of the C1 and C2 pedons (Buol and Hole, 1961). In the C1 and C2 pedons, argillans ranged from nonexistent to continuous on all sides of grains (Figures A.1-A.2 and A.7-A.8), and filled or coated a few voids (Figures A.5 and A.9). The lack of uniformly distributed argillans in the soils indicate that they formed from percolating water rather than as residual products originally present in the soils (Frei and Cline, 1949), and further support the conclusion that the argillic horizons are currently forming. In the C1 pedon, clay filled some fractures in all horizons except the Bt3 horizon (Figure A.6). Argillans along voids and in the Bt horizons in both pedons indicate that the clays originated through illuviation and translocation (Smith and Buol, 1968; Southard and

Southard, 1975; Alexander and Nettleton, 1977; Bilzi and Ciolkosz, 1977; Ciolkosz et al., 1995). Clay in the lower Bt horizons in both pedons was found to be infilling between grains rather than coating the grains (Rebecca Burt, NRCS, written commun., 2006); the lack of argillans in the Bt horizons indicates the clays also formed *in situ* (Southard and Southard, 1985). The argillans along grains and voids were thin, and brownish in color which suggests that they all formed within the same climate (Catt, 1989). However, mineral analysis of the argillans should be done to confirm this.

Further evidence of current climatic influences on argillic-horizon development is found when examining clay mineralogy. Based on a study of clay development in soils in a climate with wet and dry seasons by Sherman (1952), the mean annual precipitation rate at the soils of the Charkiln and Troughspring series is conducive of montmorillonite formation, a dominant mineral in the soil of the Troughspring series. However, the mean annual precipitation rate of the current climate is insufficient for kaolinite development (Sherman, 1952), which also is a prominent mineral in the soils of the Charkiln and Troughspring series. An additional factor that may have affected clay accumulation is infiltration and percolation of snowmelt. The mean annual snow-water equivalent for snowpack on the west side of the Spring Mountains ranges from 14.0 to 22.1 centimeters (NRCS, 2006a). Surface runoff can lead to depletion of clay on the soil surface (Graham and Wood, 1991), but infiltration and percolation of snowmelt runoff could be contributing to argillic-horizon formation in the soils of the Charkiln series.

Past research indicates that alpine glaciation and freeze/thaw cycles can affect argillic-horizon formation (Brady and Weil, 2002; Buol and Hole, 1961; Nettleton et al., 1969). These cycles must be taken into consideration because of the high elevation of the study area (~2,100 m), and current climate data from the Mt. Charleston Lodge weather station that indicates temperatures in the study area are low enough to freeze soils (data accessed December 3, 2006 at http://www.instaweather.com/KLAS/default.asp?cid=90&id=MTCHS). Additionally, alpine glaciation is suspected to have occurred on the east side of the Spring Mountains (Van Hoesen and Orndorff, 2000; 2001), and geologic units containing possible glacial or periglacial deposits (Qcf) are located on the east side of Charleston Peak (Figure 1; Page et al., 2005). Glacial activity can compact soils affecting the translocation of clays (Brady and Weil, 2002), and freeze/thaw cycles can prevent argillan formation or destroy them completely (Buol and Hole, 1961; Nettleton et al., 1969). However, studies by Ruhe (1975) and Waltman et al. (1990) show that glacial erosion and freeze-thaw cycles can have little to no affect on soils and argillic horizons, respectively. The presence of the argillic horizons and argillans in the soils of the Charkiln series indicate that freeze/thaw cycles had little to no affect on the argillic horizons, and glaciation likely was sporadic or limited to the east side of the mountain range, and not all soils were affected. Van Hoesen and Orndorff (2000; 2001) provide no information on whether the soils on the west side of the Spring Mountains were affected by the glaciation that occurred in the Pleistocene during the last Ice Age. These data suggest that the soils of the Charkiln series either were not affected by alpine glaciation and freeze/thaw cycles, or that the soils formed in the Holocene after glaciation occurred, and further supports that the soils are young.

Vegetation

Vegetation also has been found to be an important factor in the formation of argillic horizons (Aandahl, 1982). Clay illuviation occurs under Coulter pine (*Pinus coulteri* B.

Don) due to the absence of earthworms which tend to mix the soil and counteract illuviation (Graham and Wood, 1991; Johnson-Maynard et al., 2004). Coniferous needle litter contains the water-soluble organic substance, phenol (Blaschke, 1979), which can cause clay dispersion as shown in Huon pine (*Dacrydium franklinii*; Bloomfield, 1957), and dissolved organic matter leached from Douglas-fir roots contains organic acids that promote clay dispersion (Durgin and Chaney, 1984), which then could be translocated. At the Charkiln series study sites on the west side of Charleston Peak, a subhumid continental climate exists with sufficient moisture to support a pinyon (*Pinus monophylla* Torr. & Frém.)-juniper [*Juniperus osteosperma* (Torr.)] woodland with mountain mahogany (*Cercocarpus ledifolius* Nutt. var. *intermontanus* N. Holmgren) and Gambel oak (*Quercus gambelii* Nutt.; Houghton et al., 1975). Two of the pedons of the Charkiln series (C1 and C2) have a pinyon (*Pinus monophylla* Torr. & Frém.)-juniper [*Juniperus osteosperma* (Torr.)] and Gambel oak (*Quercus gambelii* Nutt) association (NRCS, 2006b; 2006d), and the third site (C3 pedon) is in a burned area that was pinyon (*Pinus monophylla* Torr. & Frém.)-juniper [*Juniperus osteosperma* (Torr.)] and Gambel oak (*Quercus gambelii* Nutt) prior to the burn. Pine likely affected clay dispersion and accumulation (Bloomfield, 1957; Durgin and Chaney, 1984; Graham and Wood, 1991; Johnson-Maynard et al., 2004) in the soils of the Charkiln series, but since vegetation and soil characteristics were similar at all three pedons, no conclusion can be made as to whether vegetation enhanced the rate at which the argillic horizons formed. Vegetation likely had the same effect on soil formation because all three pedons had similar clay percentages and depths to argillic horizons (Tables 1-3).

Time

The presence of the argillic horizons in the soils of the Charkiln series indicate that the landscape on which the soils have formed has been stable for a long enough period of time for the horizons to develop (Gile and Hawley, 1968). Argillic horizons typically require at least a few thousand years to form (Gile and Grossman, 1968; Gile and Hawley, 1968; Smeck et al., 1968; Gile, 1975; Nettleton et al., 1975; Parsons and Herriman, 1976; Bilzi and Ciolkosz, 1977; Franzmeier et al., 1985; Southard and Southard, 1985; Ciolkosz et al., 1989; Goenadi and Tan, 1989; Nettleton et al., 1989; Stolt and Rabenhorst, 1991; Hopkins and Franzen, 2003), but under unique circumstances they can form within as little as 50 years (Graham and Wood, 1991; Johnson-Maynard et al., 2004). Landform characteristics within the study area were analyzed to determine landscape stability as well as a relative age of the soils of the Charkiln series (Vincent et al., 1994; Christenson and Purcell, 1985). The fans on which the soils of the Charkiln series have formed likely are young to intermediate in age (<15,000 to 700,000 years old; Figure 5; Christenson and Purcell, 1985; Page et al., 2005), and agree with past research that state argillic horizons typically form within a few thousand years.

Relict or currently forming argillic horizons

To determine whether a soil is relict or currently forming, past research focused on soil characteristics such as rock-fragment weathering (Peltier, 1949; Mills, 1988; Soil Survey Staff, 1998), mineral composition, characteristics, and weathering (Allen and Hajek, 1989; Graham et al., 1989b; Shaw et al., 2003), soil color (Torrent et al., 1980), and lithologic discontinuities (Soil Survey Staff, 1998). Quartzite rock fragments in the Charkiln C1 pedon ranged from angular in the upper horizons to subangular in lower horizons, but remained angular with depth in the Charkiln C3 pedon. This indicates that the rock fragments have consistently undergone little weathering, and therefore a lithologic discontinuity does not exist within the soils of the Charkiln series. The lack of rinds in rock fragments throughout the soil profiles also indicates the fragments have undergone little change in their environment due to movement of the fragments, weathering processes, or climate change (Peltier, 1949; Mills, 1988). Relatively constant levels of Fe, Al, and base cations (Tables 6 and 8) and consistency in texture, color, and sand sizes were measured with depth (Tables 1-3), also indicating the lack of a discontinuity in the soils. The lack of weathering, change in soil environment, or lithologic discontinuity suggest that the soils of the Charkiln series are currently forming.

While some weathering has occurred across the three pedons as indicated by increases in CEC and clay with depth (Carroll, 1959; Tables 1-3), in general, the soils are still considered pedologically young. CEC in young soils can range in value because CEC is based on the type of clays within the soil; for example kaolinite has a lower CEC value than montmorillonite (Grim, 1953; Carroll, 1959). Regardless of the type of clays in the soils of the Charkiln series, there is a trend of increasing CEC and clay with depth in all three soils. Free Fe oxides remained relatively constant with depth in the soils of the Charkiln series which also suggests that little weathering has occurred, and thus the soils are young (Shaw et al., 2003); an increase in free Fe oxides indicates an increase in weathering (Burt, 2004). The consistency of free Fe oxides with depth among the three Charkiln pedons suggests that the degree of weathering, or lack thereof, has been similar among horizons in the soils of the Charkiln series (Table 8). The slight difference in iron content in the C2 pedon as compared to the C1 and C3 pedons likely is due to topography since the other soil-forming factors (parent material, climate, vegetation, and time; Jenny, 1994), have been the same for all three pedons. The C2 pedon is located higher in elevation than the C1 pedon, lower in elevation than the C3 pedon, and within a depression where runoff occurs and could be a collection point for minerals.

The presence of fresh microcline grains and unstable grains, such as chlorite and biotite, in the soils of the Charkiln series also indicate that the soils have undergone little weathering or change in weathering processes (Peltier, 1949; Allen and Fanning, 1983; McDaniel and Nielsen, 1985; Allen and Hajek, 1989; Graham et al., 1989b). Thin-section analysis of a clod from the Bt1 horizon of the Charkiln C1 pedon revealed a microcline grain with a prominent outline and tartan twinning (Figure A.3), and a fresh grain resembling chlorite (Figure A.4). Chlorite and biotite are unstable minerals; chlorite is typically found in relatively unaltered subsoils, and soils that have undergone reduced weathering due to climate or because the soil is young (Allen and Fanning, 1983; Allen and Hajek, 1989). Because the soils have undergone little weathering (Graham et al., 1989b) and little change in weathering processes (Peltier, 1949; McDaniel and Nielsen, 1985) suggests that the soils are young and currently forming in the present climate.

The color of a soil, specifically red hues, can indicate the relative age of a soil as redder hues increase with soil age and time (Hurst, 1977; Torrent et al., 1980; Dohrenwend et al., 1991). The most common iron compounds that are responsible for soil color are goethite, hematite, lepidocrocite, and maghemite (Hurst, 1977; Allen and Hajek, 1989); hematite is a highly effective red pigmenting agent in soils (Graham and

Buol, 1990). Soils of the Charkiln series contain yellow-red hues (7.5YR to 10YR) likely due to the presence of organic matter, goethite, and quartz (Hurst, 1977; Allen and Hajek, 1989; Graham et al., 1989b; Graham and Buol, 1990). The lack of red hues in the soils of the Charkiln series indicates that the soils have little to no hematite (Schwertmann, 1993; Ciolkosz et al., 1995) and therefore, are young (Torrent et al., 1980). The B horizons likely will redden with time because almandine, which weathers to goethite and hematite, was found in rock fragments and clods collected from the soils (Hurst, 1977; Schaetzl and Mokma, 1988; Graham et al., 1989a; Graham and Buol, 1990; Ciolkosz et al., 1995). Failure to detect goethite and hematite by XRD analysis (Table 7) however, confirms low iron oxide content, as suggested above, poor crystallinity, or that the iron oxides are present as very fine particles (Gager, 1968; Karpachevskiy et al., 1972; Gangas et al., 1973; Torrent et al., 1980).

Goenadi and Tan (1989) state that argillans in an argillic horizon, high amounts of kaolinite, and iron-oxide minerals are evidence that an intense weathering and leaching process has occurred in a soil. Silt-to-clay ratios in the Charkiln pedons decreased with depth, indicating that secondary minerals are dominant in the lower horizons (Ray, 1963, Anjos et al., 1998), and CEC values increased with depth, indicating some weathering has occurred (Carroll, 1959) in all three Charkiln pedons. However, as stated above, the lack of goethite and hematite, and the presence of fresh and unstable grains indicate that the soils of the Charkiln series have not undergone intense weathering (Peltier, 1949; Allen and Fanning, 1983; McDaniel and Nielsen, 1985; Allen and Hajek, 1989; Graham et al., 1989a; Graham et al., 1989b). Therefore, the dominant process of soil formation in the soils of the Charkiln series likely is translocation of clay rather than *in situ* weathering of minerals, because more weathered argillic horizons are an indication that some of the clay formed *in situ* (Smith and Buol, 1968). The presence of argillans on grains and pores in the C1 and C2 pedons indicate clay translocation (Southard and Southard, 1985; Bilzi and Ciolkosz, 1977, Ciolkosz et al., 1995).

CHAPTER 6

CONCLUSIONS

The soils of the Charkiln series formed on alluvium derived predominantly from a quartzite parent material, as well as from a limestone parent material and minor inputs of eolian dust. The dominant rock fragment lithology in the soils of the Charkiln series is quartzite, the dominant mineral is quartz, and the most prominent clays are mica and kaolinite. Minerals and fossils found in the soils of the Charkiln series were linked to specific parent materials within the study area that contributed to the formation of the soils. These units include the Stirling Quartzite, the Bird Spring Formation, undivided, Monte Cristo Group, Mississippian and Devonian rocks, undivided, and the Pogonip Group.

The argillic horizons in the soils of the Charkiln series are forming in the current climate regime which has sufficient moisture to translocate clays, as indicated by actively forming channel argillans in the uppermost B horizons (Buol and Hole, 1961). Active illuviation is further supported by A horizons depleted of clay and B horizons showing increases in clay (Graham and Wood, 1991). However, the climate has a low degree of weathering, as indicated by consistency in free Fe oxides with depth, and the presence of fresh microcline grains and unstable minerals such as biotite and chlorite. The lack of weathering indicates the soils of the Charkiln series are young.

Similarities in clay types, percentages, and depths to argillic horizons indicate that singleleaf pinyon pine (*Pinus monophylla* Torr. & Frém.), the dominant vegetation within the Charkiln series, did not affect argillic-horizon formation differently in any of the three soils, but may have enhanced argillic-horizon formation.

Soils of the Charkiln series have formed on fans that are young to intermediate in age, and range from less than 15,000 to 700,000 years old (Christenson and Purcell, 1985; Page et al., 2005). The age of the surfaces on which the soils formed and the presence of the argillic horizons indicate landscape stability (Gile and Hawley, 1968). The type of landform on which the soils have formed has not specifically affected soil formation, but slope has affected argillic-horizon formation. The soil of the Charkiln series with the steepest slope (the C3 pedon) has the thinnest argillic horizons of all three soils.

Consistency in rock fragment components and characteristics, relatively constant levels of Fe, Al, and base cations, and the lack of any abrupt changes in texture, sand sizes, and color with depth suggest that the soils of the Charkiln series are currently forming features.

Organic matter content, and the presence of goethite and quartz are responsible for the range in color found in the soils. The lack of red hues in the soils of the Charkiln series indicates that the soils are young.

APPENDIX

THIN-SECTION PHOTOGRAPHS AND XRD GRAPHS

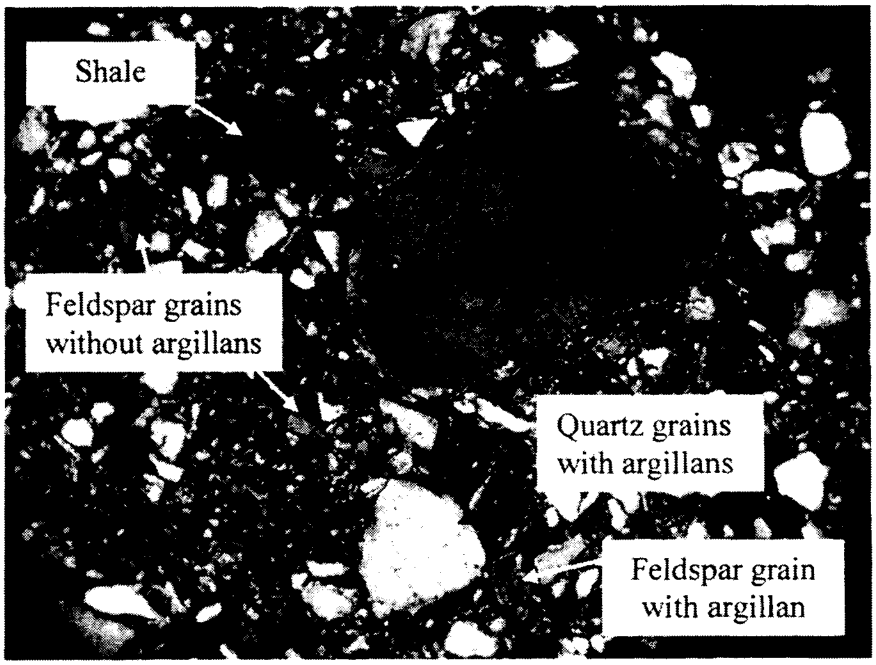


Figure A.1. Thin section of a clod collected from the A horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

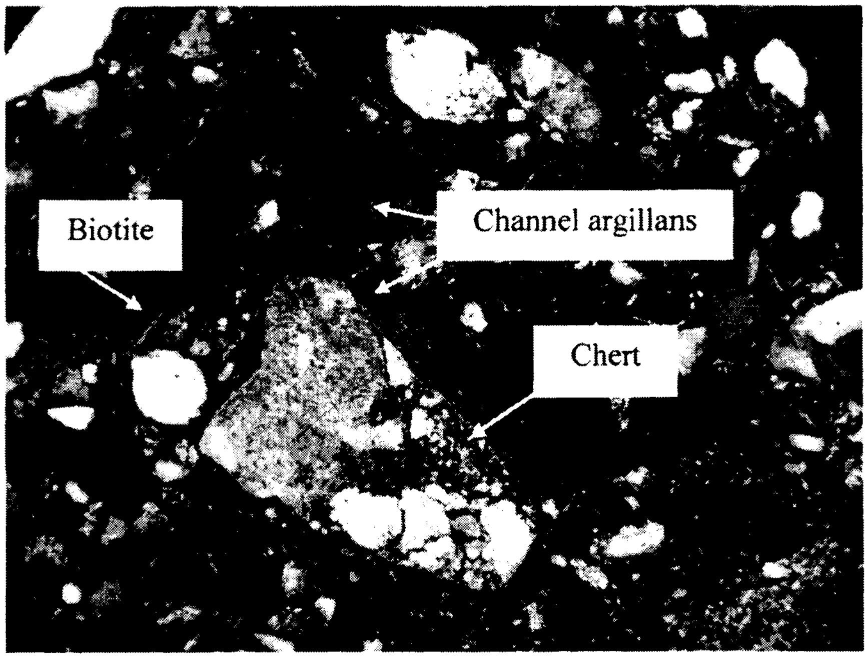


Figure A.2. Thin section of a clod collected from the AB horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

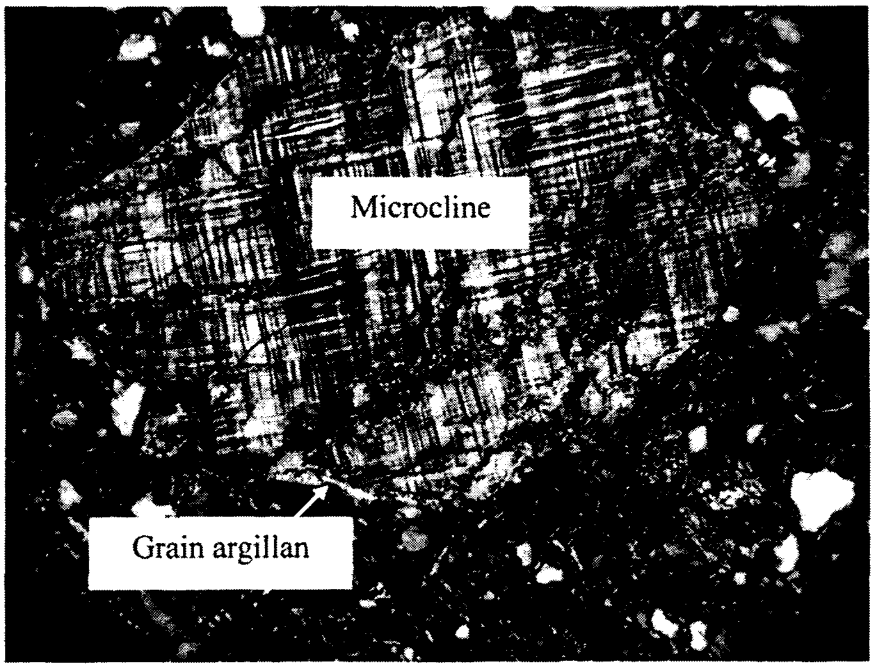


Figure A.3. Thin section of a clod collected from the Bt1 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

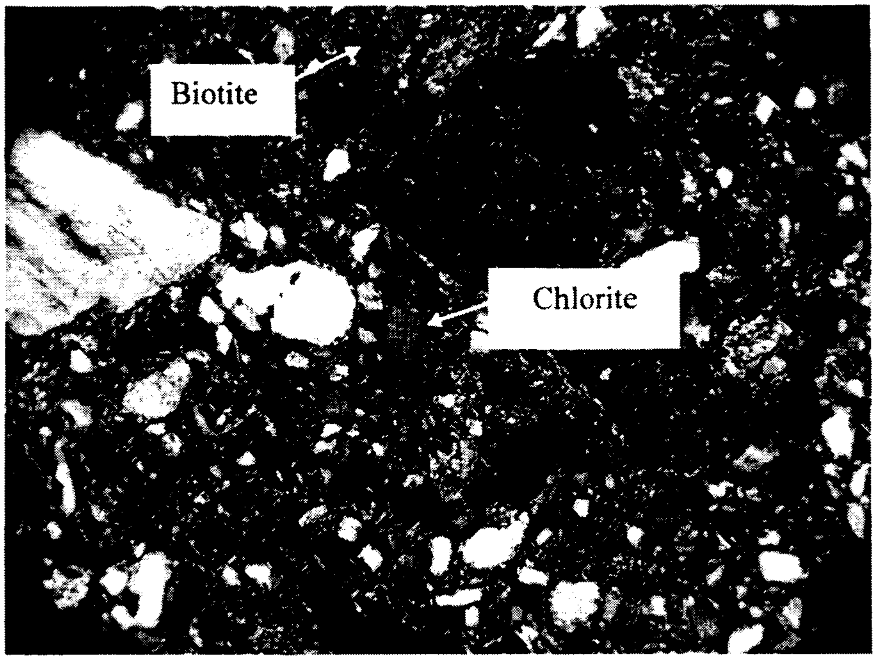


Figure A.4. Thin section of a clod collected from the Bt1 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

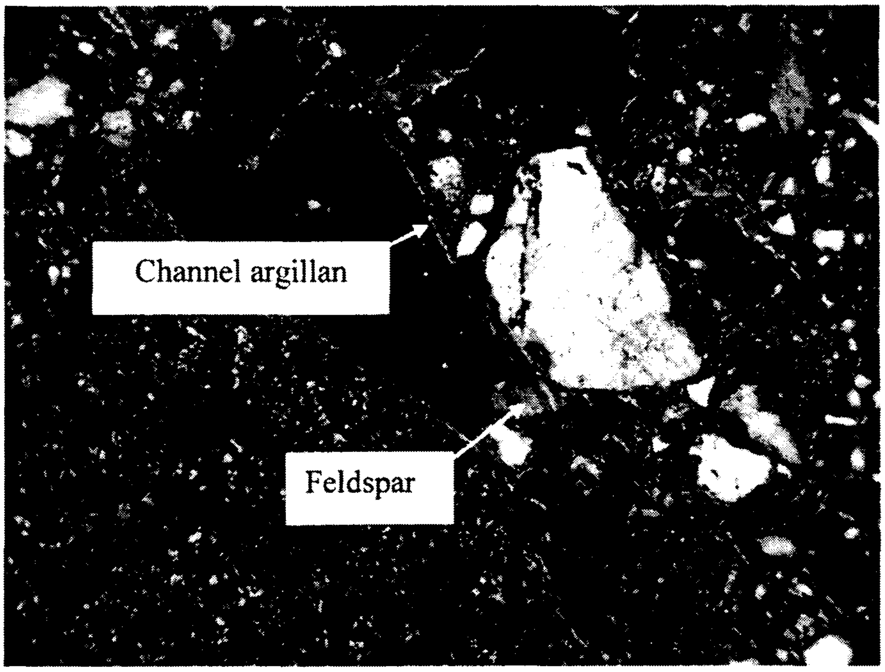


Figure A.5. Thin section of a clod collected from the Bt2 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

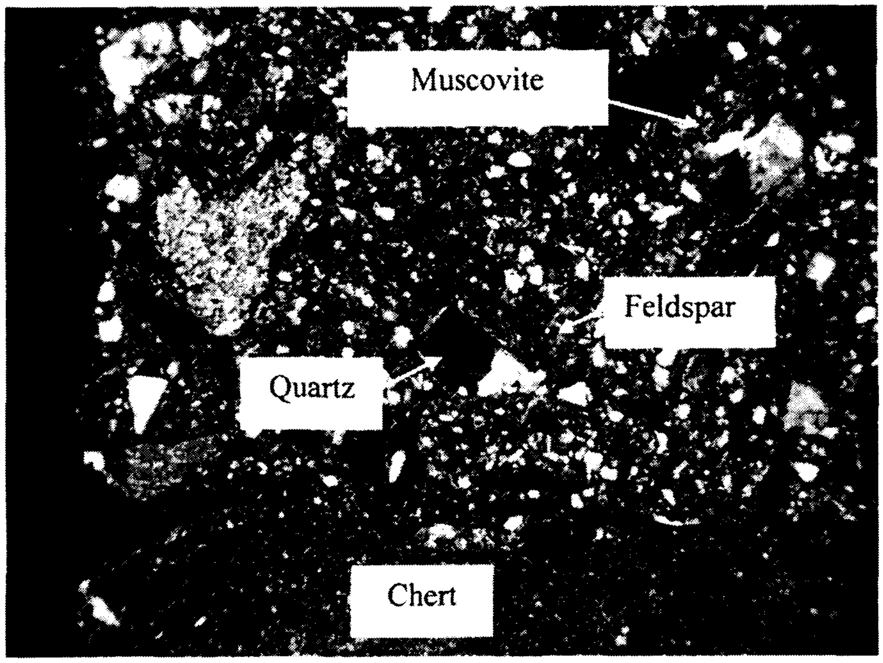


Figure A.6. Thin section of a clod collected from the Bt3 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

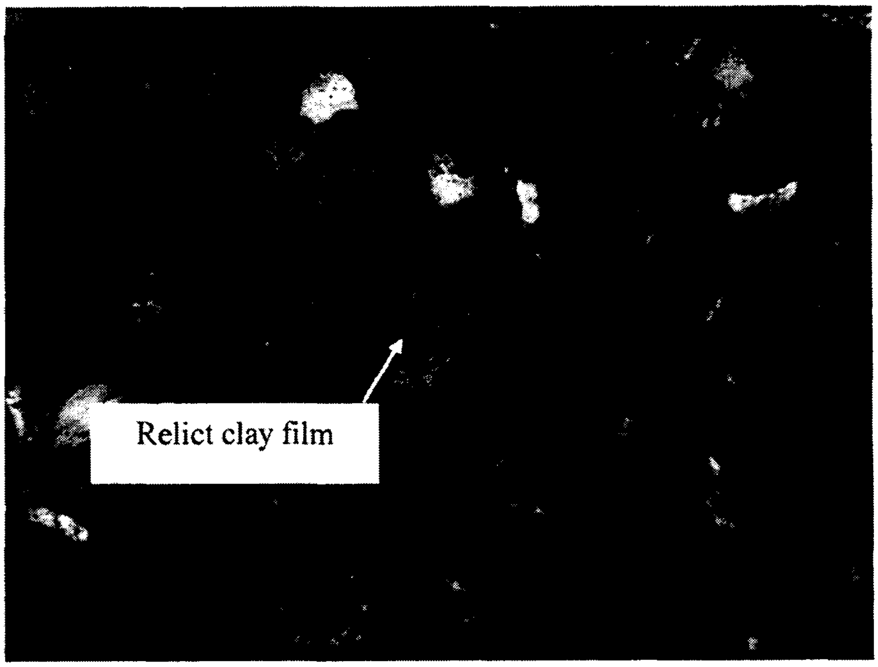


Figure A.7. Thin section of coarse silt (0.02-0.05 mm) from the Bt1 horizon in the C2 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 20X, cross polarized light. Photograph taken by the NRCS.

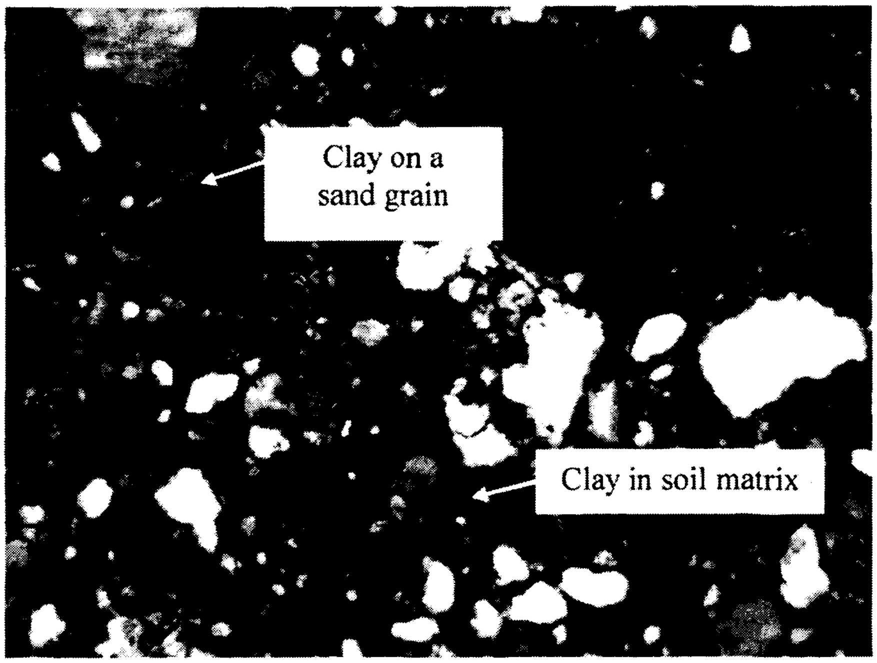


Figure A.8. Thin section of coarse silt (0.02-0.05 mm) from the Bt1 horizon in the C2 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 10X, cross polarized light. Photo taken by the NRCS.

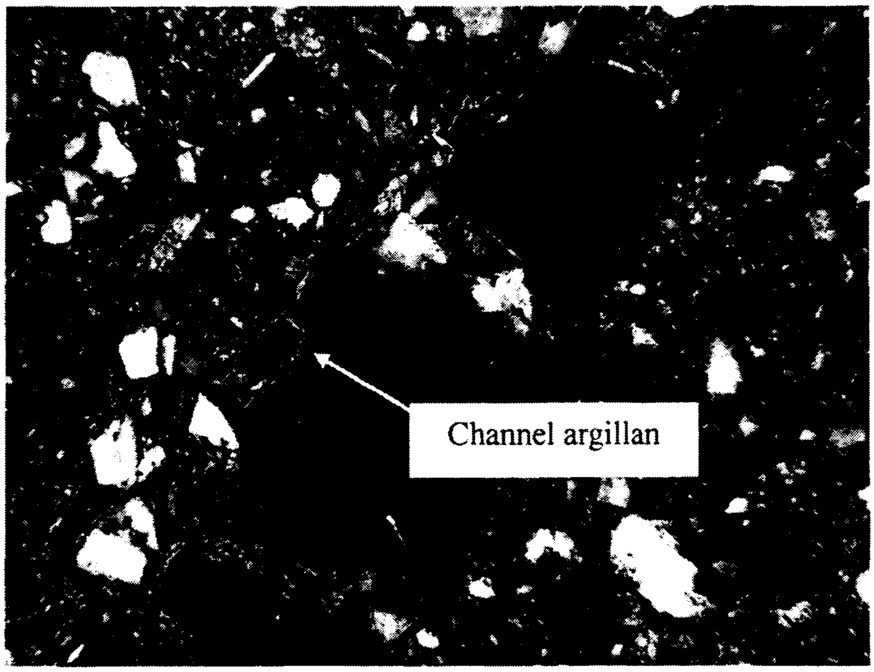


Figure A.9. Thin section of coarse silt (0.02-0.05 mm) from the Bt1 horizon in the C2 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 10X, cross polarized light. Photograph taken by the NRCS.

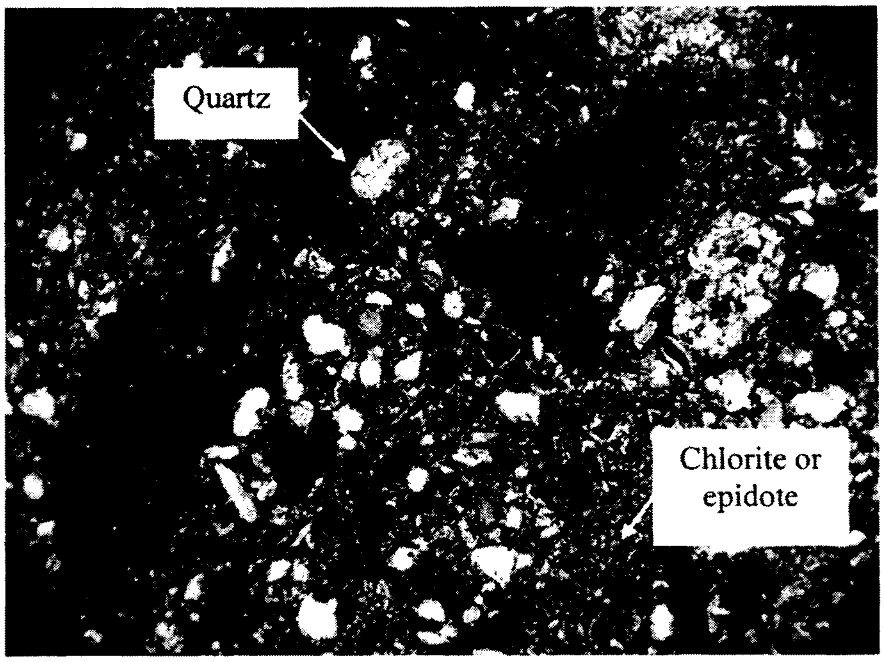


Figure A.10. Thin section of a clod from the A horizon in the Ts pedon, Troughspring series, Spring Mountains, Clark County, Nevada. Magnification 100X, cross polarized light.

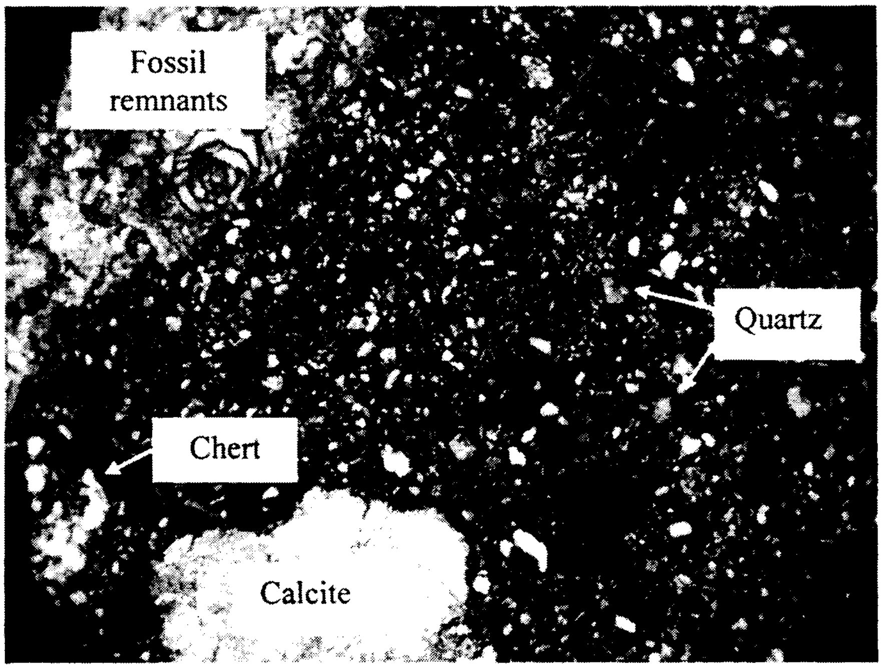


Figure A.11. Thin section of a clod collected from the Btk horizon in the Ts pedon, Troughspring series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

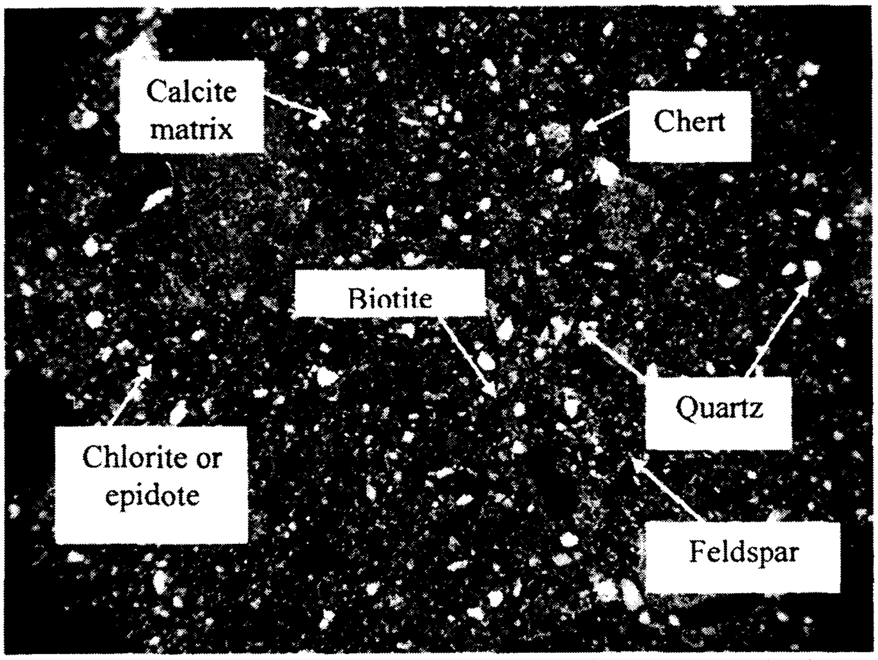


Figure A.12. Thin section of a clod collected from the Bk1 horizon in the Ts pedon, Troughspring series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

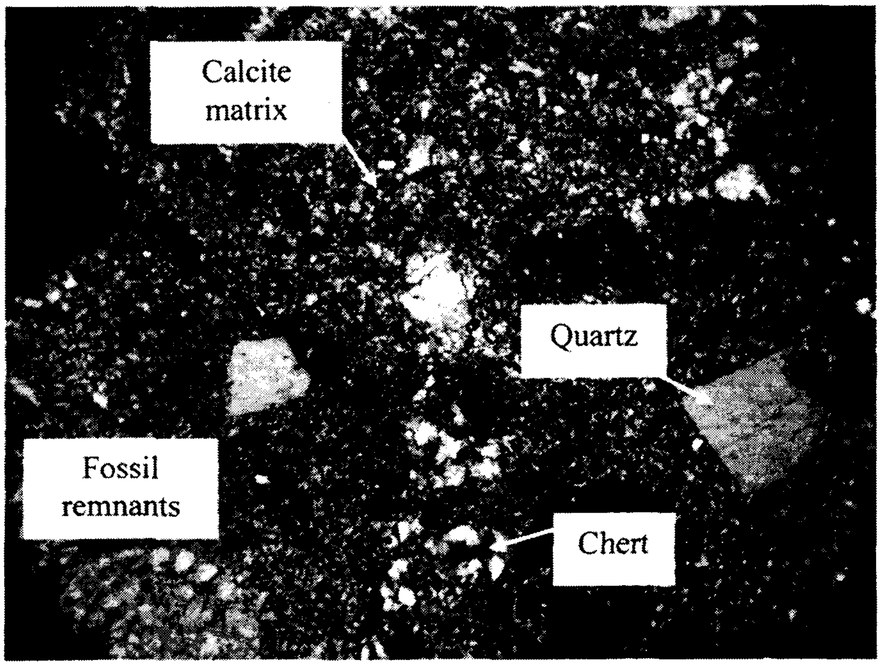


Figure A.13. Thin section of a clod collected from the 2Bk2 horizon in the Ts pedon, Troughspring series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

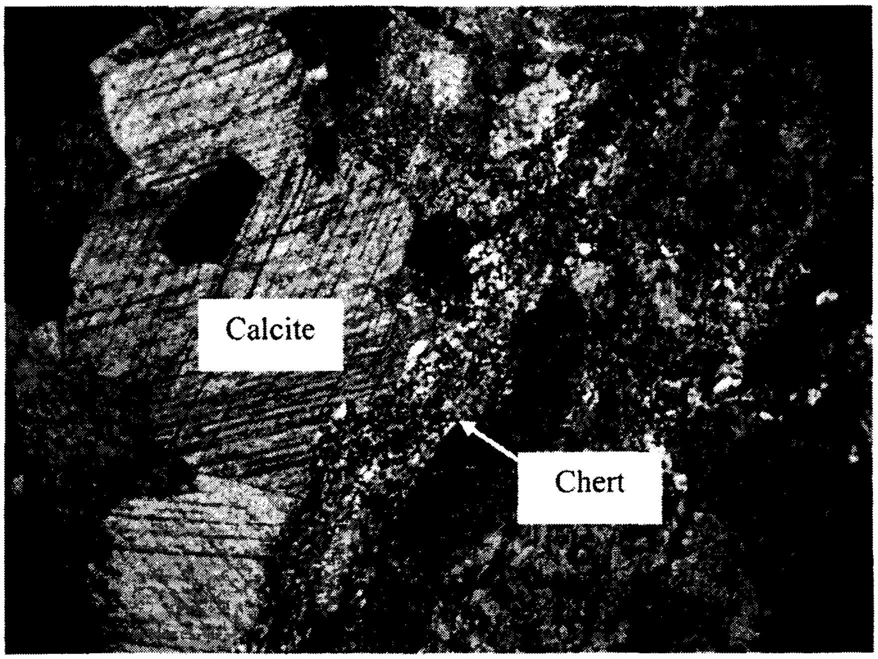


Figure A.14. Thin section of a clod collected from the 2Bkm horizon in the Ts pedon, Troughspring series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

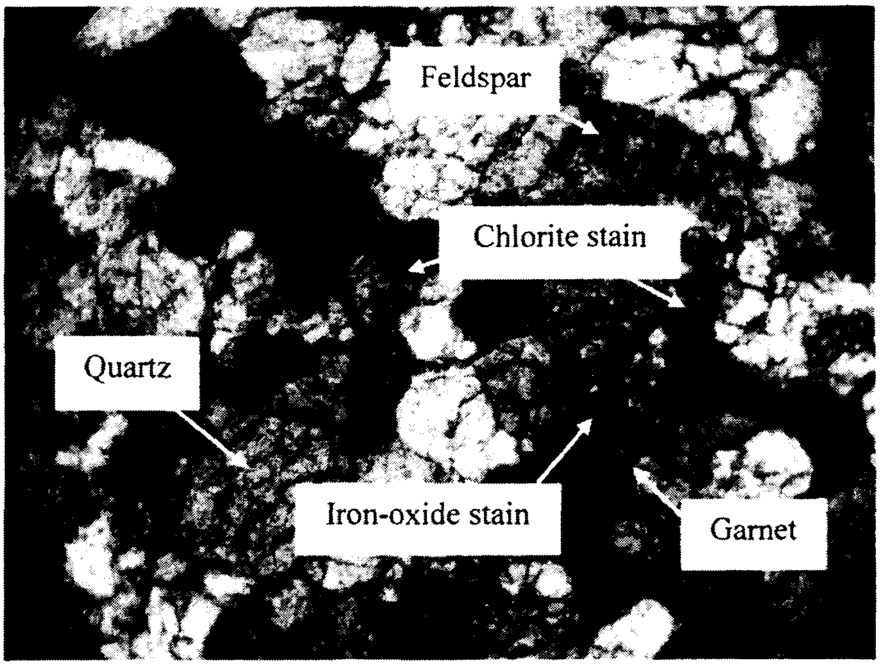


Figure A.15. Thin section of a rock fragment collected from the Bt1 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 100X, cross polarized light.

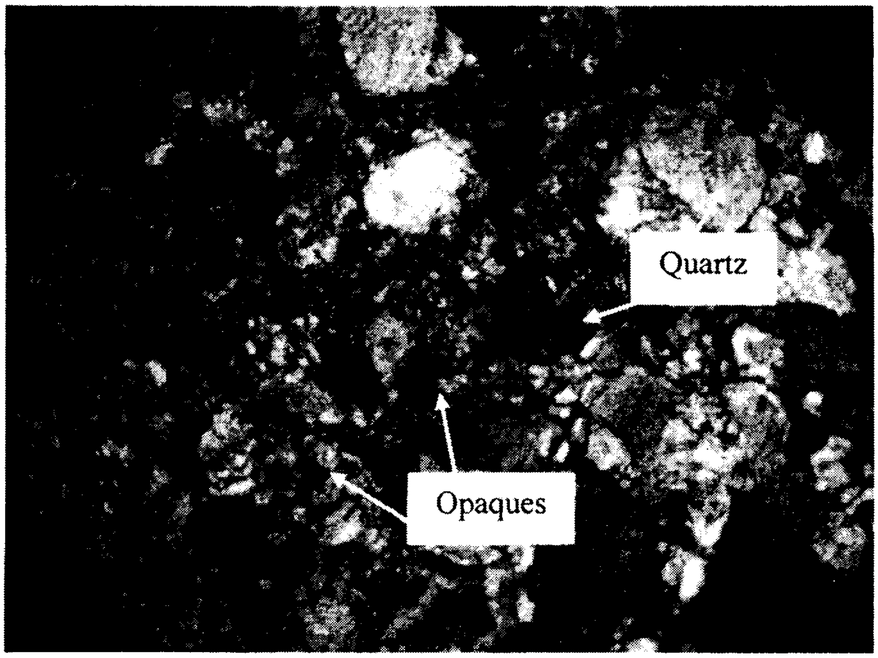


Figure A.16. Thin section of a rock fragment collected from the Bt1 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

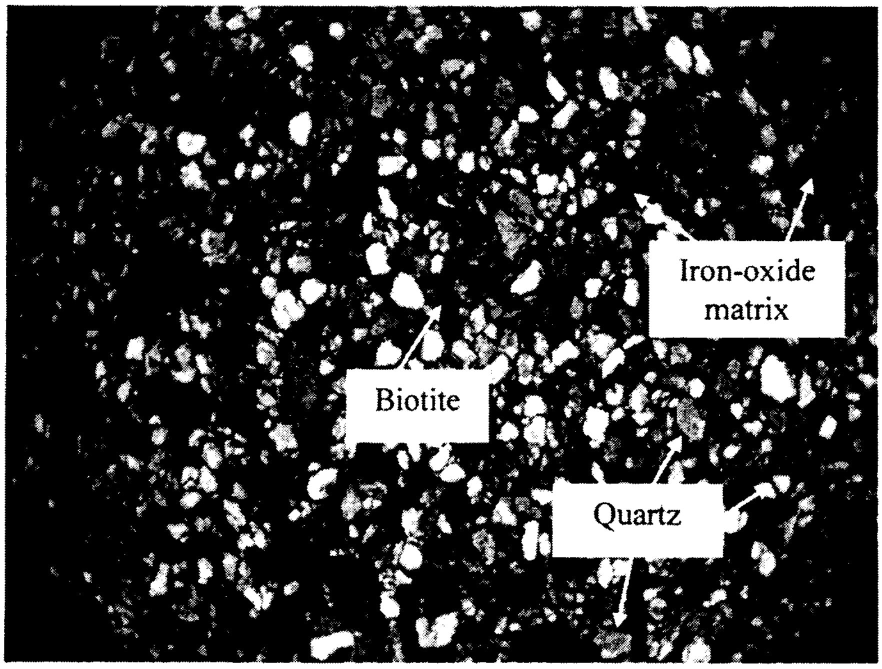


Figure A.17. Thin section of a rock fragment collected from the Bt1 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

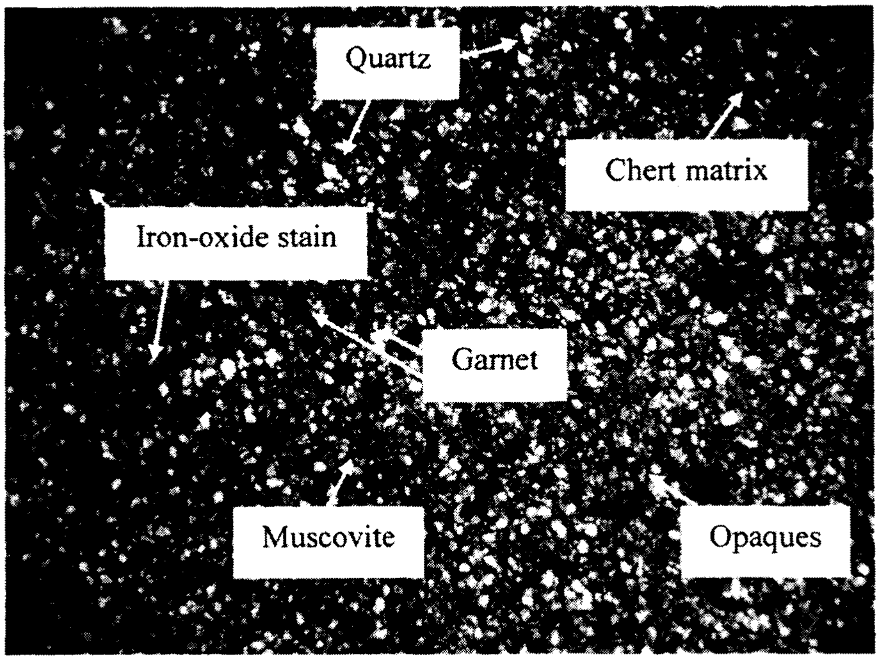


Figure A.18. Thin section of a rock fragment collected from the Bt2 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

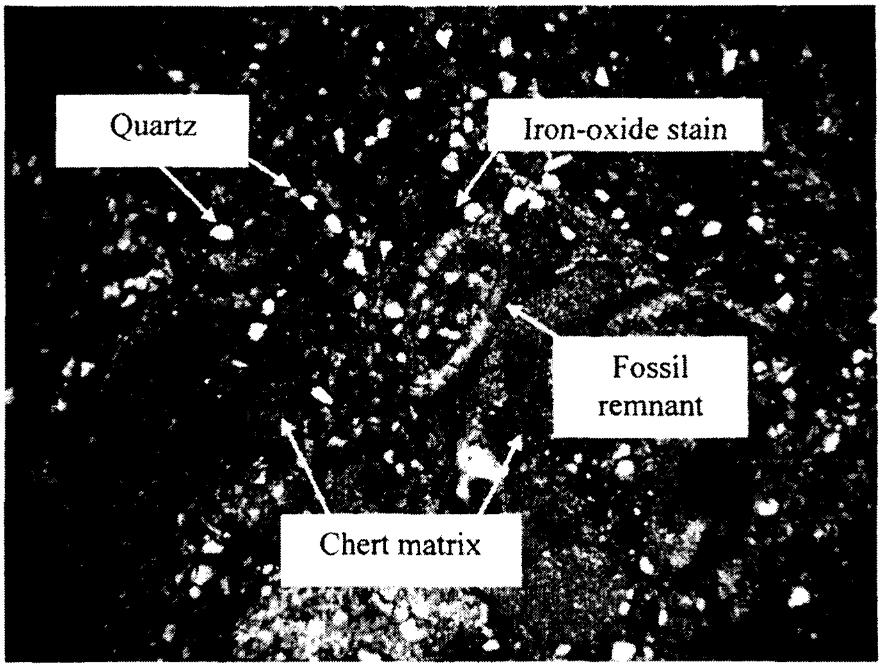


Figure A.19. Thin section of a rock fragment collected from the Bt2 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

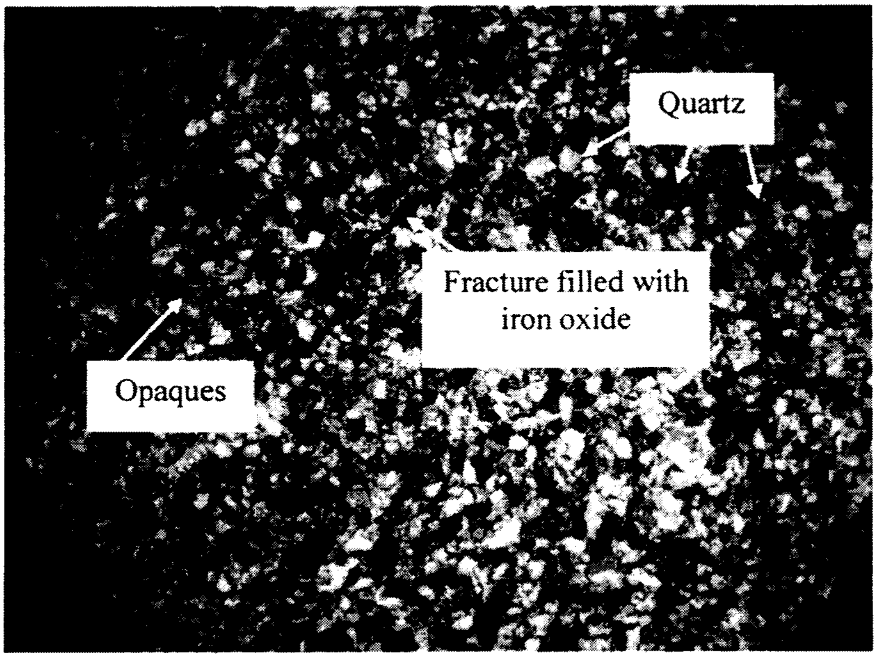


Figure A.20. Thin section of a rock fragment collected from the Bt3 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

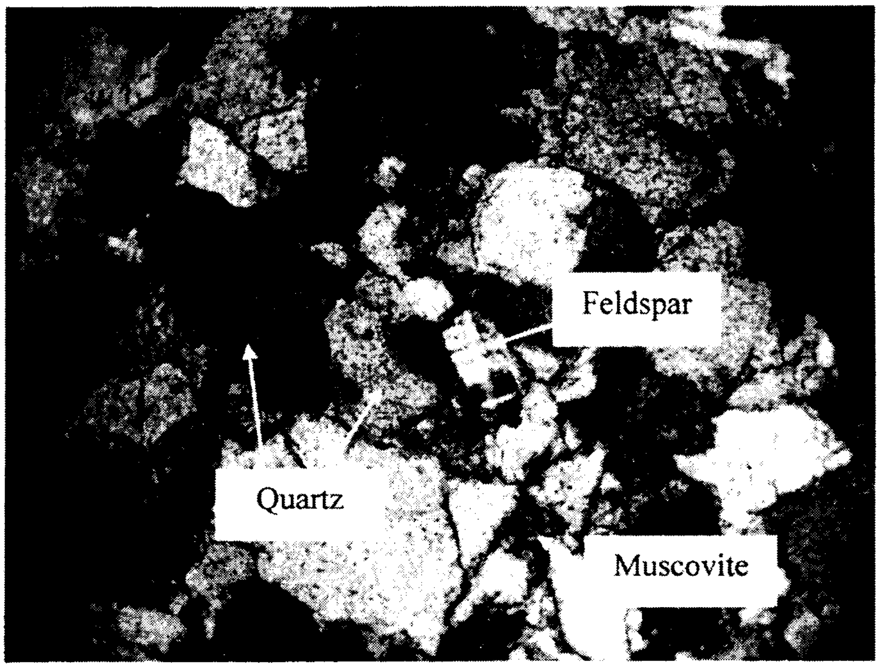


Figure A.21. Thin section of a rock fragment collected from the Bt3 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 100X, cross polarized light.

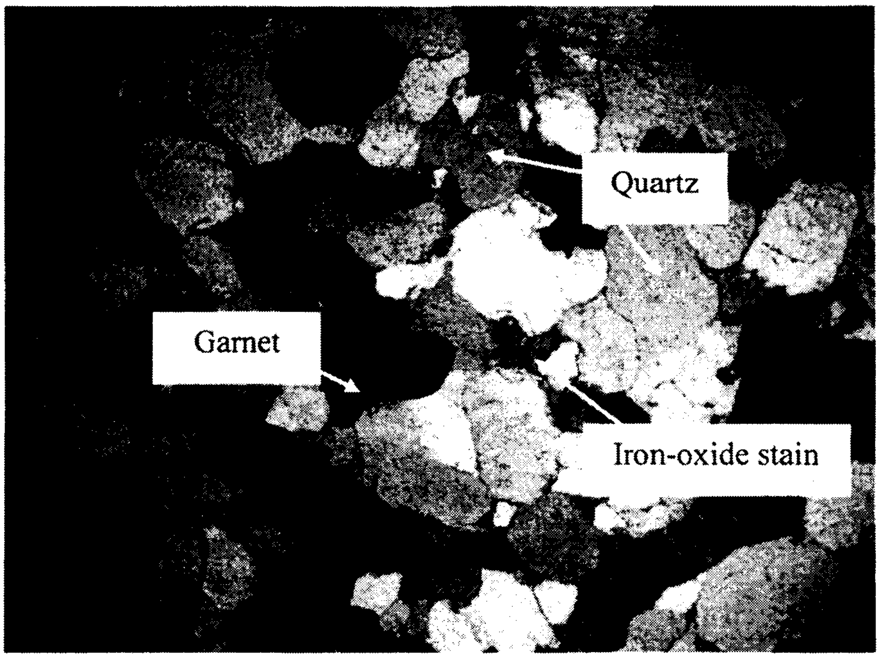


Figure A.22. Thin section of a rock fragment collected from the Bt1 horizon in the C3 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

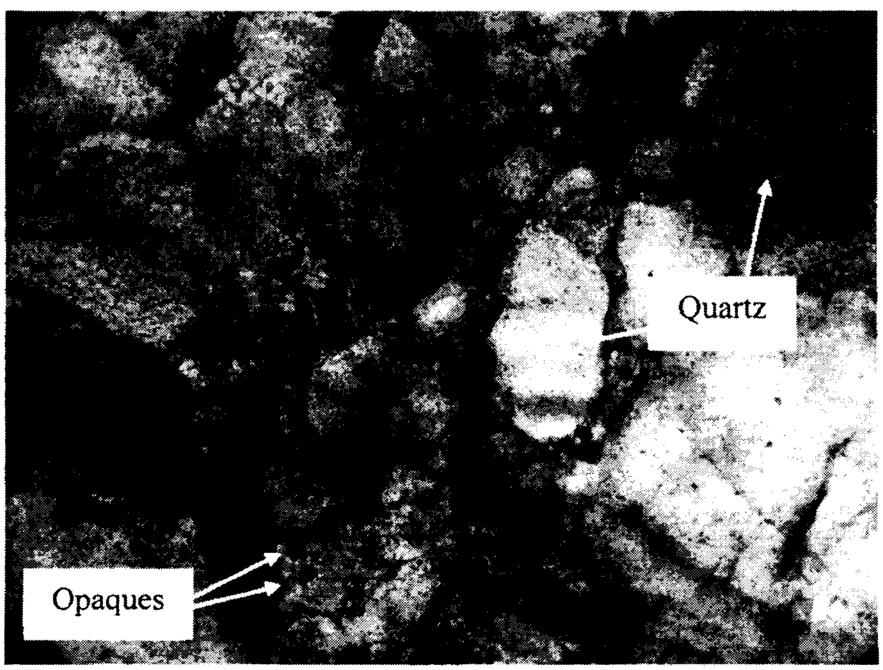


Figure A.23. Thin section of a rock fragment collected from the Bt1 horizon in the C3 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 100X, cross polarized light.

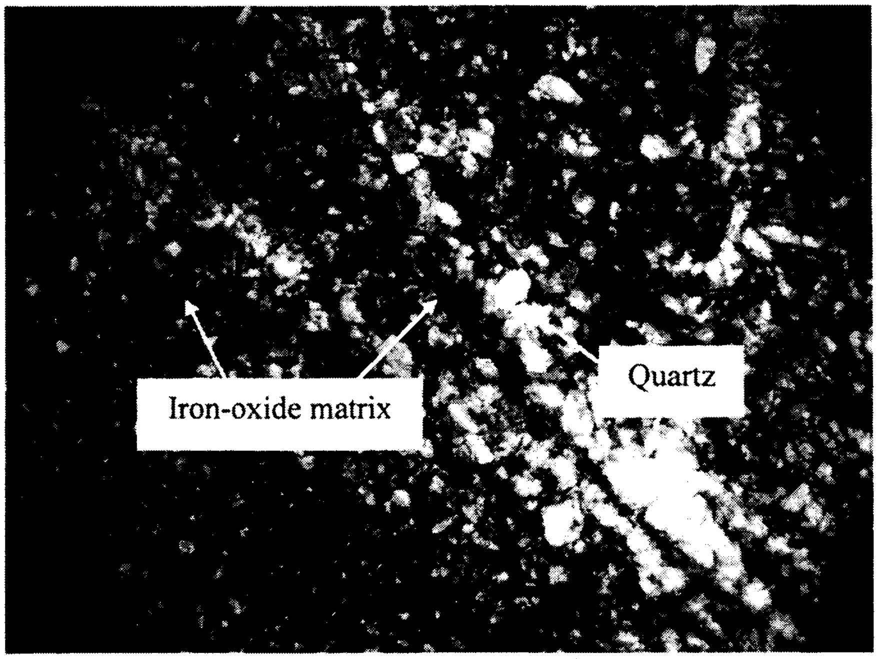


Figure A.24. Thin section of a rock fragment collected from the Bt1 horizon in the C3 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

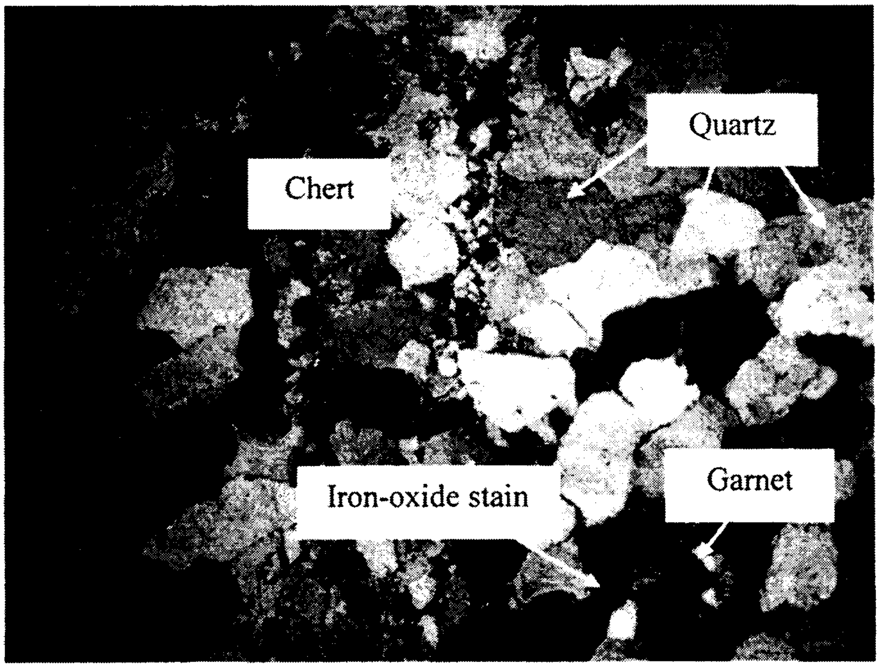


Figure A.25. Thin section of a rock fragment collected from the Bt2 horizon in the C3 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light. See Figure A.29 for a closer view of the garnet grain and iron-oxide stain.

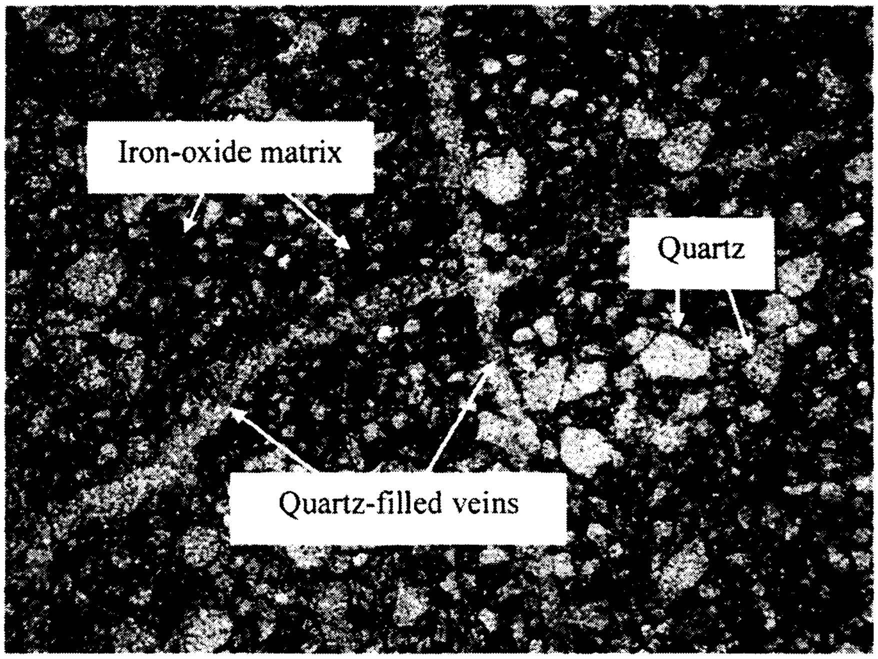


Figure A.26. Thin section of a rock fragment collected from the Bt2 horizon in the C3 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, plain polarized light.

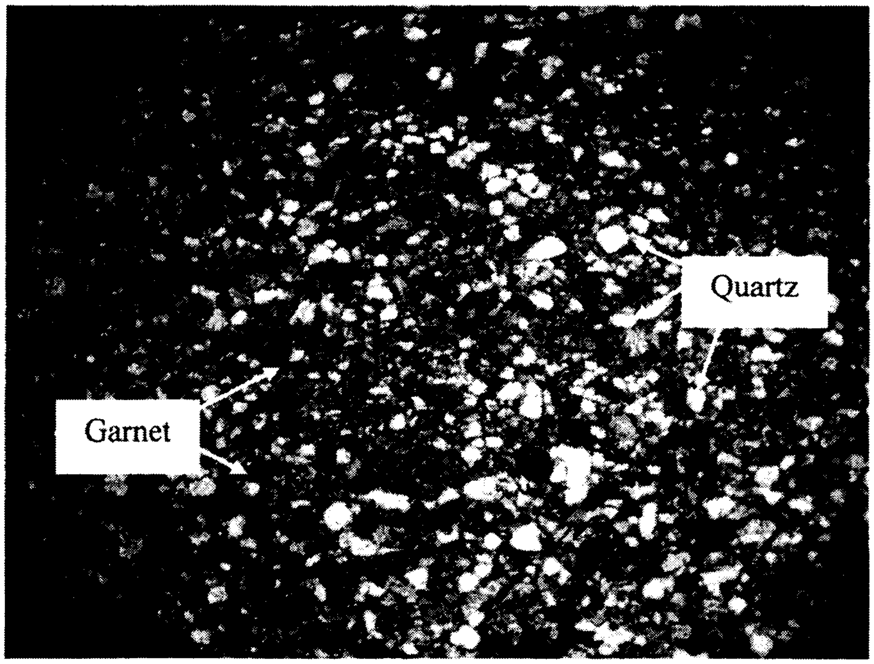


Figure A.27. Thin section of a rock fragment collected from the Bt2 horizon in the C3 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

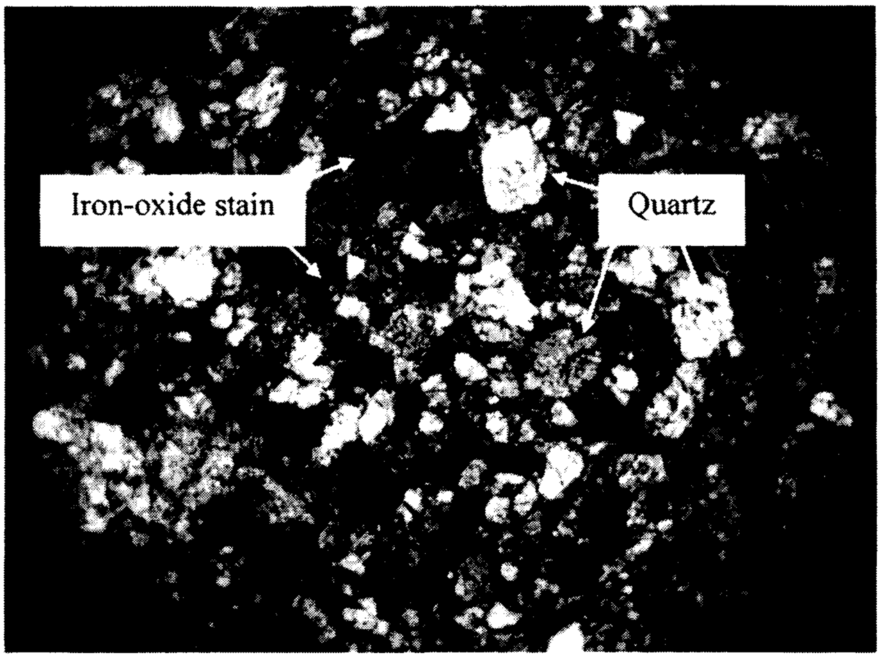


Figure A.28. Thin section of a rock fragment collected from the Bt2 horizon in the C3 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 40X, cross polarized light.

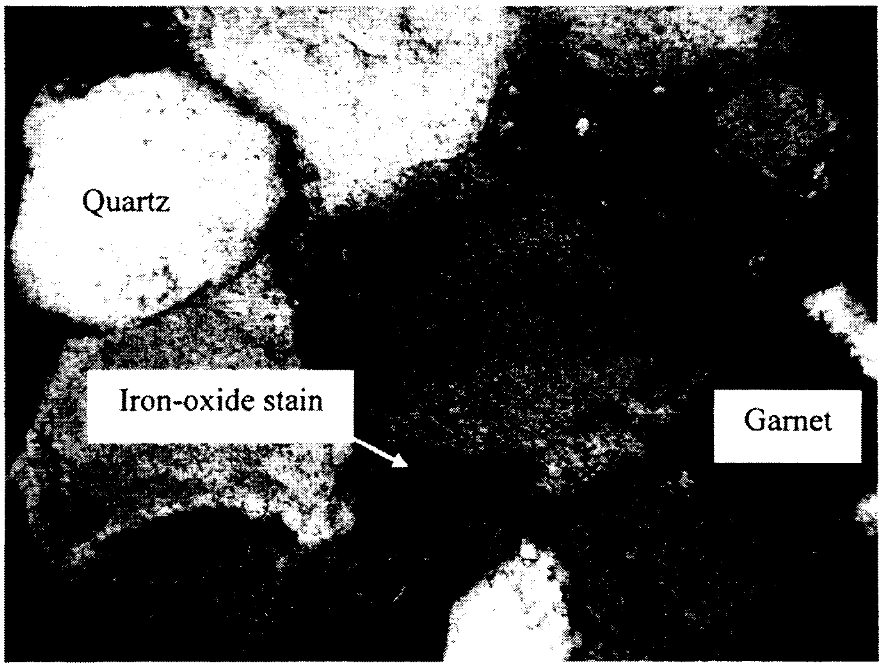


Figure A.29. Thin section of a rock fragment collected from the Bt2 horizon in the C3 pedon, Charkiln series, Spring Mountains, Clark County, Nevada. Magnification 100X, cross polarized light.

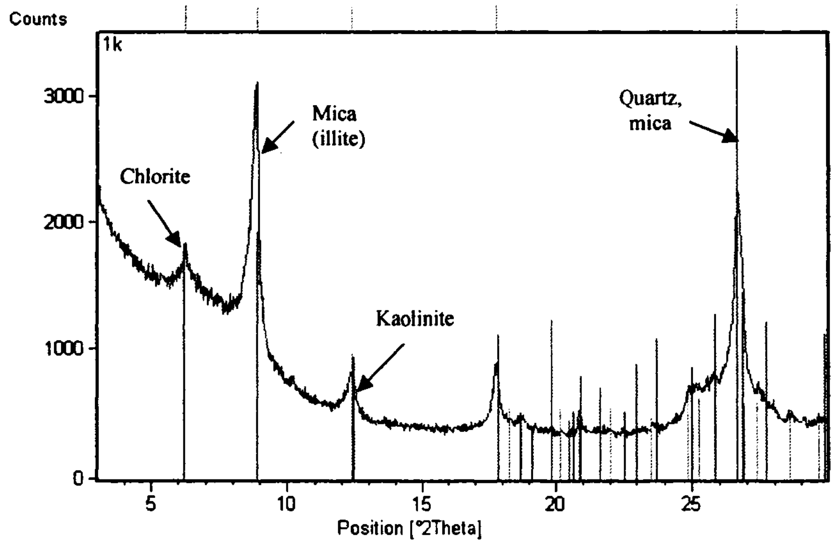


Figure A.30. XRD graph of minerals in the Bt1 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada.

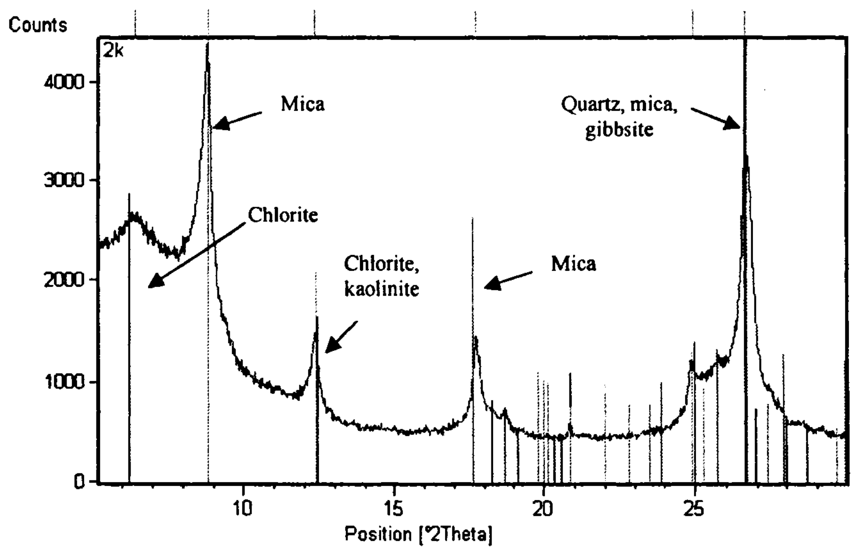


Figure A.31. XRD graph of minerals in the Bt2 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada.

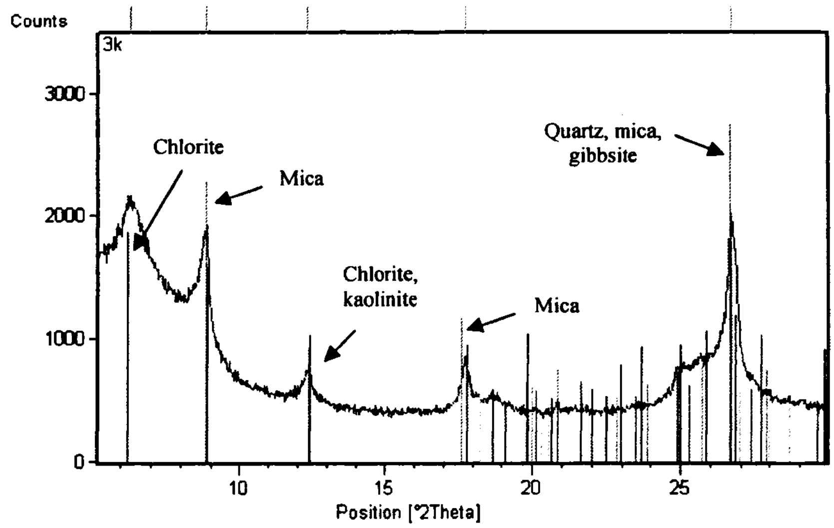


Figure A.32. XRD graph of minerals in the Bt3 horizon in the C1 pedon, Charkiln series, Spring Mountains, Clark County, Nevada.

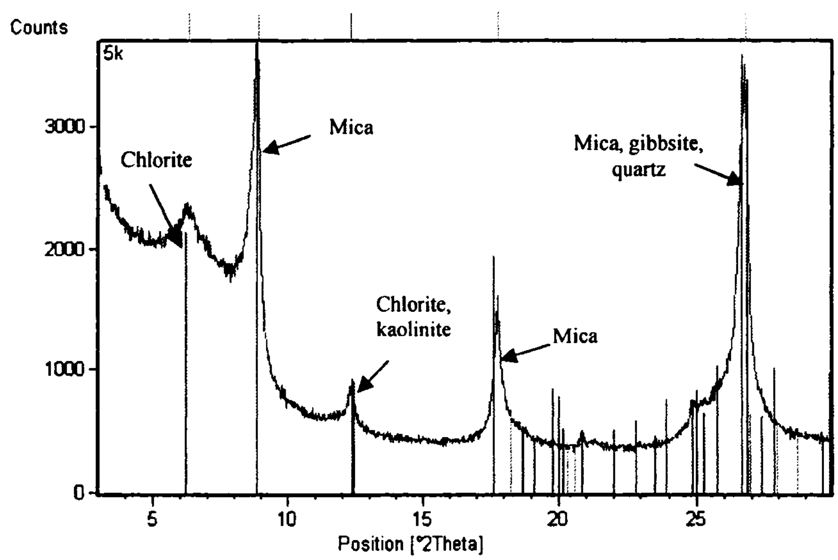


Figure A.33. XRD graph of minerals in the Bt1 horizon in the C3 pedon, Charkiln series, Spring Mountains, Clark County, Nevada.

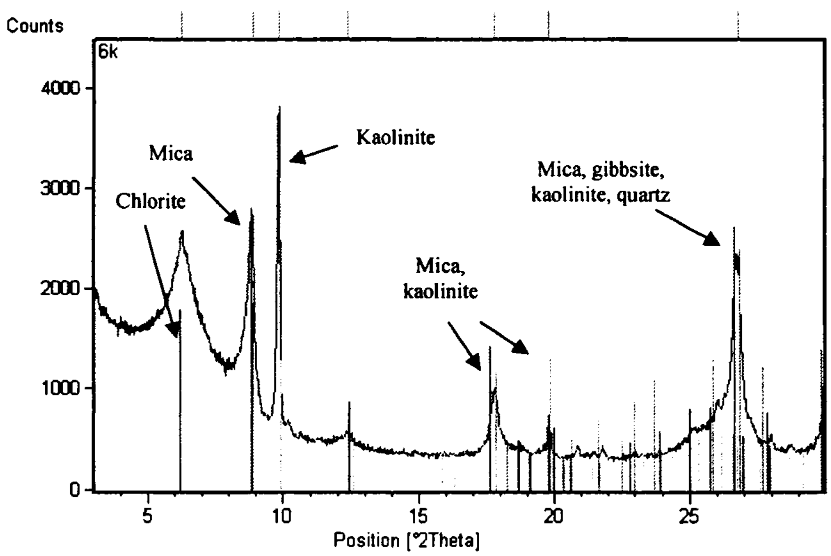


Figure A.34. XRD graph of minerals in the Bt2 horizon in the C3 pedon, Charkiln series, Spring Mountains, Clark County, Nevada.

REFERENCES

Aandahl, A.R., 1982, Soils of the Great Plains, land use, crops, and grasses: Lincoln, University of Nebraska Press, 282 p.

Alexander, E.B., 1995, Soil slope-depth relationships in the Klamath Mountains, California, and comparisons in central Nevada: Soil Survey Horizons, v. 36, p. 94-103.

Alexander, E.B., and Nettleton, W.D., 1977, Post-Mazama Natrargids in Dixie Valley, Nevada: Soil Science Society of America Journal, v. 41, p. 1210-1212.

Allen, B.L., and Fanning, D.S., 1983, Composition and soil genesis, *in* Wilding, L.P., Smeck, N.E., and Hall, G.F., eds., Pedogenesis and Soil Taxonomy I, Concepts and Interactions: Elsevier Science Publishers B.V., Amsterdam, p. 141-192.

Allen, B.L., and Hajek, B.F., 1989, Mineral occurrence in soil environments, *in* Dixon, J.B., and Weed, S.B., eds., Minerals in soil environments, second edition: Soil Science Society of America, Madison, 1244 p.

Anjos, L.H., Fernandes, M.R., Pereira, M.G., and Franzmeier, D.P., 1998, Landscape and pedogenesis of an oxisol-inceptisol-ultisol sequence in southeastern Brazil: Soil Science Society of America Journal, v. 62, p. 1651-1658.

Bilzi, A.F., and Ciolkosz, E.J., 1977, Time as a factor in the genesis of four soils developed in recent alluvium in Pennsylvania: Soil Science Society of America Journal, v. 41, no. 1, p. 122-127.

Birkeland, P.W., 1999, Soils and geomorphology: New York, Oxford University Press, 430 p.

Blaschke, H., 1979, Leaching of water-soluble organic substances from coniferous needle litter: Soil Biology Biochemistry, v. 11, p. 581-584.

Bloomfield, C., 1957, The possible significance of polyphenols in soil formation: Journal of the Science of Food and Agriculture, v. 8, no. 7, p. 389-392.

Brady, N.C., and Weil, R.R., 2002, The nature and properties of soils, thirteenth edition: Upper Saddle River, Prentice-Hall, Inc., 960 p.

Brewer, Roy, 1976, Fabric and mineral analysis of soils: Huntington, Robert E. Krieger Publishing Company, 482 p.

Buol, S.W., and Hole, F.D., 1961, Clay skin genesis in Wisconsin soils, *in* Stoops, Georges, and Eswaran, Hari, eds., Soil micromorphology: New York, Van Nostrand Reinhold Company, p. 242-244.

Burt, Rebecca, ed., 2004, Soil survey laboratory methods manual: National Resources Conservation Service Soil Survey Investigations Report No. 42, 735 p.

Cady, J.G., Wilding, L.P., and Drees, L.R., 1986, Petrographic microscope techniques, *in* Klute, Arnold, Campbell, G.S., Jackson, R.D., Mortland, M.M., and Nielson, D.R., eds., Methods of soil analysis, Part 1: Physical and mineralogical methods, second edition: American Society of Agronomy, Inc., no. 9, p. 185-218.

Carroll, Dorothy, 1959, Ion exchange in clays and other minerals: Bulletin of the Geological Society of America, v. 70, p. 749-780.

Catt, J.A., 1989, Relict properties in soils of the central and north-west European temperate region, *in* Bronger, A., and Catt, J.A., eds., 1989, Paleopedology-Nature and application of paleosols: Catena Supplement 16, p. 41-58.

Chadwick R.S., Barney, M.L., Beckstrand, D., Carley, J.A., Campbell, L., Jensen, E.H., McKinlay, C.R., Stock, S.S., and Stokes, H.A., 1975, Soil survey of Box Elder County, Utah, eastern part: USDA, Government Printing Office, Washington, D.C.

Christenson, G.E., and Purcell, Charles, 1985, Correlation and age of Quaternary alluvial-fan sequences, Basin and Range province, southwestern United States: Geological Society of America Special Paper 203, p. 115-122.

Ciolkosz, E.J., Cronce, R.C., Sevon, W.D., and Waltman, W.J., 1995, Genesis of Pennsylvania's limestone soils: Pennsylvania State University Agronomy Series No. 135, University Park, PA, 28 p.

Ciolkosz, E.J., Waltman, W.J., Simpson, T.W., and Dobos, R.R., 1989, Distribution and genesis of soils of the northeastern United States: Geomorphology, v. 2, p. 285-302.

Dohrenwend, J.C., Bull, W.B., McFadden, L.D., Smith, G.I., Smith, R.S.U., and Wells, S.G., 1991, Quaternary geology of the Basin and Range Province in California, *in* Morrison, R.B., ed., Quaternary nonglacial geology; Conterminous U.S.: Boulder, Colorado, Geological Society of America, The Geology of North America, v. K-2, p. 321-352.

Durgin, P.B., and Chaney, J.G., 1984, Dispersion of kaolinite by dissolved organic matter from Douglas-fir roots: Canadian Journal of Soil Science, v. 64, p. 445-455.

Faure, Gunter, 1991, Principles and applications of inorganic geochemistry: New York, Macmillan Publishing Company, 626 p.

Franzmeier, D.P., Bryant, R.B., and Steinhardt, G.C., 1985, Characteristics of Wisconsinan glacial tills in Indiana and their influence on argillic horizon development: Soil Science Society of America Journal, v. 49, p. 1481-1486.

Frei, Erwin, and Cline, M.G., 1949, Profile studies of normal soils of New York: II. Micromorphological studies of the gray-brown podzolic-brown podzolic soil sequence: Soil Science 68, p. 333-344.

Gager, H.M., 1968, Mössbauer spectra of deep-sea iron-manganese nodules: Nature (London), v. 220, p. 1021-1023.

Gangas, N.H., Simopoulos, A., Kostikas, A., Yassoglou, N.J., and Filippakis, S., 1973, Mössbauer studies of small particles of iron oxides in soil: Clays and Clay Minerals, v. 21, p. 151-160.

Gardner, L.R., 1972, Origin of the Mormon Mesa caliche, Clark County, Nevada: Geological Society of America Bulletin, v. 83, p. 143-156.

Gee, G.W., and Bauder, J.W., 1986, Particle-size analysis, *in* Klute, Arnold, ed., Methods of Soil Analysis, Part 1. Physical and Mineralogical Methods: American Society of Agronomy Monograph No. 9, 1188 p.

Gile, L.H., 1975, Holocene soils and soil-geomorphic relations in an arid region of southern New Mexico: Quaternary Research, v. 5, p. 321-360.

Gile, L.H., and Grossman, R.B., 1968, Morphology of the argillic horizon in desert soils of southern New Mexico: Soil Science, v. 106, p. 6-15.

Gile, L.H., and Hawley, J.W., 1968, Age and comparative development of desert soils at the Gardner Spring radiocarbon site, New Mexico: Soil Science Society of America Proceedings, v. 32, p. 709-716.

Gile, L.H., Hawley, J.W., and Grossman, R.B., 1981, Soils and geomorphology in the Basin and Range area of southern New Mexico-Guidebook to the Desert Project: New Mexico Bureau of Mines & Mineral Resources Memoir 39, 222 p.

Gile, L.H., Peterson, F.F., and Grossman, R.B., 1965, The K horizon-master soil horizon of carbonate accumulation: Soil Science, v. 99, no. 2, p. 74-82.

Gile, L.H., Peterson, F.F., and Grossman, R.B., 1966, Morphological and genetic sequences of carbonate accumulation in desert soils: Soil Science, v. 101, no. 5, p. 347-360.

Goenadi, D.H., and Tan, K.H., 1989, Micromorphology and mineralogy of illuviated clay in a Davidson soil: Soil Science Society of America Journal, v. 53, p. 967-971.

Graham, R.C., and Buol, S.W., 1990, Soil-geomorphic relations on the Blue Ridge Front: II. Soil characteristics and pedogenesis: Soil Science Society of America Journal, v. 54, p. 1367-1377.

Graham, R.C., Weed, S.B., Bowen, L.H., and Buol. S.W., 1989a, Weathering of iron-bearing minerals in soils and saprolite on the North Carolina Blue Ridge Front: I. Sand-size primary minerals: Clays and Clay Minerals, v. 37, no. 1, p. 19-28.

Graham, R.C., Weed, S.B., Bowen, L.H., Amarasiriwardena, D.D., and Buol. S.W., 1989b, Weathering of iron-bearing minerals in soils and saprolite on the North Carolina Blue Ridge Front: II. Clay mineralogy: Clays and Clay Minerals, v. 37, no. 1, p. 29-40.

Graham, R.C., and Wood, H.B., 1991, Morphologic development and clay redistribution in lysimeter soils under chaparral and pine: Soil Science Society of America Journal, v. 55, p. 1638-1646.

Grim, R.E., 1953, Clay mineralogy: New York, McGraw-Hill Book Company, Inc., 384 p.

Hopkins, D. G., and Franzen, D.W., 2003, Argillic horizons in stratified drift: Laverne end moraine, eastern North Dakota: Soil Science Society of America Journal, v. 67, p. 1790-1796.

Houghton, J.G., Sakamoto, C.M., and Gifford, R.O., 1975, Nevada's weather and climate: Nevada Bureau of Mines and Geology Special Publication 2, 78 p.

Hurst, V.J., 1977, Visual estimation of iron in saprolite: Geological Society of America Bulletin 88, p. 174-176.

Jenny, Hans, 1994, Factors of soil formation: New York, Dover Publications, Inc., 281 p.

Johnson, L.J., 1970, Clay minerals in Pennsylvania soils, relation to lithology of the parent rock and other factors: I. Clays and Clay Minerals, v. 18, p. 247-260.

Johnson-Maynard, J.L., Shouse, P.J., Graham, R.C., Castiglione, P., and Quideau, S.A., 2004, Microclimate and pedogenic implications in a 50-year-old chaparral and pine biosequence: Soil Science Society of America Journal, v. 68, p. 876-884.

Karpachevskiy, L.O., Babanin, V.F., Gendler, T.S, Opalenko, A.A., and Kuz'min, R.N., 1972, Identification of ferruginous soil minerals by Mössbauer spectroscopy: Soviet Soil Science, v. 4, p. 639.

Klein, Cornelis, and Hurlbut, C.S., Jr., 1993, Manual of mineralogy (after James D. Dana), 21st edition: New York, John Wiley & Sons, Inc., 681 p.

Longwell, C.R., Pampeyan, E.H., Bowyer, Ben, and Roberts, R.J., 1965, Geology and mineral deposits of Clark County, Nevada: Nevada Bureau of Mines Bulletin 62, 218 p.

Machette, M.N., 1985, Calcic soils of the southwestern United States, *in* Weide, D.L., and Faber, M.L., eds., Soils and Quaternary geology of the southwestern United States: Geological Society of America Special Paper 203, p. 1-21.

McDaniel, P.A., and Nielsen, G.A., 1985, Illuvial versus inherited clays in a cryoboralf of the Boulder Batholith, Montana: Soil Science Society of America Journal, v. 49, p. 156-159.

Mills, H.H., 1988, Surficial geology and geomorphology of the Mountain Lake area, Giles County, Virginia, including sedimentological studies of colluvium and boulder streams: U.S. Geological Survey Professional Paper 1469, 57 p.

Natural Resources Conservation Service, 2006a, Nevada snowcourse historical tables: U.S. Department of Agriculture, accessed August 13, 2006 at http://www.nv.nrcs.usda.gov/snow/nvcourse.html.

Natural Resources Conservation Service, 2006b, Official soil series descriptions: U.S. Department of Agriculture, accessed September 5, 2006 at http://soils.usda.gov/Soils/technical/classification/osd/index.html.

Natural Resources Conservation Service, 2006c, Soil moisture regimes map: U.S. Department of Agriculture, accessed September 23, 2006 at http://soils.usda.gov/use/worldsoils/mapindex/smr.html.

Natural Resources Conservation Service, 2006d, Soil Survey of Clark County area, Nevada: U.S. Department of Agriculture, 1818 p.

Nettleton, W.D., Flach, K.W., and Brasher, B.R., 1969, Argillic horizons without clay skins: Soil Science Society of America Proceedings, v. 33, p. 121-125.

Nettleton, W.D., Gamble, E.E., Allen, B.L., Borst, G., and Peterson, F.F., 1989, Relict soils of subtropical regions of the United States, *in* Bronger, A., and Catt, J.A., eds., Paleopedology: Nature and application of paleosols: W. Germany, Catena Verlag, Catena Supplement 16, p. 59-93.

Nettleton, W.D., Olson, C.G., and Wysocki, D.A., 2000, Paleosol classification: Problems and solutions: Catena, v. 41, p. 61-92.

Nettleton, W.D., Witty, J.E., Nelson, R.E., and Hawley, J.W., 1975, Genesis of argillic horizons in soils of desert areas of the southwestern United States: Soil Science Society of America Proceedings, v. 39, p. 919-926.

Ogg, C.M., and Baker, J.C., 1999, Pedogenesis and origin of deeply weathered soils formed in alluvial fans of the Virginia Blue Ridge: Soil Science Society of America Journal, v. 63, p. 601-606.

Ogg, C.M., Baker, J.C., and Edmonds, W.J., 2001, Quartzite-cobble weathering in alluvial-fan soils of the Virginia Blue Ridge: Soil Science, v. 166, no. 2, p. 127-136.

Page, W.R., Lundstrom, S.C., Harris, A.G., Langenheim, V.E., Workman, J.B., Mahan, S.A., Paces, J.B., Dixon, G.L., Rowley, P.D., Burchfiel, B.C., Bell, J.W., and Smith, E.I., 2005, Geologic and geophysical maps of the Las Vegas 30' x 60' quadrangle, Clark and Nye Counties, Nevada, and Inyo County, California: U.S. Geological Survey Scientific Investigations Map 2814.

Parsons, R.B., and Herriman, R.C., 1976, Geomorphic surfaces and soil development in the upper Rogue River Valley, Oregon: Soil Science Society of America Journal, v. 40, p. 933-938.

Peltier, L.C., 1949, Pleistocene terraces of the Susquehanna River, Pennsylvania: Pennsylvania Geological Survey Fourth Series Bulletin G23, 158p.

Peterson, F.F., 1981, Landforms of the Basin and Range province, defined for the soil survey: Soil Survey Technical Bulletin 28, 52 p.

Ray, L.L., 1963, Silt-clay ratios of weathering profiles of Peorian loess along the Ohio Valley: Journal of Geology, v. 71, no. 1, p. 38-47.

Reheis, M.C., 1988, Pedogenic replacement of aluminosilicate grains by $CaCO_3$ in Ustollic Haplargids, south-central Montana, U.S.A.: Geoderma, 41, p. 243-261.

Reheis, M.C., and Kihl, Rolf, 1995, Dust deposition in southern Nevada and California, 1984-1989: Relations to climate, source area, and source lithology: Journal of Geophysical Research, v. 100, no. D5, p. 8893-8918.

Reheis, M.C., Sowers, J.M., Taylor, E.M., McFadden, L.D., and Harden, J.W., 1992, Morphology and genesis of carbonate soils on the Kyle Canyon fan, Nevada, U.S.A.: Geoderma, v. 52, p. 303-342.

Ruhe, R.V., 1965, Quaternary paleopedology, *in* Wright, H.E., Jr., and Frey, D.G., eds., The Quaternary of the United States: New Jersey, University Press, p. 755-764.

Ruhe, R.V., 1975, Geomorphology, geomorphic processes and surficial geology: Boston, Houghton Mifflin Company, 246 p.

Schaetzl, R.J., and Mokma, D.L., 1988, A numerical index of podzol and podzolic soil development: Physical Geography, v. 9, no. 3, p. 232-246.

Schoeneberger, P.J., Wysocki, D.A., Benham, E.C., and Broderson, W.D., eds., 2002, Field book for describing and sampling soils, Version 2.0: Natural Resources Conservation Service, National Soil Survey Center, Lincoln, NE.

Schwertmann, U., 1993, Relations between iron oxides, soil color, and soil formation, *in* Bigham, J.M., and Ciolkosz, E.J., eds., Soil Color: Soil Science Society of America Special Publication No. 31, p 51-69.

Shaw, J.N., Odom, J.W., and Hajek, B.F., 2003, Soils on Quaternary terraces of the Tallapoosa River, central Alabama: Soil Science, v. 168, no. 10, p. 707-717.

Sherman, G.D., 1952, The genesis and morphology of the alumina-rich laterite clays, *in* Problems of clay and laterite genesis symposium at annual meeting of The American Institute of Mining and Metallurgical Engineers, St. Louis, Missouri, February 19-22, 1951, p. 154-161.

Smeck, N.E., Wilding, L.P., and Holowaychuk, N., 1968, Genesis of argillic horizons in Celina and Morley soils of western Ohio: Soil Science Society of America Proceedings, v. 32, p. 550-556.

Smith, B.R., and Buol, S.W., 1968, Genesis and relative weathering intensity studies in three semiarid soils: Soil Science Society of America Proceedings, v. 32, p. 261-265.

Sobecki, T.M., and Wilding, L.P., 1983, Formation of calcic and argillic horizons in selected soils of the Texas Coast Prairie: Soil Science Society of America Journal, v. 47, p. 707-715.

Soil Survey Staff, 1998, Keys to soil taxonomy, eighth edition: U.S. Department of Agriculture, Natural Resources Conservation Service, 326 p.

Soil Survey Staff, 1999, Soil Taxonomy: A basic system of soil classification for making and interpreting soil surveys: U.S. Department of Agriculture, Natural Resources

Conservation Service Agriculture Handbook 436, U.S. Government Printing Office, Washington DC, 869 p.

Soil Survey Staff, 2006, National soil survey characterization data: Soil Survey Laboratory, National Soil Survey Center, USDA-NRCS-Lincoln, NE, accessed September 5, 2006 at http://ssldata.nrcs.usda.gov.

Southard, R.J., and Southard, A.R., 1985, Genesis of cambic and argillic horizons in two northern Utah aridisols: Soil Science Society of America Journal, v. 49, p. 167-171.

Stolt, M.H., and Rabenhorst, M.C., 1991, Micromorphology of argillic horizons in an upland/tidal marsh catena: Soil Science Society of America Journal, v. 55, p. 443-450.

Taylor, R.M., McKenzie, R.M., Fordham, A.W., and Gillman, G.P., 1983, Oxide minerals, *in* Soils: an Australian viewpoint: Division of Soils, CSIRO, Melbourne, London Academic Press Ltd., p. 309-334.

Thurman, N.C., Ciolkosz, E.J., and Dobos, R.R., 1994, Soil Characterization Laboratory Methods Manual, second edition: Pennsylvania State University Agronomy Series 17, 119 p.

Torrent, J., Schwertmann, U., and Schulze, D.G., 1980, Iron oxide mineralogy of some soils of two river terrace sequences in Spain: Geoderma, v. 23, no. 3, p. 191-208.

Udden, J.A., 1898, The mechanical composition of wind deposits: Augustana Library Publication Number 1, 69 p.

United States Geological Survey, 1984, 7.5 minute topographic map of the Wheeler Well, Nevada quadrangle: United States Geological Survey, 1:24,000 scale.

Van Hoesen, J.G., and Orndorff, R.L., 2000, Evidence for Pleistocene glaciation in the Spring Mountains, Nevada: Geological Society of America Abstracts with Programs, v. 32, no.7, p. A-16.

Van Hoesen, J.G., and Orndorff, R.L., 2001, Using GIS to predict the spatial distribution of perennial snow and ice under modern and Pleistocene conditions in the Spring Mountains, Nevada: Geological Society of America Abstracts with Programs, v. 33, no. 3, p. A67.

Vincent, K.R., Bull, W.B., and Chadwick, O.A., 1994, Construction of a soil chronosequence using the thickness of pedogenic carbonate coatings: Journal of Geological Education, v. 42, p. 316-324.

Waltman, W.J., Cunningham, R.L., and Ciolkosz, E.J., 1990, Stratigraphy and parent material relationships of red substratum soils on the Allegheny Plateau: Soil Science Society of America Journal, v. 54, p. 1049-1057.

Weisenborn, B.N., and Schaetzl, R.J., 2005, Range of fragipan expression in some Michigan soils: I. Morphological, micromorphological, and pedogenic characterization: Soil Science Society of America, v. 69, p. 168-177.

Wentworth, C.K., 1922, A scale of grade and class terms for clastic sediments: Journal of Geology, v. 30, p. 377-392.

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Belcher, W.R., Sweetkind, D.S., and Elliott, P.E., 2002, Probability distributions of hydraulic conductivity for the hydrogeologic units of the Death Valley regional ground-water flow system, Nevada and California: U.S. Geological Survey Water-Resources Investigations Report 02-4212. http://water.usgs.gov/pubs/wri/wrir024212/

Bonner, L.J., Elliott, P.E., Etchemendy, L.P., and Swartwood, J.R., 1998, Water resources data, Nevada, water year 1997: U.S. Geological Survey Water-Data Report NV-97-1, 636 p.

Elliott, P.E., Beck, D.A., and Prudic, D.E., 2006, Characterization of surface-water resources in the Great Basin National Park area and their susceptibility to ground-water withdrawals in adjacent basins, White Pine County, Nevada [abs]: 2006 Annual Nevada Water Resources Association Conference, February 21-23, 2006, Mesquite, Nevada.

Elliott, P.E., Beck, D.A., and Prudic, D.E., 2006, Characterization of surface-water resources in the Great Basin National Park area and their susceptibility to ground-water withdrawals in adjacent valleys, White Pine County, Nevada: U.S. Geological Survey Scientific Investigations Report 2006-5099, 156 p. http://pubs.water.usgs.gov/sir2006-5099/

Laczniak, R.J., Smith, J.L., Elliott, P.E., DeMeo, G.A., Chatigny, M.A., Roemer, G.J., 2001, Ground-water discharge determined from estimates of evapotranspiration, Death Valley regional flow system, Nevada and California: U.S. Geological Survey Water-Resources Investigations Report 01-4195, 51 p. http://water.usgs.gov/pubs/wri/Wri014195/

Reiner, S.R., Laczniak, R.J., DeMeo, G.A., Smith, J.L., Elliott, P.E., Nylund, W.E., and Fridrich, C.J., 2001, Ground-water discharge determined from measurements of evapotranspiration, other available hydrologic components, and shallow water-level changes, Oasis Valley, Nye County, Nevada: U.S. Geological Survey Water-Resources Investigations Report 01-4239, 65 p. http://water.usgs.gov/pubs/wri/wri014239/

Reiner, S.R., Laczniak, R.J., DeMeo, G.A., Smith, J.L., Nylund, W.E., and Elliott, P.E., 1999, Preliminary estimate of mean annual evapotranspiration, Oasis Valley, Nye County, Nevada [abs.]: Eos, American Geophysical Union Transactions, v. 80, no. 44.

Vincent, K.R., Elliott, P.E., and Beck, D.A., 2003, Geomorphic and hydrologic analyses of the Wendover piedmont, Elko County, Nevada [abs]: Nevada Water Resources Association Annual Conference, February 26-28, 2003, Sparks, Nevada, p. 34.

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