The paleosols of the Upper Jurassic Morrison and Lower Cretaceous Cloverly formations, Northeast Bighorn Basin, Wyoming

by

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INTRODUCTION

General Statement

This report presents the results of a local study of the paleosols of the Upper Jurassic Morrison and Lower Cretaceous Cloverly formations in the northeast Bighorn Basin, Wyoming.

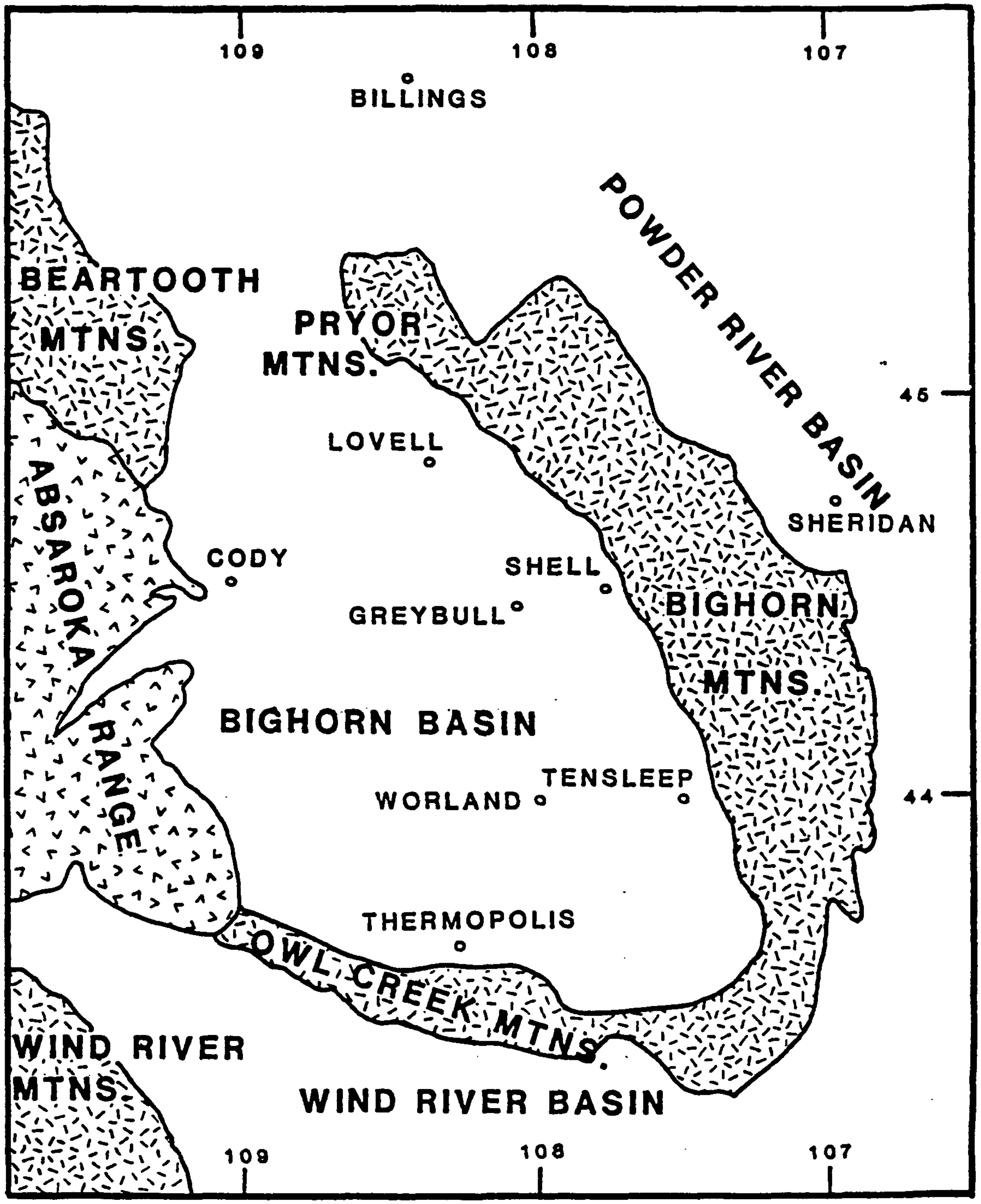
Extensive investigations of the Morrison and Cloverly date back to the late 1800s, but no detailed investigation of the paleosols has taken place. Early studies were primarily directed toward the lithologic and stratigraphic nature of the Morrison and Cloverly, and toward the rich assemblage of dinosaur remains preserved in these deposits.

In the present study the writer has focused major attention on the following aspects of the Morrison and Cloverly formations: 1) the orders of the paleosols in the study area, 2) the paleoclimatologic interpretations based upon the soil orders, and 3) the depositional history of the mudstones within which the paleosols occur.

Physiography and Location of the Study Area

The Bighorn Basin is located in northwest Wyoming and southcentral Montana. It is bounded by the Absaroca Range and the Beartooth Mountains to the west, the Pryor Mountains to the northeast, the Bighorn Montains to the east, and the

Figure 1. Map of the Bighorn Basin, northwest Wyoming



(modified after Kvale,1986)

Owl Creek Mountains to the south (Figure 1). The Absaroka Range is a volcanic range which was active during Eocene time. The rest of the mountains were formed during the Laramide compressional deformation of the North American craton (Coney, 1972) during Late Cretaceous (about 80 mys ago) to early Eocene time. The Laramide mountain ranges are characterized by exposed Precambrian cores flanked by Paleozoic and Mesozoic strata. Cenozoic rocks comprise the intermontane basin fill.

The study area is located along the western flank of the northern Bighorn Mountains. Two primary study areas were used for this project: 1) the badlands at Coyote Basin just east of the Beaver Creek Anticline, 2) and the badlands along the South Nose of Sheep Mountain Anticline. Field work was also performed at Rose Dome located to the northeast of the North Nose of Sheep Mountain Anticline, Douglas Draw near the town of Shell, and Devil's Kitchen east from the town of Greybull (Figure 2). The badlands at Coyote Basin are located at the confluence of Beaver Creek and Cedar Creek (Figure 2), and were of most interest because of the excellent exposures of both formations. A composite stratigraphic section of the Morrison and Cloverly formations from this area prepared by Kvale (1986) will be used here to indicate the stratigraphic position of the

Figure 2. Map of the study areas

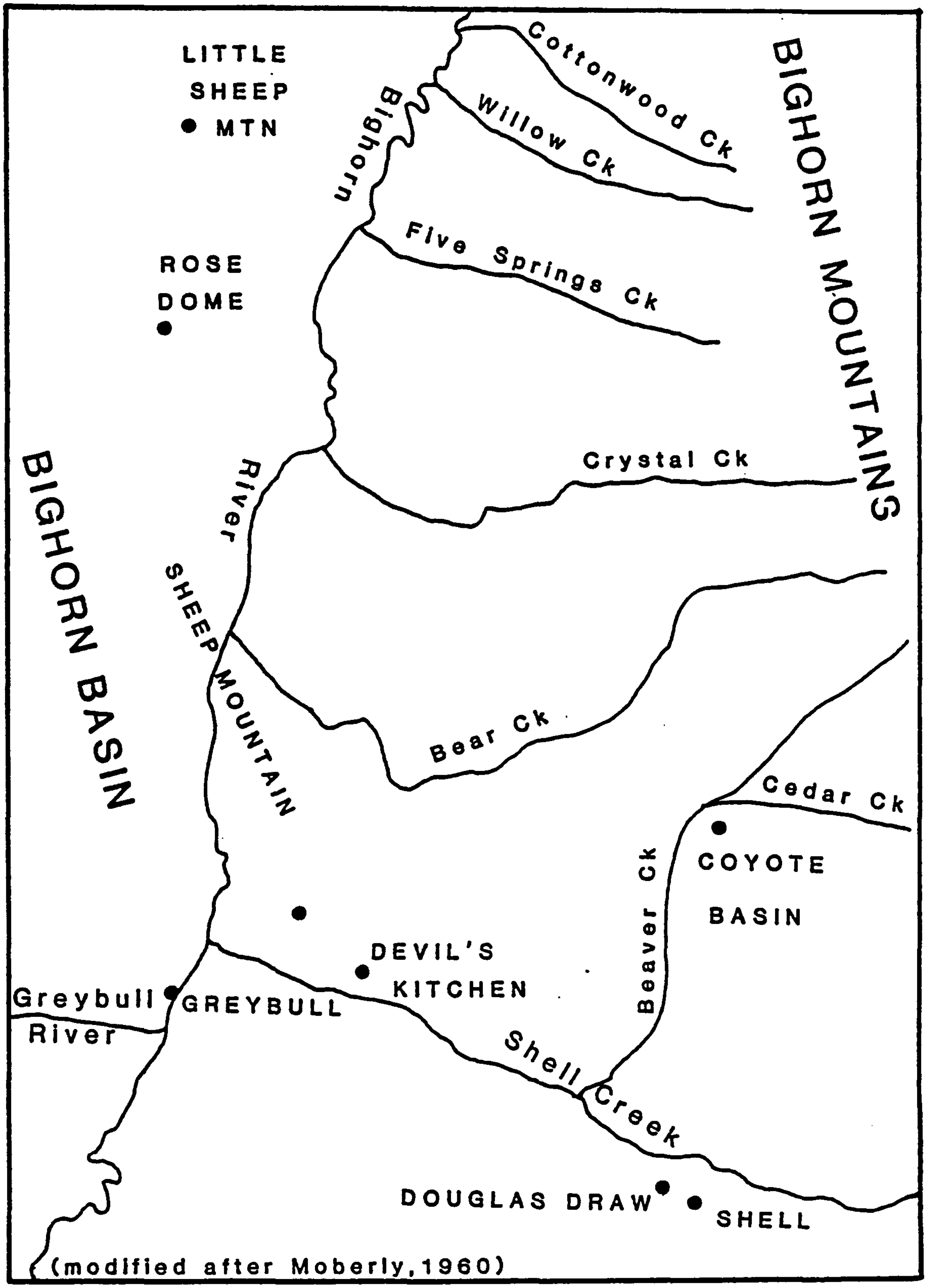
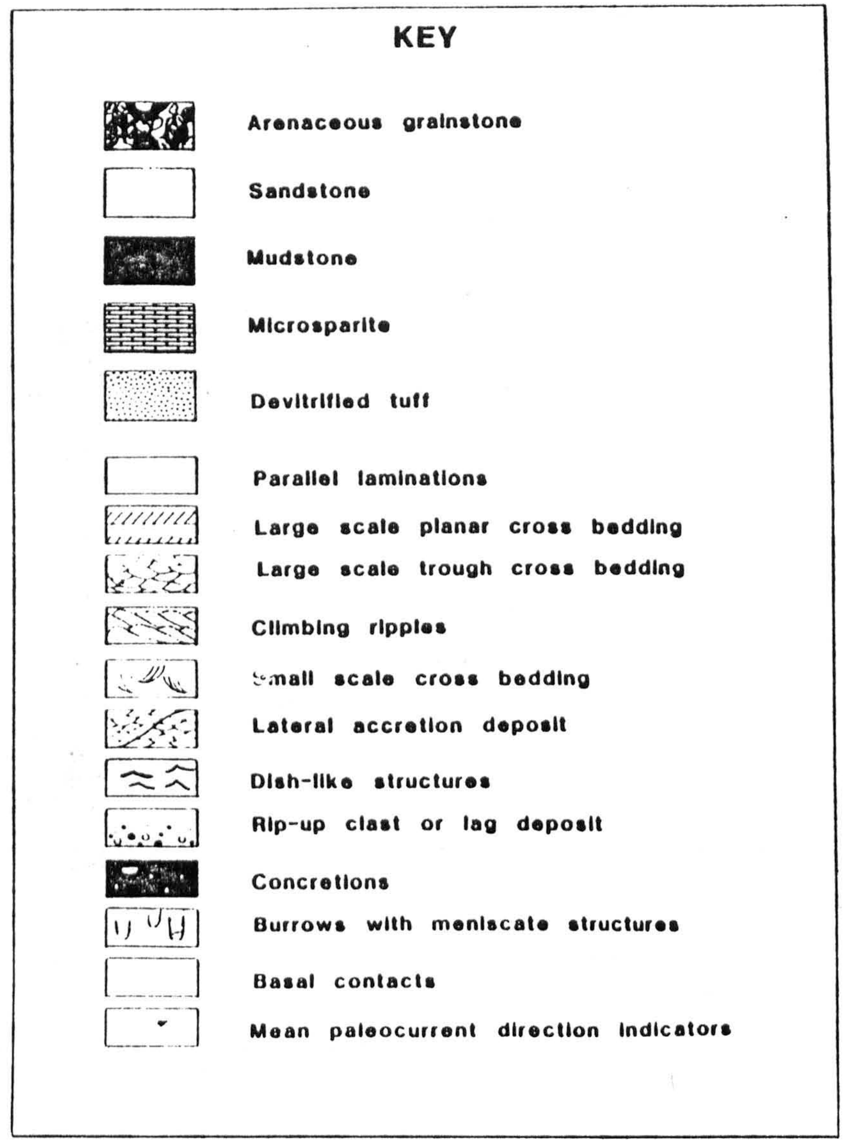
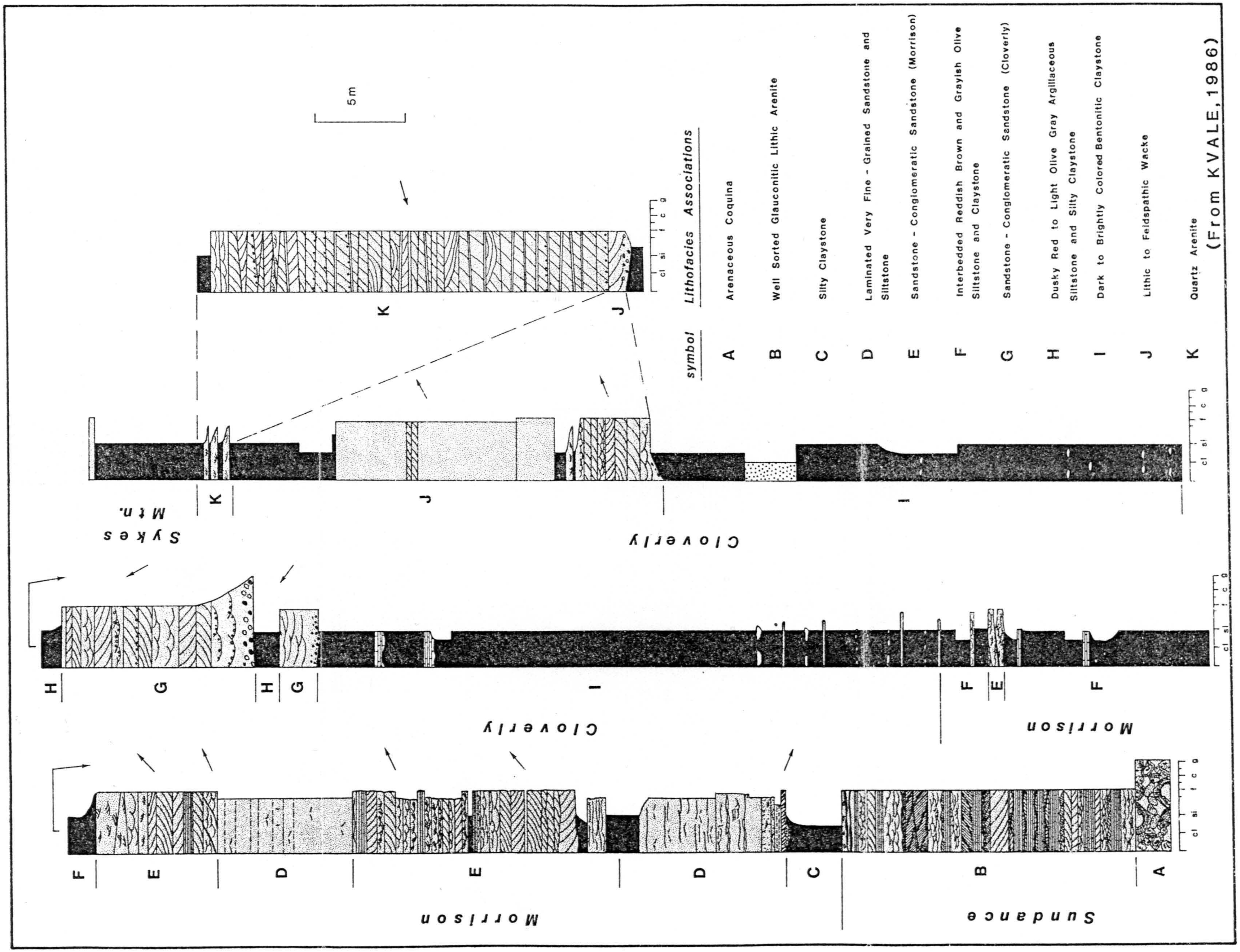

LITTLE SHEEP MTN
ROSE DOME
Bighorn River
Cottonwood Ck
Willow Ck
Five Springs Ck
BIGHORN MOUNTAINS
Crystal Ck
BIGHORN BASIN
SHEEP MOUNTAIN
Bear Ck
Cedar Ck
COYOTE BASIN
Beaver Ck
DEVIL'S KITCHEN
Greybull River
GREYBULL
Shell Creek
DOUGLAS DRAW
SHELL
(modified after Moberly, 1960)

Figure 3.
Composite stratigraphic section of the Morrison and Cloverly formations, Coyote Basin;
Erratum: Lithofacies Association E should read: Sandstone




5 m
symbol
Lithofacies Associations
A Arenaceous Coquina
B Well Sorted Glauconitic Lithic Arenite
C Silty Claystone
D Laminated Very Fine - Grained Sandstone and Siltstone
E Sandstone - Conglomeratic Sandstone (Morrison)
F Interbedded Reddish Brown and Grayish Olive Siltstone and Claystone
G Sandstone - Conglomeratic Sandstone (Cloverly)
H Dusky Red to Light Olive Gray Argillaceous Siltstone and Silty Claystone
I Dark to Brightly Colored Bentonitic Claystone
J Lithic to Feldspathic Wacke
K Quartz Arenite
(From KVALE, 1986)
Sundance
Morrison
Cloverly
Sykes Mtn.
cl si f c g

paleosols. The data concerning the Morrison Formation were obtained entirely from this area.

The exposures along the South Nose of Sheep Mountain Anticline, located 3 kilometers northeast from the town of Greybull, were used for the investigation of the upper part of the Little Sheep Mudstone Member of the Cloverly Formation as well as the Himes Member. The nearby Devil's Kitchen Badlands offered excellent exposures of the Himes Member.

The uppermost Cloverly, more specifically the overbank deposits of the Greybull sandstones and overlying mudstones just below the contact of the Cloverly with the overlying Sykes Mountain Formation was investigated at Rose Dome. Finally, the Douglas Draw area offered spectacular exposures of the Cloverly and less such of the Morrison.

Method of Investigation

During the summers of 1984 and 1985, a total of four months were spent studing the mudstones of the Morrison and Cloverly formations in the field. The major portion of the field work consisted of measurement, detailed description, and sampling of exposures in the study areas (Figure 2).

The Morrison and Cloverly exposures studied were essentially flat-lying and a Jacob staff and a tape measure were used for thickness measurements. Oriented rock samples were collected in the field to provide for a more complete description. The field description of exposures included color (using the Munsell Color Chart, 1954), lithology, texture and sedimentary structures, as well as pedogenic features such as plant roots, burrows, cutans, and calcium carbonate accumulations of different forms (glaebules, etc.).

Laboratory studies of the samples consisted primarily of thin-section and grain-mount examination, x-ray diffraction analysis, SEM (scanning electron microscope) analysis (Smart and Tovey, 1981; 1982), and a series of chemical analyses to determine free iron content using the sodium dithionite-citrate-bicarbonate method of Aguilera and Jackson (1953), ferrous iron content using the colorimetric orthophenanthroline method of Jackson (1958), aluminum content using the colorimetric aluminon method of Jackson (1958), carbonate content by hydrocloric acid treatment, and organic matter content by hydrogen peroxide treatment (Hutton, 1950; cited in Brewer, 1964). Mechanical analysis was also performed in an attempt to determine the textural characters of the different soil horizons. Since the study was concerned mainly with the description and interpretation of paleosols little attention was directed towards the paleontological aspects of the Morrison and Cloverly formations.

PREVIOUS WORK

Morrison and Cloverly formations

N. H. Darton (1904) was the first to describe the Morrison Formation in the Bighorn Basin comparing it with exposures in the Black Hills and the Colorado Front Range, and was also the one who introduced the name "Cloverly" Formation in his 1904 paper. Darton (1906) published a description of a typical exposure of the Cloverly about 1.5 miles west of the type locality where the then deserted Cloverly post office had existed.

Washburne (1909), Hewett (1914), Hintze (1915) and Lupton (1916) recognized the two formations throughout the Bighorn Basin but disagreed on the contact between them. Hewett and Lupton (1917) recognized the upper sandstone of the Cloverly and named it the "Greybull Sandstone". Hares (1917) introduced the name "Pryor Conglomerate" for the basal Cloverly and designated that unit as the lowest Member of the Cloverly, exposed primarily in the northeastern part of the Bighorn Basin along the west flank of the Pryor Mountains.

Wilmarth (1925) and Lee (1927) included the "rusty beds" as part of the Cloverly and recognized the Greybull Sandstone as the upper Member of the Cloverly. Pierce and Andrews (1940) argued that the Cloverly constitutes only the upper half of what is recognized as Cloverly in this report. They did not recognize the "rusty beds" and extended the Morrison Formation to include the part of the Cloverly which is refered to as the Little Sheep Mudstone in this report. Blackstone and Sternberg (1947) equated the Pryor Conglomerate Member with the Lakota and the Cloverly Formation with the Fuson Formation. Andrews et al. (1947) did not recognize the Cloverly formation but included it as part of the Morrison Formation. Rogers (1948) and Pierce (1948) argued that the entire nonmarine sequence belonged to the Morrison Formation while the Greybull Sandstone was assigned to the "rusty beds" and equated with the Cloverly Formation.

Moberly (1960) introduced new names and concepts which are generally accepted today. He recognized the Morrison Formation as overlying the marine Jurassic Sundance Formation. The Cloverly Formation conformably overlies the Morrison and according to Moberly (1960) can be divided into three members: 1) the Pryor Conglomerate Member, 2) the Little Sheep Mudstone Member, and 3) the Himes Member. The Pryor Conglomerate Member constitutes the basal unit of the Cloverly. It is a lenticular chert pebble conglomerate that occurs primarily in the northern Bighorn Basin. The Little Sheep Mudstone Member resulted from subaqeous weathering of volcanic debris in seasonal lakes and swamps. The Himes

Member consists of channeled sandstones and overbank mudstones. Moberly did not recognized the Greybull Sandstone but he named the "rusty beds" as Sykes Mountain Formation.

Ostrom (1970) divided the sequence into calcareous (labeled Units I, II, and III) and noncalcareous (Units IV, V, VI, and VII). The calcareous units correspond to Moberly's (1960) Morrison Formation while the noncalcareous units to the Cloverly Formation. Ostrom's (1970) stratigraphy of the Morrison and Cloverly formations is rather confusing, but his paleontologic study of the fossil vertebrates and their stratigraphic distribution is the most important product of his investigation.

Kvale (1986) studied the paleonvironments and tectonic significance of the Morrison and Cloverly formations. He divided the Himes Member into lower and upper units. The lower unit includes channel sandstones and overbank mudstones. The upper unit constitutes the Greybull sandstone which he interpreted as of fluvial origin with possible tidal influence.

Paleosols

"Although paleosols were recognized as early as 1795 along major unconformities in southestern Scotland by James Hutton, they have remained little studied because of the subsequently divergent aims and methods of soil and geological sciences" (Geol. Soc. Amer. News and Info., 1986). In the last 30 years geologists have studied paleosols ranging from Precambrian to Tertiary. Retallack (1984) described the problems and promise of Precambrian paleosols. Other significant studies of Precambrian paleosols are those of Collins (1925), Sharp (1940), Rankama (1955), Eskola (1963), Davidson (1965), Roscoe (1968), Williams (1968), McKee (1969), Donaldson (1969), Frarey and Roscoe (1970), Fraser et al. (1970), Kalliokosi (1975), Patel (1977), and Blades and Bickford (1979).

Boucot et al. (1974) studied Ordovician-Silurian paleosols of the Dunn Point and Moydart formations in Nova Scotia. Paleosols seem to be particularly abundant and well developed within the rocks of the Devonian System. Friend et al. (1970), Woodrow et al. (1973), Allen (1973, 1974a, 1974b) and Leeder (1976) described paleosols in the Catskill Formation in New York and Old Red Sandstone in Great Britain. These paleosols are characterized by calcrete horizons and are interpreted as being developed under semiarid paleoclimatic conditions. Similar paleosols were studied by McPherson (1979) in Victorialand, Antarctica, and Molenaar (1984) in Northern Ardennes, Belgium. Kidston and Lang (1921), Maiklem (1971), and Wardlaw and Reinson (1971) also described Devonian paleosols. Pennsylvanian age paleosols were studied by Keller et al. (1954) in Missouri, Hubert (1960) in Colorado, Walkden (1974) in England, and Walls et al. (1975) in Kentucky. Huddle and Patterson (1961), Roeschmann (1971) and Feofilova (1977) also recognized Carboniferous paleosols.

Permian and Triassic paleosols were described by Ortlam (1971) in southern Germany, and Danilov (1968) and Chalyshev (1969) in the USSR. Retallack's (1977a, 1977b) detailed work on Lower Triassic paleosols near Sidney, Australia, concluded that these soils developed under coniferous forests. Dunham (1969), Bernoulli and Wagner (1971), Bosellini and Rossi (1974), Steel (1974), Watts (1976, 1978), Estaban and Pray (1977), and Hubert (1977a, 1977b) have studied Triassic and Jurassic paleosols. McBride (1974) described Late Cretaceous terrestrial paleosols in Mexico. Faugeres and Robert (1969) in Greece, and Freytet (1971; 1973) in France described Late Cretaceous to Early Tertiary organic forest paleosols. Allen (1959, 1976) studied Late Cretaceous paleosols with red B and leached A horizons in England and France.

Paleosols studied in most detail are of Cenozoic age. Andreis (1972), and Spalletti and Mazzoni (1978) described savanna paleosols of the Eocene and Oligocene Sarmiento Group in Chubut Province, Argentina. Reeves (1970) studied caliche paleosols in the Great Plains of North America.

Schultz et al. (1955), Schultz and Stout (1955), and Harvey (1960) described floodplain paleosols of the Tertiary Orella Member of the Brule Formation in Nebraska. Johnson and Heron (1965), Estaban (1972), Buurman (1975), Abbott et al. (1979), and Peterson and Abbott (1979) described Eocene paleosols. Bown and Kraus (1981a, 1981b) described in great detail the Lower Eocene Willwood Formation in the Bighorn Basin, Wyoming and described Spodosols and Ultisols (classification according to the Soil Survey Staff, 1975). Buurman (1980) studied Paleocene paleosols in the Reading beds, Alum Bay, UK. Retallack (1979, 1983a) described Lower Tertiary paleosols in the Badlands National Monument, South Dakota. The Commission for Paleopedology of INQUA (1971) established criteria for the recognition and classification of paleosols which will be followed in this report.

STRATIGRAPHY AND SEDIMENTARY FACIES

The Morrison Formation (Oxfordian and Purbeckian; Moberly, 1960) conformably overlies the marine Jurassic Sundance Formation. The contact is very distinct due to color and lithology changes from green glauconitic lithic arenite to black claystone (Figure 4a). The thickness of the Morrison Formation at Coyote Basin is about 60m. The lower 15m maybe tidally influenced and no strong pedogenic features are present to suggest the existence of paleosols. It consists of silty claystone, siltstone and sandstone facies. In the southern part of the Bighorn Basin and in the Wind River Basin the Lower Morrison is described as an eolian deposit (Dan Weed, Dept. of Earth Sciences, Iowa State Univ., pers. comm., 1986). The rest of the Morrison Formation is characterized by interbedded sandstone and variegated mudstone facies which were interpreted by Mantzios and Vondra (1986a; 1986b) and others as fluvial channel and overbank deposits. At Coyote Basin the channels are mostly medium- to fine-grained quartz arenites, fining upwards to siltstones, with large-scale trough cross-bedding. The sandstone facies is interpreted as channel margin and channel interior deposits, point bar complexes and mid-channel bars (Kvale, 1986) which are typical of low sinuosity streams. The mudstone facies dominate the upper part of the Morrison Formation. Their color (Munsell Color

Figure 4a. The contact between the marine Jurassic Sundance and the nonmarine Upper Jurassic Morrison formations at Coyote Basin

Figure 4b. The contact of the Upper Jurassic Morrison and Lower Cretaceous Cloverly formations at Coyote Basin

Chart, 1954) ranges from dark reddish brown (5YR3/2) to weak red (10R4/3) and from grayish olive green (5GY3/2) to greenish gray (5GY6/1). These are stratigraphically persistent horizons with variable thickness ranging from 15cm to 170cm, and show pedogenic features such as plant roots, burrows, glaebules, pedotubules, cutans, color mottling and clay-rich zones (Mantzios and Vondra, 1986a).

Bones commonly occur within the variegated mudstone facies of the upper Morrison. They are most abundant in the greenish (A) horizons of the paleosols. Ostrom (1970) identified fossil remains of Stegosaurus, Camptosaurus, Allosaurus and Apatosaurus. The famous Howe Dinosaur Quarry (Brown, 1935; 1937) is located near the Coyote Basin study area where over 4000 dinosaur bones were collected by Barnum Brown and his American Museum team. The bones occur in a greenish mudstone beneath a channel sandstone within the variegated mudstone facies. They were identified by Dale Russel (Canadian Museum of Natural History, 1986, pers. comm.) as to belong mostly to juvenile Diplodocus.

The Cloverly Formation (Albian to Aptian; Ostrom, 1970) overlies the Morrison Formation with a paraconformity (Figure 4b). At Coyote Basin the Pryor Conglomerate Member is exposed truncating the Morrison Formation. Its dominant lithology is black chert pebble conglomerate. Large-scale trough cross-bedding indicates a north to south paleocurrent direction. It is interpreted by Moberly (1960) as a braided stream complex. The Pryor Conglomerate Member is exposed in the northern part of the Bighorn Basin where its thickness reaches more than 10m (Moberly, 1956) near the Pryor Montains, but it thins toward the southern part of the Basin. Kvale (1986) questioned the usefulness of accepting this as a formal unit, because similar conglomerates are reported in the literature at different stratigraphic positions and are not always confined in the basal Cloverly.

The Little Sheep Mudstone Member overlies the Morrison Formation with a distinct contact due to a color change from red to gray (Figure 4b). It has a variable thickness of 80-110m. Its dominant lithology is bentonitic mudstone ranging in color from dark gray (5Y4/1) to olive gray (5Y5/2) to grayish red purple (5RP4/2). The lower part of the Little Sheep Mudstone Member is dominated by dark gray and olive gray colors but the upper part is mostly dark gray and purple. Calcium carbonate and silica nodules, septaria, bone fragments, and silicified wood fragments occur in the mudstone facies. Sandstone facies interbedded with the mudstone facies represent flashy ephemeral streams (Kvale, 1986). A clay-playa complex is believed to have been the site of deposition of the mudstones with the flashy ephemeral streams supplying the sediments. Similar environments have been described in the ancient record by

Tunbridge (1984) and in the modern record by Hardie et al. (1978). The pedogenic features of the Little Sheep Mudstone clay-playa complex are similar to those in the Morrison Formation. A conglomeratic sandstone with an associated overbank mudstone occurs stratigraphically in the middle of the Little Sheep Mudstone Member. This is a braided stream complex indicating an episode of rapid deposition. The overbank mudstone consist mainly of paleosols with features similar to those of the upper Morrison Formation. Dinosaur bone fragments occur at the uppermost green horizon and at the base of the channel deposit. The upper part of the Little Sheep Mudstone Member is characterized by mudstones with laterally persistent horizons of chert-carbonate nodules. The mudstones are gray (7.5YR4/0) and grayish red purple (5RP4/2) in color. Lenticular devitrified tuff beds are present at various stratigraphic levels. They are locally reworked and contain cobble-size rounded quartzite and chert clasts. The underlying paleosols have been reworked and/or truncated. The depositional environment of the upper Little Sheep Mudstone Member was interpreted by Moberly (1960) as a seasonal lake (?) deposit and by Kvale and Vondra (1985) as a playa deposit.

The Himes Member disconformably overlies the Little Sheep Mudstone Member. The lower portion of the Himes Member is composed of channel sandstones, lithic to feldspathic wackes that contain abundant clasts of zoned plagioclase, polycrystalline quartz and volcanic rock fragments (Kvale, 1986). Large-scale trough cross-bedding indicates a west to east paleocurrent for the lower Himes channels. A densely mottled red mudstone overlies the lower Himes channels. It is interpreted as floodplain deposit with an associated paleosol (Vertisols; Mantzios and Vondra, 1986b).

The Greybull sandstone is the uppermost unit of the Himes Member. It was deposited mostly under fluvial conditions though some tidal influence is suspected (Kvale, 1986). A detailed study of the Greybull sandstone by Kvale (1986) revealed the presence of three major channel types which he termed type I, II, and III from youngest to the oldest. Type I are low sinuosity sand filled channels composed of fine-grained quartz arenites with unidirectional large-scale planar cross-bedding. Type II are high sinuosity, mud- and sand-filled channels composed of massive to ripple laminated mudstones and very fine-grained quartz arenites which pass into thicker large-scale trough cross-bedded sandstones near the center of the channel. Type III are high sinuosity sand-filled channels composed of ripple laminated very fined-grained quartz arenites mostly confined to point bars and massive overbank mudstones. Paleocurrent analysis shows an east to west direction of flow (Kvale, 1986). Finally, massive, mottled, red (5R2/2) and olive gray (5Y3/2) overbank mudstones with turtle and crocodile remains occupy the uppermost part of the Cloverly Formation. They interfinger with the overlying siltstones of the tidally influenced Sykes Mountain Formation.

Vertebrate bones are not as abundant in the Cloverly Formation. Ostrom (1970) discovered that "three taxa are represented by large enough samples to provide meaningful stratigraphic ranges. These are *Tenontosaurus tilletti* (gen. and sp. nov.), *Sauropelta edwardsi* and *Deinonychus Antirrhopus*".

PALEOSOL FEATURES

Paleosols are preserved ancient soils whose recognition is based on various features of modern soils (Retallack, 1977a; 1981; 1986). Bown and Kraus (1981a) defined paleosols as "any broadly tabular rock bodies that show evidence of having been exposed on the ground surface after deposition for long enough periods to have been modified by physical and chemical weathering including the actions of organisms". Some criteria used to recognize and classify modern soils such as moisture content and rainfall regime, soil temperature, vegetation, soil organisms, base saturation, and original soil pH and Eh are not available for the study of paleosols (Bown and Kraus, 1981a). Post-pedogenic diagenesis can also alter pedogenic features such as peds, cutans and soil structure. Therefore, conclusions regarding the existence of paleosols must be made by using features which can be preserved in the geologic record such as plant roots (rhizoliths), burrows, color mottling, soil horizon development, bioturbation, and carbonate accumulation in the form of glaebules and/or pedotubules (Brewer and Sleeman, 1964; Brewer, 1964). Geochemical and petrographic data along with field observations can be used effectively to support conclusions about the occurrence of paleosol.

Paleosols (also called fossil soils) are studied for their potential as tools for reconstructing ancient environments (Retallack, 1981). They can also provide information regarding paleoclimate (Mantzios and Vondra, 1986b), paleoecology and paleogeography. This study presents evidence concerning the climatic conditions during the deposition of the Morrison and Cloverly formations and attempts to provide a detailed reconstruction of the local depositional environments. A more extensive study on a regional scale (hopefully to be undertaken by the writer as Ph.D. dissertation) will provide information concerning the regional depositional environments, paleoclimate and paleogeography of the Morrison and Cloverly formations in the Bighorn and adjoining intermontane basins in Wyoming.

Morrison Formation

The mudstones of the Morrison Formation show a number of pedogenic features suggesting extended periods of soil formation. These features are discussed below and are based on both field observations and laboratory analyses.

Soil Horizons

Two master soil horizons (Soil Survey Staff, 1975) are recognized in the Morrison Formation, those being the A and the B (defifitions in Appendix A). The A horizons are green in color and always overlie the red B horizons. They possess greater amount of organic matter content than the B horizons

(Table 2), and plant roots and burrows penetrate through them down to the B horizons. Their thickness is variable depending on the duration of non-sedimentation periods. The contact between the A and B horizons is rather sharp meaning that no diagenetic processes controlled the coloration of the two horizons (Bown and Kraus, 1981a). According to the Soil Survey Staff (1975), the A horizons represent diagnostic surface horizons called ochric epipedons (definition in Appendix). The B horizons are well developed subsurface horizons with illuvial concentrations of clay, iron and aluminum. They are classified as spodic (Bs), argillic (Bt) and/or calcic (Bk) diagnostic horizons (Soil Survey Staff, 1975; definitions in Appendix). The Morrison Formation is characterized by mature (well developed) soil horizons. Stratigraphically, the oldest paleosol is made up of O (?), A, Bs (spodic) and Bt (argillic) horizons (Figure 5). The remaining paleosols are made of A and Bt (argillic) or Btk (combination of argillic and calcic) horizons (Figure 6). Superimposed horizons, though difficult to demonstrate, are present and are indicative of long periods of non-sedimentation (Birkeland, 1984). The soil horizons present in the Morrison Formation along with the pedogenic features present in them will be used in an attempt to determine the soil orders of the Morrison paleosols according to the modern taxonomy of the Soil Survey Staff (1975).

Figure 5. Spodosols underlying channel sandstone at Coyote Basin (the outcrop height is 4m). This represents horizons 5 through 9 in Table 2

Figure 6. Argillic-calcic (Btk) horizon at Coyote Basin (horizon 10, left side of photo is 95cm)

Color

Two colors, green and red, characterize the mudstones of the Morrison Formation. The red mudstones fall in the 10R-5YR range and the green in the 5GY range of the Muncell Color Chart (1954). The origin of color in the clastic sedimentary rocks is still a matter of debate (Morad, 1983). Diagenetic processes have been suggested by several authors. Friend (1966) concluded that the differentiation between red and non-red occurs shortly after deposition and is the result of differences in oxidation-reduction potentials which can be determined by the position of the water table. Walker (1967) suggested a postdepositional origin for hematite in red sediments due to in situ alteration of unstable iron-bearing detrital minerals. Thompson (1970) made similar conclusions as Friend (1966) suggesting that the green and gray colors are due to reduction and dissolution of ferric oxide or ferric hydrate pigment. Van Houten (1972) noted that although some ferric oxides might have been contributed by alteration of iron-silicates for a tropical savannah alluvium, enough soil-derived ferric oxides were present to produce the red color by postdepositional dehydration and ageing. Turner (1980) proposed the process of "ageing" during which iron oxides convert into hematite (which constitutes the red pigment). The conversion to hematite could be partly of pedogenic origin in semiarid and arid climatic conditions (Walker, 1967; Bown and Kraus, 1981a). Morad (1983) suggested a diagenetic origin of color based on petrographic examination of the clay minerals and on iron content analysis of the total sample.

The red mudstones of the Morrison Formation owe their color to the distribution and amount of iron oxides. Grain-mount examination revealed silt-size hematite but no detrital iron-bearing minerals (biotite, hornblende, etc.). This supports the postdepositional origin of hematite of Walker (1967). Hematite is partly of pedogenic nature due to the semiarid to arid paleoclimate which is suggested by the writer for the Morrison Formation. Hematite occurs as a coating around grains suggesting a diagenetic origin (Wilson and Pittman, 1977).

The red mudstones in the Morrison Formation are also characterized by mottling (definition in Appendix A), which is analogous to the modern soil "gleying" in which the reduction and mobilization of iron, manganese and other elements has taken place at the air/groundwater interface as a result of a fluctuating water table (McPherson, 1979). The mottles are of green and olive green color and occur in the soil matrix and around burrow fillings and plant scars suggesting a periodic, possibly seasonal, waterlogging of the Morrison soil profiles.

The green mudstones of the Morrison Formation are similar to the red mudstones in texture, structure and bulk mineralogy except for the absence of hematite and the presence of ferrous iron (Table 2). The green mudstones grade to a gray color, with an increase in organic matter. This usually occurs where the mudstones are associated with channel fill sandstone units. The green mudstones contain abundant fossil remains of dinosaur vertibrate and have a higher percentage of organic carbon than the red mudstones (Table 2). The contact between the green and red mudstones is sharp and does not show any mottling.

Plant Roots

The occurrence of plant roots is a strong indication of paleosol existence regardless of their degree of pedological development (Retallack, 1983b). They extend downward from the green (A) horizons into the red (B) horizons. They form downward branching tubes penetrating the bedding at all angles (Figures 7a and 7b), and display irregular widths and directions. These tubes are filled with carbonate or very fine sediment (clay) and vary in diameter from less than 1mm to 2cm. No organic matter is preserved in the fossil roots due to diagenesis. Their penetration into the deeper parts of the B horizons indicates that the permanent water table was well below in most of these soils and that the water was available at depth to the larger roots during severe dry seasons (Retallack, 1983b). The reducing (green color) microenvironment around root scars is due to the decay of organic matter under anaerobic conditions after burial (Black, 1968).

Burrows and Bioturbation

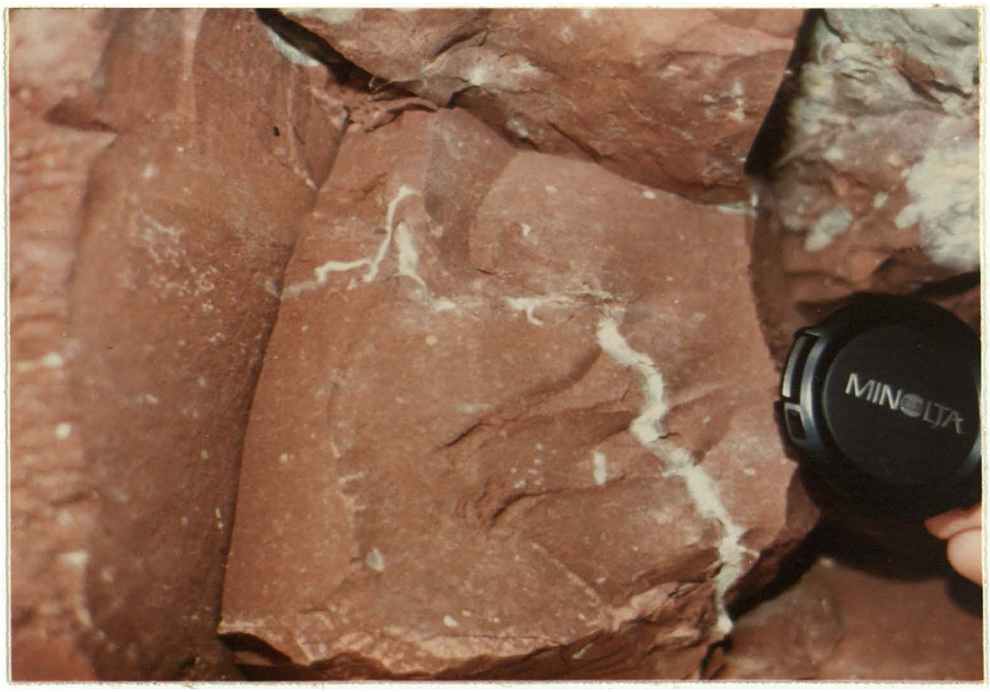
Burrowing is somewhat common in the Morrison Formation, as evidenced by a number of tubules with diameters up to 2cm. These tubules (Brewer, 1964) are believed to be fossil feeding and dwelling paths of soft-sediment burrowing organisms. They are usually vertical to subvertical and at least partly filled with carbonates and/or very fine sediment (clay). Bioturbation due to burrowing is evident mostly in the B horizons (Figures 8a and 8b). Burrows often occur in the deeper parts of the B horizons indicating a low water table (Retallack, 1983b).

Clay Illuviation - Cutans

Clay cutans are the result of mechanical infiltration (Teruggi and Andreis, 1971; Brewer, 1972). The term cutans was proposed by Brewer (1960) "for a broad group of pedological features, including the so called clay skins, associated with the surfaces of the skeleton grains, peds, and various kinds of voids within soil materials". They are classified on the basis of their composition and the type of surface affected (argillans, ferrans, calcitans etc.; Brewer, 1964). Their thickness varies, depending on the duration of

Figure 7a. Plant root from horizon 10 at Coyote Basin

Figure 7b. SEM photo of a plant root from horizon 16, Coyote Basin (scale: bottom side is 120u, u is micrometers)


MINOLTA

the interruption of sedimentation processes. In the Morrison Formation illuviation and complex cutans are the most common with thicknesses less than 1mm. Illuviation cutans (ferro-argillans) are found on grain surfaces (Figure 9a) and complex cutans are found on ped surfaces (Figure 9c). The latter are due both to illuviation and stress. They occur in the B horizons where clay was precipitated from solution and then moved downward with percolating water during wet seasons. During dry periods when water evaporation was high, clay covered pore and ped surfaces. The illuviation cutans formed when clay particles were deposited in the B horizons on the walls of voids or around grains and formed coverings of oriented clay platelets called argillans (visible in SEM photographs; Figure 9b). Complex cutans formed on ped surfaces due to shrink-swell processes.

Glaebules

These are pedological features described by Brewer and Sleeman (1964), characterized by their distinct boundary with the enclosing material (definition in Appendix A). They are classified based on: 1) internal fabric, 2) minaralogy, 3) distinctness as a unit, and 4) shape. The Morrison Formation mudstones contain glaebules (Figure 10a) which can be classified according to Brewer and Sleeman (1964) as follows: 1) consisting of undifferentiated fabric, 2) calcareous, 3) having sharp to rather sharp boundaries with the enclosing mudstones, and 4) consisting of mostly convolute shapes with less spherical and irregular shaped glaebules also being present. Based on these features the Morrison glaebules are classified as "nodules" formed in situ "by accretion of soluble constituents which accumulated by diffusion or crystallization from solution in matrix voids" (Brewer and Sleeman, 1964). The accretion took place during the begining of dry periods when solution concentrated in voids initiating their formation. Their sharp boundaries also indicates that they formed during dry periods (Drosdoff and Nikiforoff, 1940). The calcareous glaebules (immature stage of the cornstones of Allen, 1960) are calcium carbonate rich and occur mostly in the uppermost Morrison Formation indicating the existence of a drier climate (Reeves, 1970).

Pedotubules

These are pedological features defined by Brewer and Sleeman (1963) with tabular external form and complex internal composition and fabric (definition in Appendix A). They are classified according to: 1) their internal fabric and composition, 2) details of external form, 3) distinctness, and 4) by a comparison of their fabric and composition with that of the horizons of the soil profile. Their general morphology suggests that they formed as fillings in voids caused by faunal and floral (root) activity

Figure 8b. Burrow filled with carbonate, horizon 10, Coyote Basin

Figure 8a. Bioturbation with carbonate nodules and plant roots in horizon 10, Coyote Basin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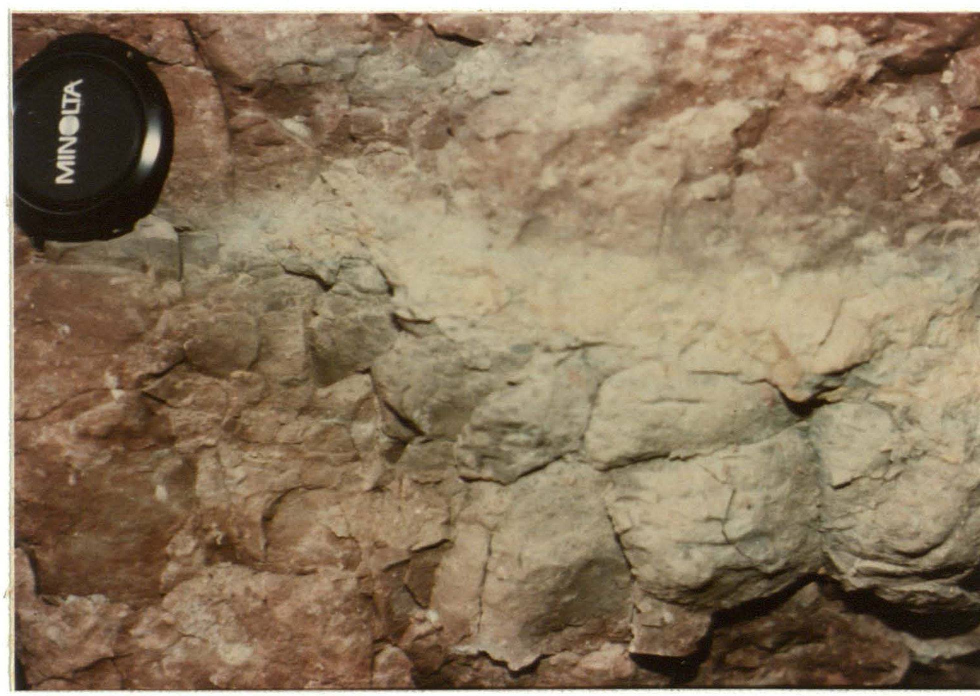
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Figure 9a. SEM photo showing clay surrounding a grain from horizon 5, Coyote Basin, the clay mineral is montmorillonite (scale: bottom side of photo is 50u)

Figure 9b. SEM photo showing oriented clay from horizon 10, Coyote Basin (scale: bottom side of photo is 8u)

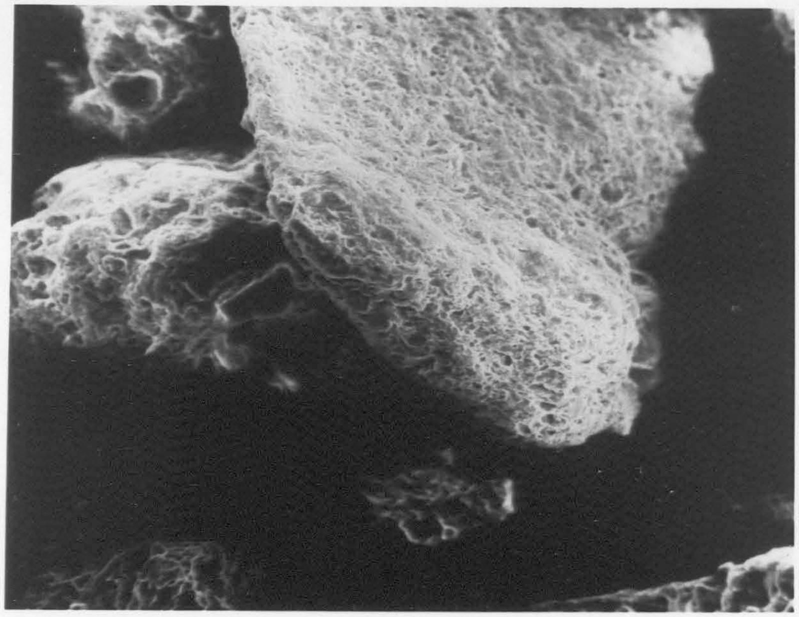
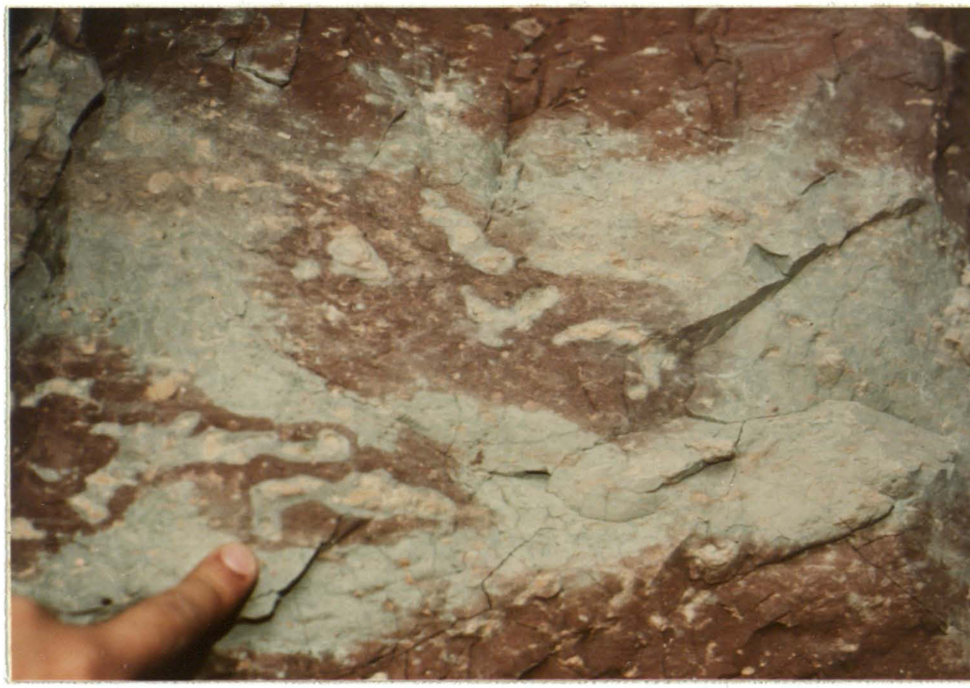


Figure 9c. Complex cutan on a ped surface from horizon 10, Coyote Basin; index finger at lower right corner for scale

Figure 10a. Glaebules, horizon 10, Coyote Basin

Figure 10b. Pedotubules at Coyote Basin scattered on the surface



(Brewer and Sleeman, 1963). In the Morrison Formation the pedotubules have an internal structure similar to the isotubules of Brewer and Sleeman (1963). They are composed of "skeleton grains and plasma that are not organized into recognizable aggregates and within which the basic fabric shows no directional arrangement with regard to external form". Their plasma is ferro-calcareous with a bright red color and they are characterized by a discrete distinctness. The material making up the pedotubules has been derived from the soil material of the horizon in which they occur, therefore are classified as orthotubules (Brewer and Sleeman, 1963). They occur in the red B horizons and cover the badlands of the Beaver Creek area due to the erosion (Figure 10b).

Cloverly Formation

The paleosols of the Cloverly Formation are not as well developed as those of the Morrison Formation, but possess pedogenic features that are similar to those of the Morrison. They will be described here using the background information presented for the pedogenic features of the Morrison Formation.

Soil Horizons

Both A and B master horizons are recognized in the paleosols of the Cloverly Formation. The playa mudstones of the Little Sheep Mudstone Member have almost uniform characteristics in all horizons. The A horizons are identified on the basis of higher organic carbon content (Table 3), lower iron and clay content, and presence of transitional contacts with the underlying B horizons (Soil Science Society of America, 1965). Their color falls into the 7.5YR3/0 to 7.5YR4/0 range of the Munsell Color Chart (1954). The B horizons are much thicker and show compound pedogenesis (Bown and Kraus, 1981a), that is incorporation of the A horizon of an older soil into the B horizon of a younger soil. This is the likely explanation of some multicolored mottled units with spherical glaebules (pedorelicts) discussed by Brammer (1968). Interbedded bentonite and devitrified tuff indicates that soil-forming processes were often interrupted. The overbank mudstones of the Little Sheep Mudstone Member are silmilar to those of the upper Morrison mudstones in that they are characterized by green (A) and red (B) horizons. The latter have a high carbonate content (Table 3) and are designated as (Bk) horizons. In the Himes Member, the overbank paleosols show only B horizons, with the A horizons absent probably due to erosion and subsequent sedimentation. They are densly mottled argillic (Bt) horizons. The uppermost Himes Member mudstones which immediately underlie the Sykes Mountain Formation are B horizons of olive gray (5Y5/1 to

5Y6/1) and blackish red (5R2/2) color with high clay and iron content (Table 3).

Color

The gray mudstones owe their color to high organic carbon content (McBride, 1974) which is at least twice as much as that in the green and red mudstones (Table 3). The iron content is also significantly less in the gray mudstones. The olive mudstones have a higher iron content than the gray and probably owe their color to the presence of bentonite. The cause of purple color is controversial (McBride, 1974). Ortlam (1971) ascribed the purple color "to reduction processes by the organic matter of plant or tree roots". Tree trunks (?) (Figure 11a and 11b) similar to those found in the Paleocene Fort Union Formation in the Bighorn Basin, Wyoming (Dr. Mary Kraus, Univ. of Colorado, 1986, pers. comm.) occur in the upper Little Sheep Mudstone Member in the Coyote Basin and Douglas Draw areas.

Plant roots and Burrows

Both plant roots and burrows occur in the Cloverly Formation but they are not as abundant as in the Morrison Formation. Plant roots occur throughout the Formation. They are usually filled with black clay (Figure 11c) and they branch in all different directions. Their diameter is variable, ranging from a few mm to 2cm. Where the A horizons are present, plant roots are more abundant

Figure 11a. Tree trunks (?) from the upper Little Sheep Mudstone Member at Douglas Draw, horizon 39 in Table 3

Figure 11b. Tree trunk in the upper Little Sheep Mudstone Member at Bear Creek, horizon 39

Figure 11c. Plant roots filled with dark clay from horizon 24, lower Little Sheep Mudstone Member, Coyote Basin.

penetrating down into the middle parts of the B horizons. Where the A horizons are absent, the plant roots are sparse due to the fact that sedimentation was active and vegetation was not able to be sustained. Burrows are vertical, up to 3cm in diameter, similar to those of the Morrison Formation.

Clay Illuviation - Cutans

The Little Sheep Mudstone Member is a clay-rich unit. Illuviation and stress cutans are the most common types. The former were formed during wet periods due to the movement of clay in solution and subsequent deposition around grains and voids. The stress cutans were formed during dry periods when desiccation occurred. Based on Brewer (1964) they may be classified as argillans and are common in the B horizons. Their thickness is variable, mostly less than 1mm. The Himes Member cutans are interpreted as complex (Brewer, 1964) because they occur on ped surfaces and their appearance is indicative of some local movement (stress cutans). Where clay and iron oxides are present they are classified as ferro-argillans (Brewer, 1964).

Vein Networks and Gilgai Topography

Vertical vein networks occur in the Little Sheep Mudstone Member. They extend downward 30 to 85cm and are up to 5cm thick. They are believed to have been formed as a result of alternating wetting and drying and are considered to be desiccation cracks (Figures 12a and 12b). Similar features have been described in present-day soils from playa-like depressions in southwestern United States by Neal and Motts (1967) and by White (1972) and are attributed to the intense evaporation and shrink-swell processes that occur in clay-rich soils. McPherson (1979) also described similar vein networks in the Upper Devonian fluvial overbank rebeds (Aztec Siltstone) in southern Victorialand, Antarctica. In the lower Little Sheep Mudstone Member the veins are filled with chalcedony and calcite (Figure 13a and 13b). The chalcedony was derived from the alteration of volcanic ash (Moberly, 1960). The calcite is at least partly pedogenic in origin having formed due to aridity of the climate at the time of deposition. The veins have a radiating patern and thin upsection. In the upper Little Sheep Mudstone (at Douglas Draw) the veins are filled with siltstone and sandstone.

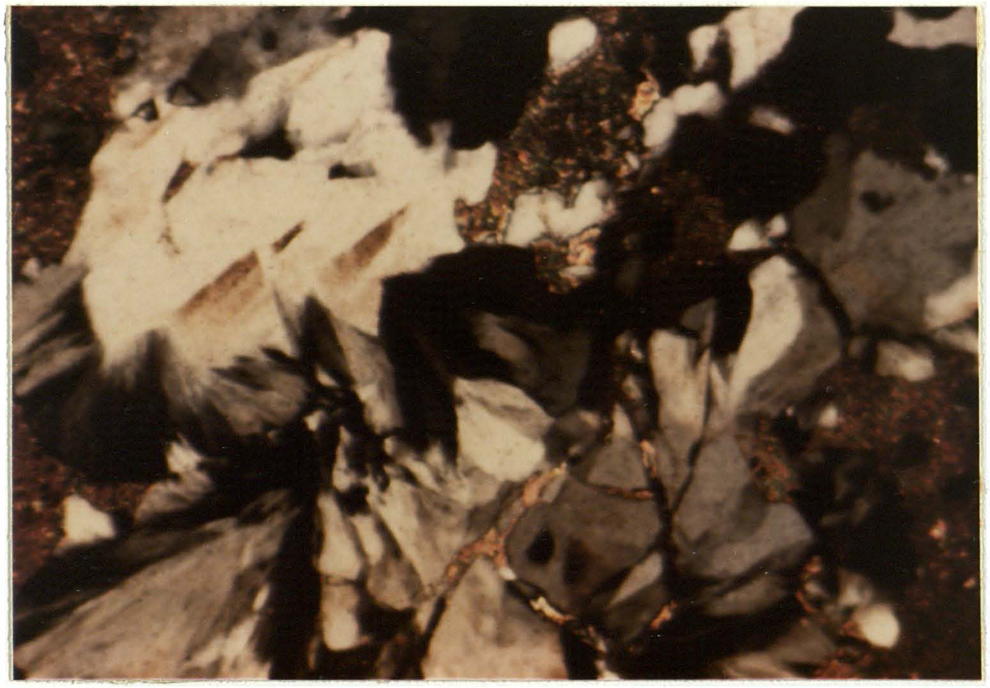
Gilgai topography forms as a result of the shrink-swell processes that take place in clay-rich soils during alternating wet-dry periods. The resulting landscape is that of an undulating topography of highs and lows. The resulting depressions were filled with volcanic ash during the volcanic eruptions to the west. Gilgai topography is characteristic of Vertisols (Soil Survey Staff, 1975) and will be discussed in more detail below.

Figure 12a. Desiccation cracks, horizon 24, lower Little Sheep Mudstone Member, Coyote Basin; width of the cracks is 5cm.

Figure 12b. Closer view of a desiccation crack of figure 12a; width of the crack is 5cm.

Figure 13a. Thin-section showing replacement of chalcedony by calcite crystals (scale: bottom side is 200um (um is microns))

Figure 13b. Closer view of calcite replacing the chalcedony (scale: bottom side is 500um)



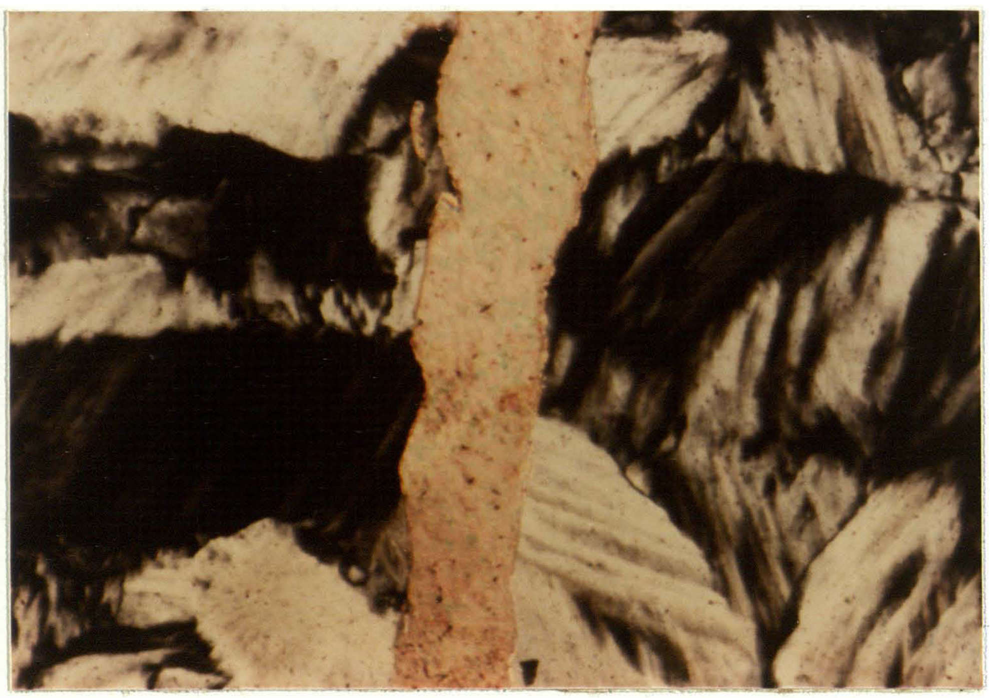


Figure 14. Chalcedony nodule (silcrete ?), the horizontal diameter of the nodule is 10cm, from horizon 25, Coyote Basin

Figure 15. Septarian nodule with coarse calcite crystals, horizon 24, Coyote Basin

Nodules

Calcareous and silica nodules are common features in the mudstones of the Cloverly Formation. They occur in distinct horizons throughout the host rock and vary in size from a few mm to tens of cm.

Two kinds of silica nodules, probably reflecting different depositional environments, commonly occur in the Little Sheep Mudstone Member. In the lower Little Sheep Mudstone unit chalcedony nodules are common. These are white rounded subspherical nodules (Figure 14) that occur where bentonite (product of the alteration of volcanic ash) is most abundant. They formed as silica solubility was lowered as a result of decreasing pH (Hay, 1968). A detailed geochemical analysis of these nodules will provide data for more definite conclusions. Silica precipitation requires alkaline conditions but it is not easy to establish the timing of silica diagenesis (Smale, 1973). Silica nodules in some sequences of continental origin compare with silcrete nodules of present-day soils and are thought to have formed at an early diagenetic stage (Summerfield, 1983; 1984). Similar nodules have been described by Mountain (1951), Hutton, Twidale and Milnes (1978), Senior (1978), and Wopfner (1983). Their origin is controversial. Alteration of the heavy minerals which occur in volcanic ash may be the possible explanation of silica nodule formation (Jenkins, 1985). The second kind of silica nodules occurs in the upper Little Sheep Mudstone and is discussed below.

Septarian nodules (Figure 15) occur in the lower Little Sheep Mustone Member and can be traced laterally for hundreds of feet. They show a pattern of irregular lenticular cracks filled with coarse calcite crystals (and/or rarely barite crystals also found). The cracking reflects a synaeresis-type contraction due to the dewatering of a gel-like mass of clay minerals (Collinson and Thompson, 1982). Synaeresis is a process during which "subaqueous shrinkage cracks result from loss of pore water from the sediment because of a reorganization of originally highly porous clays, either due to flocculations or because of salinity-induced changes of volume of certain clay minerals" (Collinson and Thompson, 1982). The latter seems to be the case for the Little Sheep Mudstone Member.

The upper Little Sheep Mudstone Member is characterized by stratigraphically persistent chert and carbonate nodules which can be traced for kilometers (Figures 16a and 16b). They are present at the Coyote Basin, Douglas Draw and Sheep Mountain Anticline localities but they are ubiquitous in the Bighorn Basin. The chert is purple and white in color and shows cracks which are filled with calcium carbonate (Figures 17a and 17b). These cracks were formed during

Figure 16a. Chert-carbonate nodules from Coyote Basin

Figure 16b. Chert-carbonate nodules at South Nose of Sheep Mountain Anticline

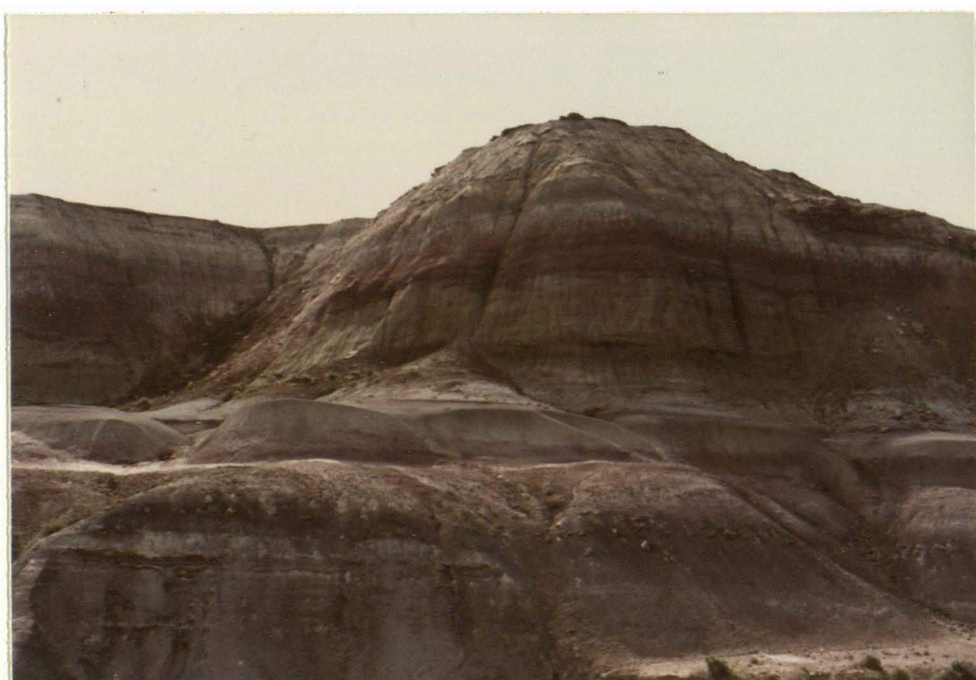
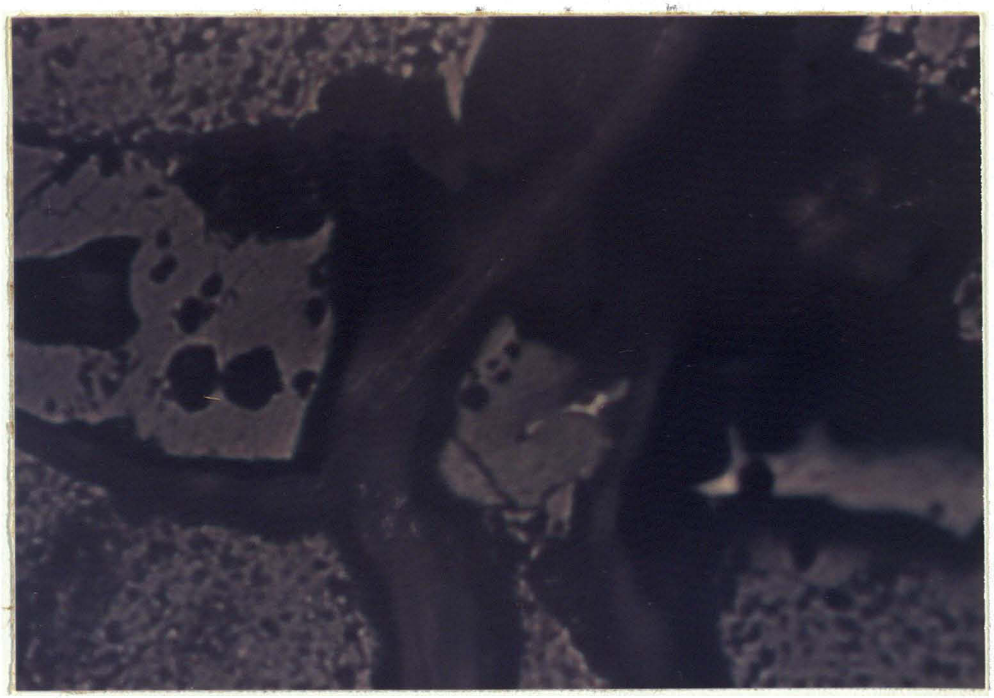


Figure 17a. Chert-carbonate relationship from polished section, horizon 37, Coyote Basin (scale: bottom sibe of the photo is 200um)

Figure 17b. Polished sample showing the relationship between chert and carbonate, horizon 37, Coyote Basin (scale: bottom side of the sample is 6.5cm)



subaerial exposure and were then filled with calcium carbonate of pedogenic origin (Mantzios and Vondra, 1986b). Similar chert has been reported by Surdam, Eugster and Mariner (1972) in the Morrison Formation in southern Wyoming as well as in Eocene through Pleistocene rocks in Wyoming. This chert called Magadi-type chert By Surdam, Eugster and Mariner (1972) was formed from a hydrous sodium-silicate precursor such as magadiite. The depositional environment and the mode of origin of this chert will be discussed below. Calcium carbonate accumulated as coating on the chert surface and as crack filling. With time the carbonate deposition plugged the chert horizon and its build up continued to variable degree as its thickness is not constant throughout.

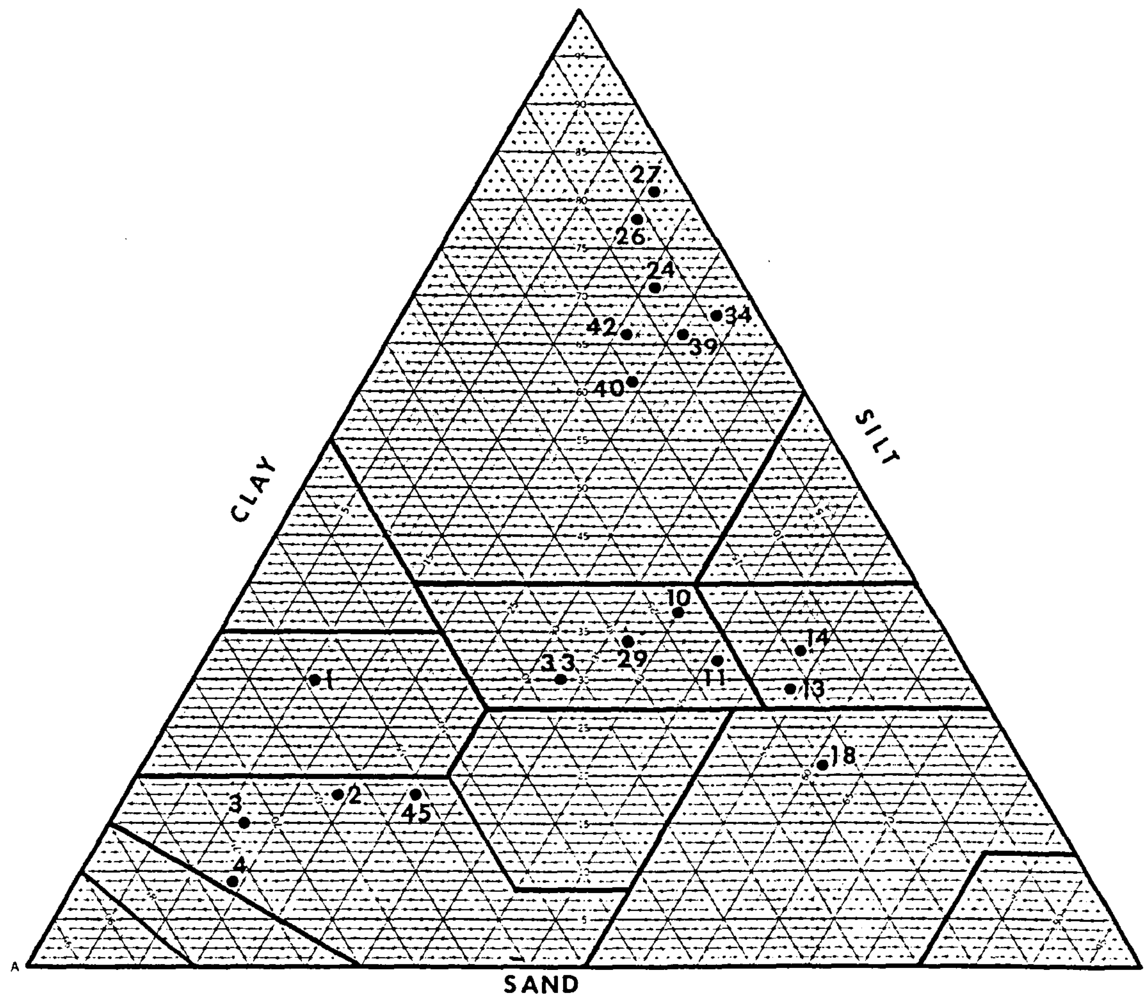
PETROGRAPHY

Mudstones

The mudstones constitute at least 50% of the Morrison Formation. They are calcareous, usually massive, often bioturbated siltstones to silty claystones (Folk, 1968) with a small amount of sand and (or) gravel present (Figure 18) and contain calcareous nodules of pedogenic origin. The mudstones were classified using the Soil Science Society of America (1965) guide for textural classification of soils. The mudstones also constitute at least 65% of the Cloverly Formation. The Little Sheep Mudstone Member is characterized by non-calcareous, locally fissile, siltstones and claystones (Figure 18). They are bentonitic (montmorillonite-rich), organic-rich and show desiccation cracks which were formed due to subaerial exposure. The mudstones of the Himes Member are massive, densely mottled siltstones. Their textural classification is shown in Figure 18. The textural analysis was performed by mechanical analysis using sieve and pipette methods.

Visual aids (petrological and binocular microscopes) were used for textural and compositional examination of the mudstones. Thin sections were rather limited but grain mounts of 30 samples were studied for mineralogical analysis. The most abundant mineral for both formations is

Figure 18. Textural classification of the Morrison and Cloverly formations. Numbers correspond to soil horizons in Tables 2 and 3


CLAY
SILT
SAND
A
27
26
24
34
42
39
40
10
14
33
29
11
13
1
18
3
2
45
4

quartz but chert, calcite, hematite and feldspars (mostly plagioclase and microcline) are also present. Volcanic contributions are not very common in the Morrison Formation, but tourmaline, epidote, fluorite (Figure 19a) and zircon (Figure 19b) were found. The Little Sheep Mudstone Member is rich in bentonite and devitrified tuff. Biotite, apatite, zircon, garnet and sphene are also found. Chert (Figure 20a) and feldspars (plagioclase and microcline) are present (Figures 19c and 19d) but quartz is again the most abundant mineral occurring in polycrystalline, monocrystalline as well as in overgrowth form (Figures 20b-d). Chalcedony (Figure 14a and 14b) is abundant throughout the Cloverly and is probably associated with the volcanic ash. The volcanic ash was suggested by Moberly (1960) as having its source to the west where volcanic activity had increased from the Late Jurassic to the Early Cretaceous. Calcite and silica nodules found throughout the Cloverly Formation will be discussed below.

Clay Minerals

X-ray diffraction was performed on 28 samples. Similar analyses have been reported for both the Morrison and Cloverly formations by Moberly (1960) and Kvale (1986). All three independent analyses agree on the results. The Morrison Formation consists of illite, montmorillonite,

Figure 19a. Fluorite under plain polarized light (scale: bottom side is 200um)

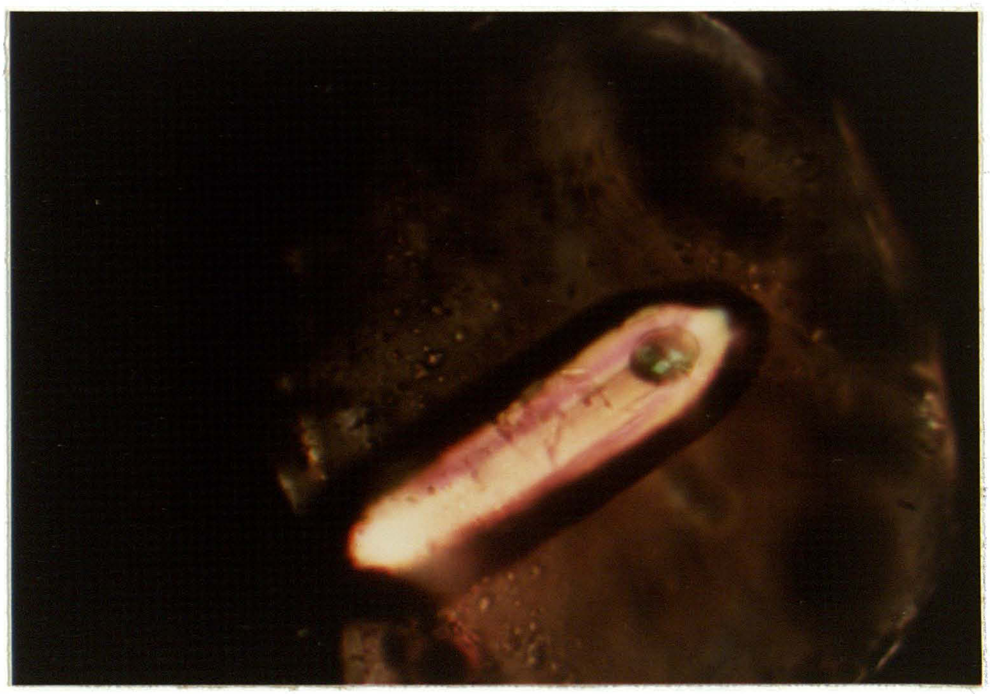


Figure 19b. Zircon (scale: bottom side is 200um)

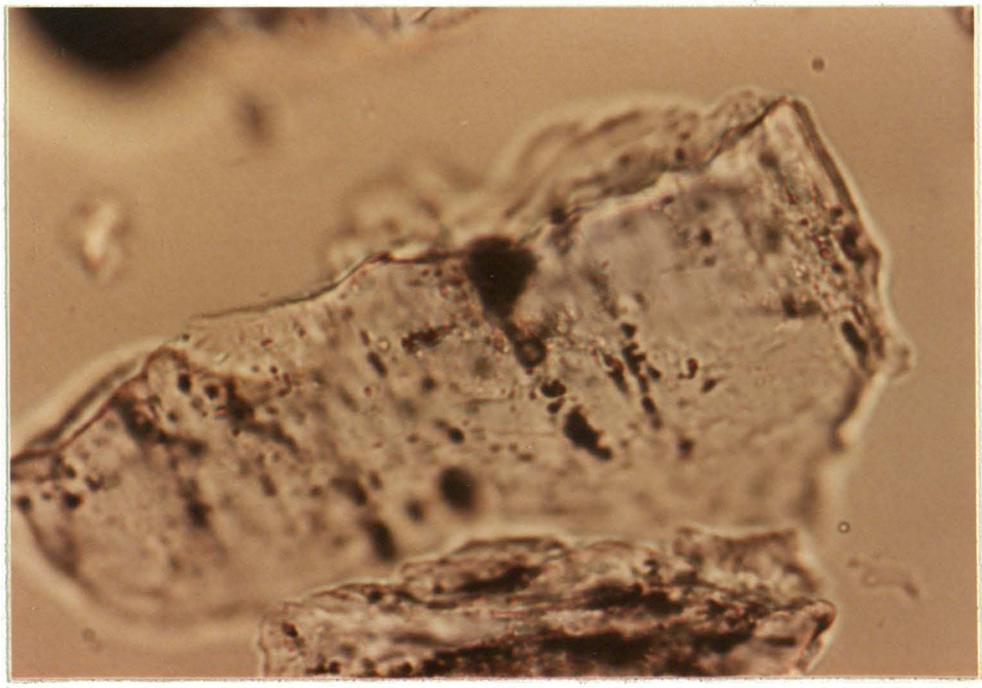


Figure 19c. Plagioclase under plain polarized light (scale: bottom side is 200um)

Figure 19d. Microcline (scale: bottom side is 200um)

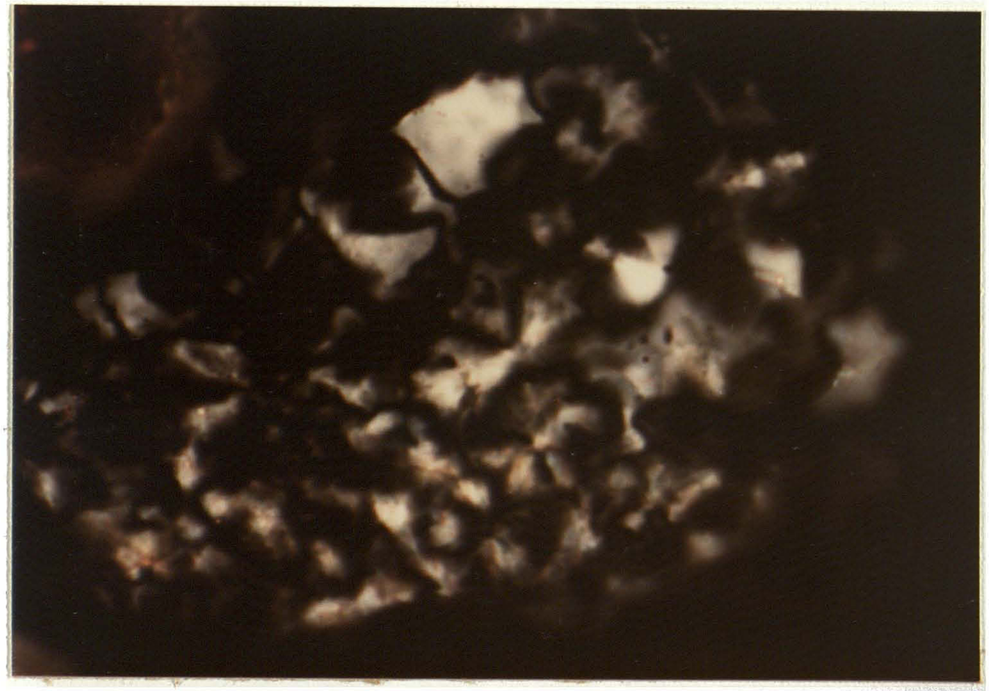


Figure 20a. Chert (scale: bottom side is 200um)

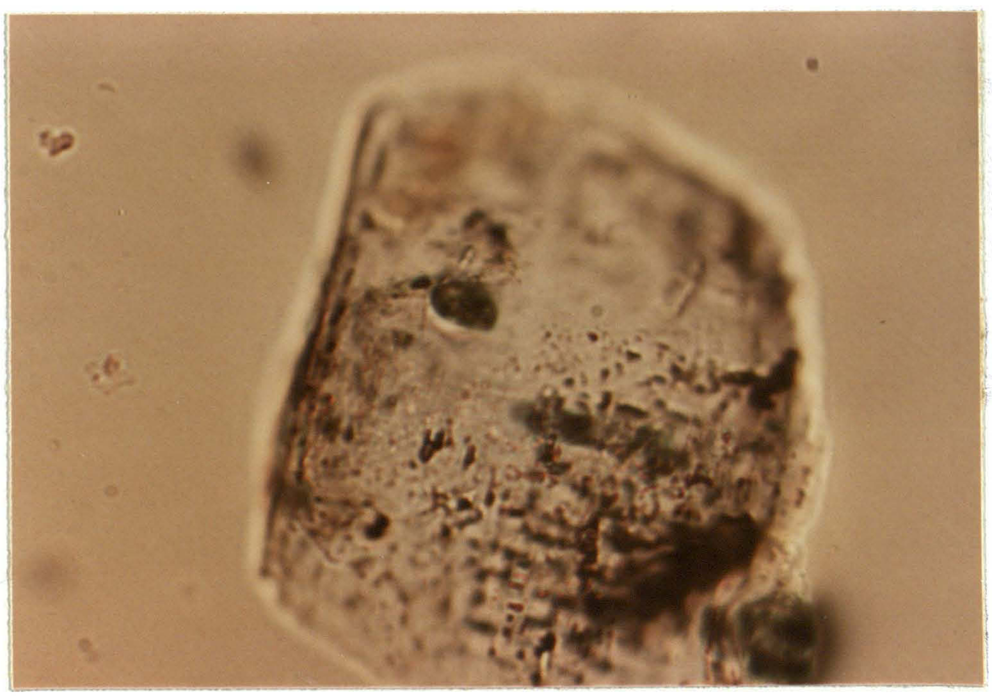


Figure 20b. Quartz, well rounded with inclusions under plain polarized light (scale: bottom side is 200um)

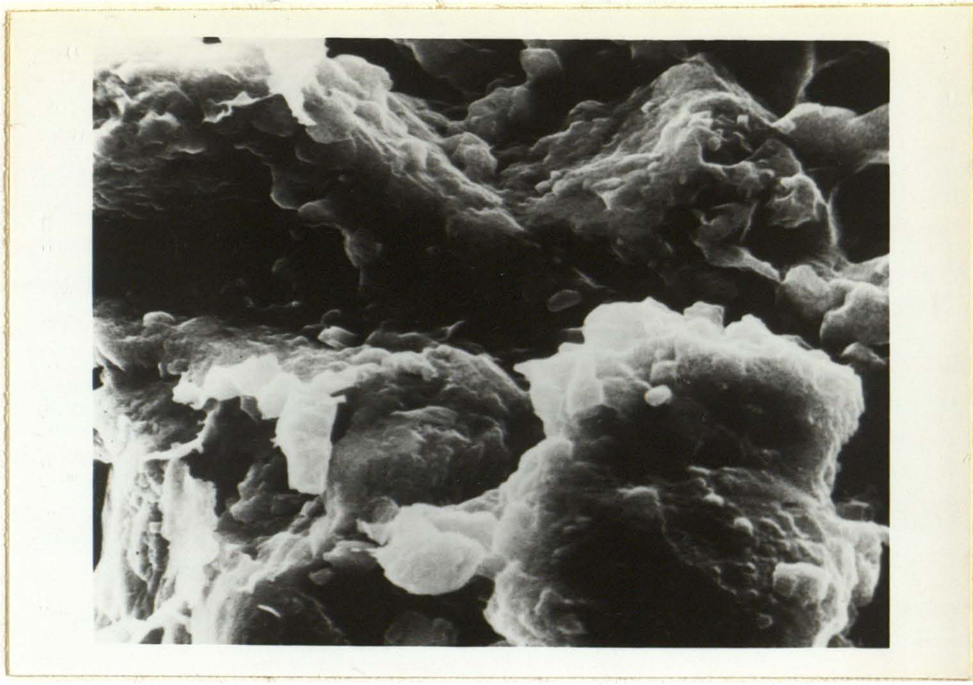


Figure 20c. SEM photo showing quartz overgrowth (scale: bottom side is 40u)

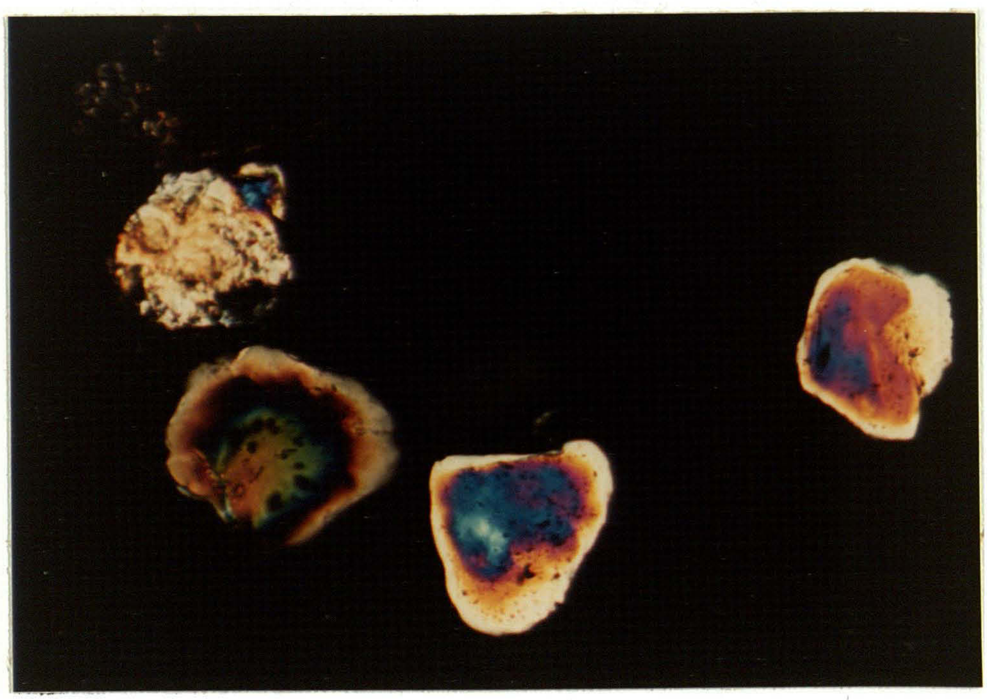


Figure 20d. Quartz crystals with inclusions (scale: bottom side is 200um)

Table 1. Clay Mineralogy, Il-illite, Mo-montmorillonite, Il-Mo-mixed-layer illite-montmorillonite, Ka-kaolinite

Formation	Sample Number	Clay Minerals
Morrison	BC6228502	Il, Mo, Il-Mo, Ka
Morrison	BC6228503	Il, Mo
Morrison	BC6228504	Il, Mo, Il-Mo, Ka
Morrison	BC6228505	Il, Il-Mo, Mo
Morrison	BC7168501	Il, Mo, Il-Mo
Morrison	BC7168502	Il, Mo, Il-Mo
l. LSM	BC7178506	Mo, Il-Mo, Il
l. LSM	BC7188502	Mo, Il-Mo
l. LSM	BC7188503	Mo. Il-Mo
l. LSM	BC7188513	Mo, Il-Mo
l. LSM	BC7208504	Mo, Il-Mo, Il
l. LSM	BC7208507	Mo, Il-Mo, Il
m. LSM	BC7208511	Il, Il-Mo, Mo
m. LSM	BC7208514	Il, Il-Mo, Mo
u. LSM	BC8018503	Mo, Il-Mo, Il
u. LSM	BC8018504	Mo, Il-Mo
u. LSM	BC8018506	Mo, Il-Mo
u. LSM	BC8018508	Mo, Il-Mo, Il
Himes	SSM7018503	Mo, Il-Mo, Il
Himes	SSM7018504	Mo, Il-Mo, Il
Himes	SSM7018505	Mo, Il-Mo, Il
Himes	RD6148502	Il, Il-Mo, Mo

mixed-layer illite-montmorillonite and subordinate kaolinite while the Cloverly Formation is predominantly rich in montmorillonite, mixed-layer illite-montmorillonite and illite (Table 1; Mantzios and Vondra, 1986b). Singer (1980, 1984) attempted to relate clay minerals and paleoclimate but he cautioned the reader regarding the sole use of clay minerals as paleoclimatic indicators. His studies indicated however, that clay minerals can be used in combination with pedogenic features for paleoclimatic interpretations. Retallack (1981) noted that "... clay minerals are notoriously susceptible to diagenetic alterations of such a scope and complexity (Millot, 1970) that it is very difficult to be certain of the original clay mineralogy of paleosols". Other scientists have also noted the difficulty in using clay minerals as tools for various kinds of geologic interpretations (Weaver, 1967; Wilson and Pittman, 1977; Millot, 1977). Birkeland (1984) indicated that soils are characterized by several clay mineral species which are formed either due to weathering of primary non-phyllosilicate minerals during which cations are released and recombined to form phyllosilicate clay minerals, or due to weathering of primary phyllosilicates and alteration in the solid state (this process is called transformation). Dunoyer De Segonzac (1970), Reynolds and Hower (1970) and Rettke (1981) discuss clay mineral diagenesis. The clay minerals Morrison Formation at the time of deposition had to be predominantly montmorillonite considering the input of the volcanic ash from the west. Illite and mixed-layer illite-montmorillonite probably formed as alteration products of the montmorillonite. The subordinate kaolinite is probably the product of feldspar weathering. In the Cloverly Formation where montmorillonite is the predominant clay mineral, the alteration product is mostly mixed-layer illite-montmorillonite with less illite.

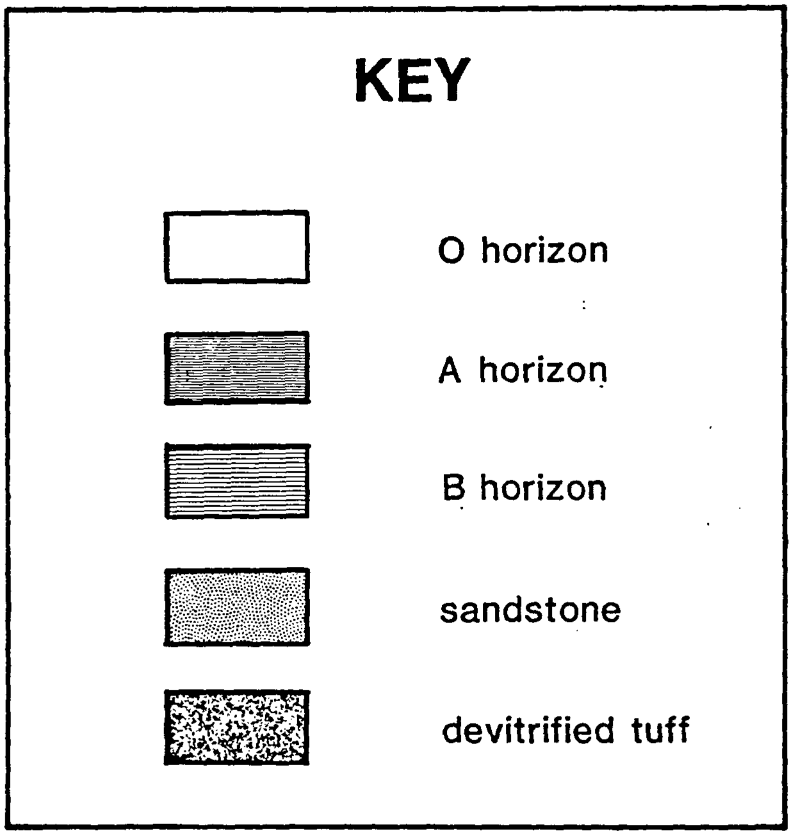
PALEOSOL INTERPRETATION

Paleosol orders are based on the modern soil taxonomy (Soil Survey Staff, 1975). During the last ten years researchers have been able to identify paleosol orders even though some of the criteria used to classify modern soil such as original soil pH and Eh, soil temperature, vegetation, moisture content and rainfall regime, and base saturarion are not available for the study of paleosols (Bown and Kraus, 1981a).

There are ten soil orders recognized in the modern soil classification. The three orders (Spodosols, Aridisols and Vertisols) suggested for the Morrison and Cloverly paleosols occur today in semiarid to arid climatic regions with the Spodosols being rather transitional to Aridisols (Dregne, 1976). Their recognition is based on both field and laboratory observations but cannot be as conclusive as in modern soils for the reasons mentioned above and because of diagenesis which altered other usefull features.

The first order recognized in the Morrison Formation includes the paleosols developed on floodplains associated with meandering channels. These are Spodosols characterized by accumulations of sesquioxides and organic matter complexes. The processes of Spodosol formation are summarized by Buol, Hole and McCracken (1973) as follows: 1) accumulation of organic matter, 2) leaching and acidification, 3) weathering, 4) translocation of Fe and Al (with some Mn and clay) from the A to B horizon, 5) imobilization of humic and fulvic acids (and some clay) in the B, 6) pelleting of humus coating, 7) reduction in bulk density, and 8) cementation. The Morrison Spodosols show an upper horizon rich in organic carbon content (Table 2) which is interpreted as an O horizon. This horizon is light gray (7.5YR7/0) in color and has a thickness of 10cm or less. It directly underlies channel sandstones. Directly below is an A horizon of greenish gray (5GY6/1) color whose thickness is 45cm. The organic carbon content of the A horizon is about half that of the O horizon. A spodic (Bs) horizon lies between the A and an argillic (Bt) horizon and is rich in sesquioxides and clay which were translocated from the upper horizons (O and A). It displays mottling and contains glaebules in the lower portion of the horizon, and is low in organic carbon content. These Spodosols have a high iron to organic carbon ratio (12:1) and this is characteristic of the suborder Ferrods which in modern soils are not very wet (freely drained) soils, indicate some increase in dry periods, probably one to two months, and whose iron to organic carbon ratio must be greater than 6:1. The texture of the Spodosols is sandy to loamy (Figure 18a).

The second order of paleosols recognized in the Morrison Formation is the Aridisols. These have thin A



EXPLANATION

Total Iron: the sum of ferric and ferrous iron

Ferric Iron: Fe_2O_3

Ferrous Iron: FeO

Aluminum: Al_2O_3

Table 2. Elemental analyses' data of the Morrison Formation

Table 3: Elemental analysis' data of the Cloverly Formation, explanation and key as in Table 2

MORRISON FORMATION

	THICKNESS (cm)	SAMPLE NUMBER	HORIZON	COLOR	TOTAL IRON	FERROUS IRON	FERRIC IRON	ALUMINUM	ORGANIC CARBON	CALCIUM CARB.
20	40	BC7168508	A	10Y6/2					0.33	
19	95	BC7168507	Btk	5YR3/2						19.5
18	70	BC7168506	A	5GY3/2					0.39	
17	85	BC7168505	Btk	5YR3/2	3.75	1.02	2.73			17.3
16	170	BC7168504	Btk	10R4/3						15.6
15	45	BC7168503	A	5GY3/2					0.46	4.4
14	55	BC7168502	Btk	10R4/2						17.4
13	40	BC7168501	Btk	5YR3/2	4.01	0.73	3.28	0.12		18.3
12	40	BC6228512								
11	55	BC6228511	Btk	10R5/6						16.8
10	250	BC6228510	Btk	10R3/4	4.12	1.14	2.98			17.1
9	70	BC6228509								
8	10	BC6228508	O	7.5YR7/0					1.16	
7	95	BC6228507	A	5GY6/1	4.17	3.19	0.98		0.51	
6	85	BC6228506	Bs	10R4/2	3.65	0.71	2.94	0.73		11.6
5	140	BC6228505	Bt	5YR4/2				0.23		12.3
4	10	BC6228504	O	7.5YR7/0					1.07	
3	45	BC6228503	A	5GY6/1					0.44	
2	95	BC6228502	Bs	10R4/3	3.35	1.02	2.33	0.87	0.19	12.1
1	90	BC6228501	Bt	5YR4/2	3.73	1.45	2.28	0.20	0.10	10.8

CLOVERLY FORMATION

	THICKNESS (cm)	SAMPLE NUMBER	HORIZON	COLOR	TOTAL IRON	FERROUS IRON	FERRIC IRON	ALUMINUM	ORGANIC CARBON	CALCIUM CARB.
46	50	RD6148504		5Y5/2						
45	80	RD6148503		5R2/2						10.1
44	210	RD6148502		5Y6/1						
43	145	RD6148501		5Y5/1 10R3/4						
42	365	SSM7018505	Bt	5R4/6 7.5YR5/0	3.06	1.38	1.68			10.6
41	160	SSM7018504	Bt	10R3/4 5GY6/1						
40	70	SSM7018503	Bt	10R3/4 5GY8/1						
39	100	BC8018508	A	7.5YR5/0	1.61	0.90	0.71		0.98	10.3
38	300	BC8018507								
37	110	BC8018506	Bt	5Y4/1						11.5
36	60	BC8018505	Bt	5Y6/1						
35	35	BC8018504	A	7.5YR4/0						
34	195	BC8018503	Bt	5RP4/2	2.31	0.95	1.36		0.31	12.1
33	90	BC7208514	A	10Y6/2	3.00	2.19	0.81		0.41	
32	105	BC7208513	Btk	5R3/4						15.4
31	90	BC7208512 BC7208511	Btk	10R4/2	3.40	1.02	2.38	0.06		16.1
30	100	BC7208510	Btk	5R3/4						16.2
29	210	BC7208509 BC7208508	Btk	10R4/2						16.5
28	315	BC7208507 BC7208506 BC7208505 BC7208504	Bt	5Y4/1					0.79	8.5
27	85	BC7188513	A	7.5YR3/0					1.21	6.1
26	200	BC7188511 BC7188510 BC7188509	Bt	5Y4/1	2.03	1.15	0.88		0.65	7.7
25	150	BC7188508 BC7188507 BC7188506 BC7188505	Bt	5Y5/2						8.1
24	520	BC7188503 BC7188502 BC7188501 BC7178507 BC7178506 BC7178505	Bt	5Y6/1						8.7
23	35	BC7178504	A	7.5YR4/0					1.13	5.8
22	60	BC7178503	Bt	5YR4/1	2.83	1.39	1.44		0.49	9.6
21	45	BC7178502	Bt	5YR4/1						10.1

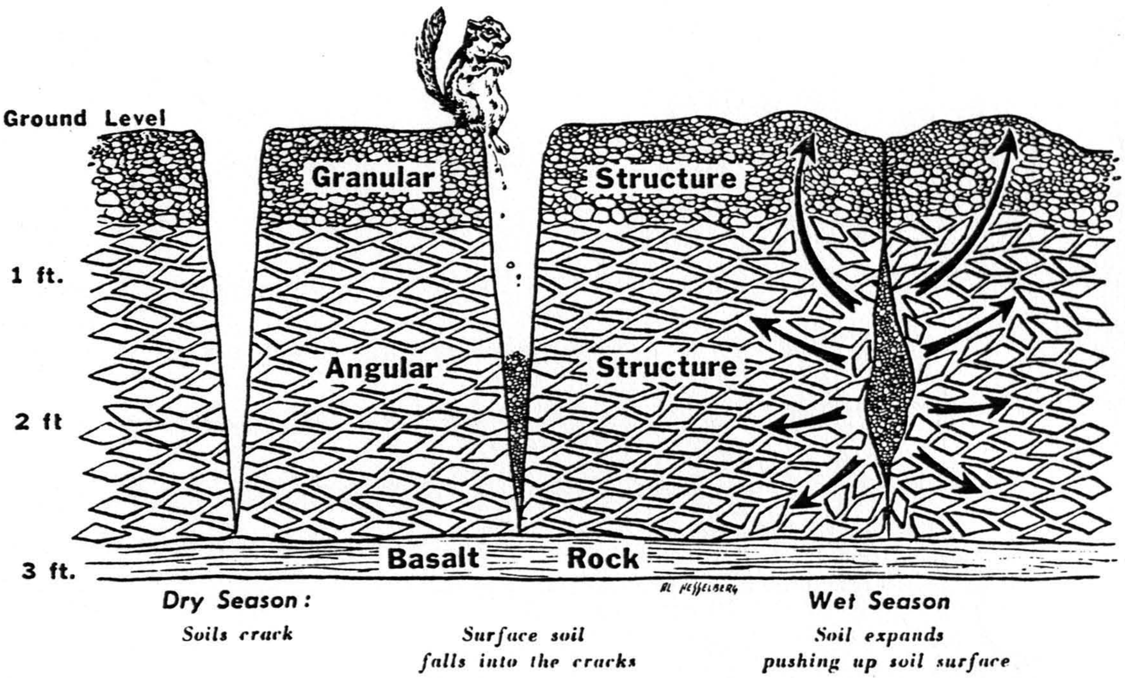
horizons of low organic carbon content and are classified as ochric epipedons. The B horizons are rich in clay and calcium carbonate and are classified as argillic (Bt), calcic (Bk) or as a combination of the two (Btk). The Aridisols are well oxidized soils as is evident by their low organic carbon content (Table 2). The process by which the argillic horizons formed is by translocation of clay from the surface horizons eventhough in situ clay formation is suspected due to the aridity of the paleoclimate and the relatively high amount of clay present in the argillic horizons. Both of these processes of clay formation are common in modern Aridisols (Buol, Hole and McCracken, 1973). The most striking feature of the Aridisols is the carbonate accumulation. In the Morrison Formation the carbonate is found in the form of glaebules and pedotubules. Plant roots penetrate down to the lower portions of the B horizons indicating that water was available during dry seasons. The presence of the argillic horizons classifies the Aridisols as Argids (suborder) and Haplargids (great group).

The third paleosol order, the Vertisols, occurs in the Cloverly Formation. These are clay-textured, dark-colored soils with a low level of profile development (immature soils). Modern Vertisols develop on both lacustrine and alluvial deposits, and usually occupy depressions (playas) or broad alluvial plains having minimal runoff. They are rich in smectite clay minerals which are either inherited or associated with volcanic ash. Gilgai microrelief, cracking, and slickensides are associated with smectitic mineralogy as well as the amount of clay. The most unique feature of Vertisols is their ability to shrink and crack as drying progresses, and then to swell and become very plastic and cohesive on wetting (Figure 21). The extent and nature of shrinking and swelling are functions of clay content, clay type, soil texture, degree of hydration, absorbed cations, rainfall distribution and natural vegetation (Birkeland, 1984).

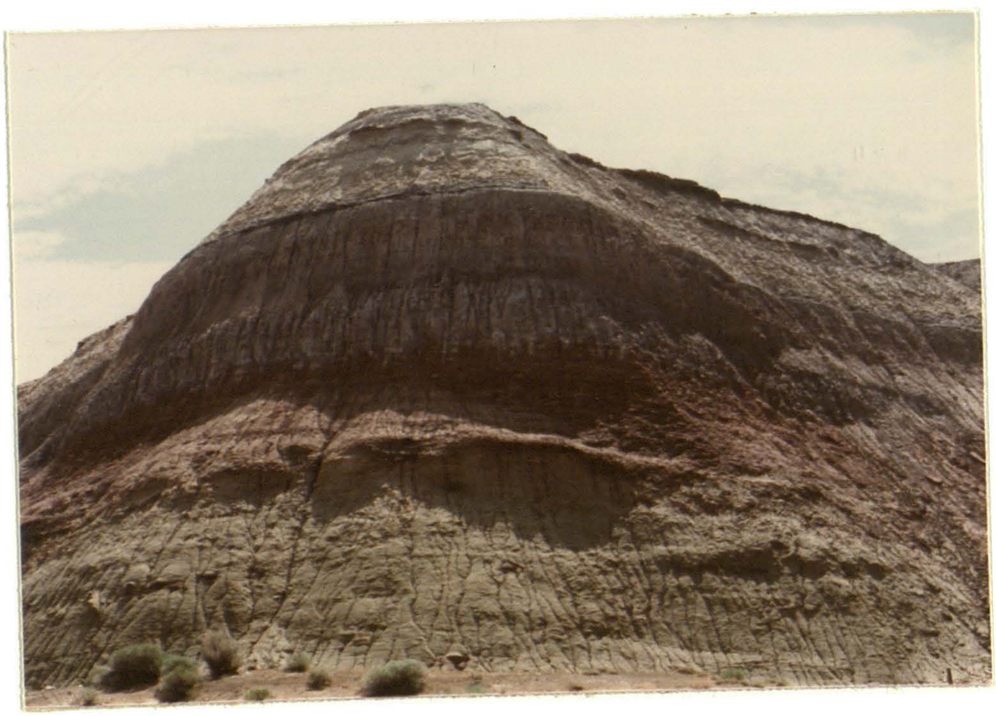
The Vertisols of the Little Sheep Mudstone Member are immature, clay-rich, with horizonation so weakly expressed that the profiles appear to be the same throughout because of the shrink-swell processes. Textural differences in the profiles are minimal, therefore horizon differentiation is primarily based upon color and chemical content. Both A and B horizons are recognized. The A horizons show plant roots and desiccation cracks that penetrate down in to the B horizons. Gilgai microrelief is difficult to distinguish because the sediments overlying have more or less destroyed it. Mottling in the lower horizons suggests a periodic, seasonal waterlogging. The Vertisols of the lower Little Sheep Mudstone Member Vertisols show at least three different soil profiles superimposed on top of each other.

Figure 21. Sketch illustrating the effect of wet and dry cycles

Figure 22. The lower Himes Vertisols from the South Nose of Sheep Mountain Anticline area



(from Buol, Hole and McCracken, 1980)



They are classified as Torrets. The Upper Little Sheep Mudstone Member Vertisols show thicker horizons and less well developed pedogenic features. Tree trunks occur at the base of the thick devitrified tuff beds. The Himes Member Vertisols (Figure 22) are densely mottled (also suggesting a seasonal waterlogging), clay rich, with slickensides along ped surfaces and thick superimposed argillic (Bt) horizons. The A horizon is absent probably due to truncation by the Bt horizons or due to erosion by the overlying Greybull channels. They are classified as Usterts. Finally, the paleosols of the uppermost Himes Member are not assigned to a soil order due to their poor development. The transgression of the Sykes Mountain seas probably destroyed the soils and their pedogenic features.

PALEOCLIMATIC INTERPRETATION

Based on the paleosol orders just discussed, a paleoclimatic interpretation of the Morrison and Cloverly formations can be attempted. Other properties that correlate with climate and are taken into consideration here are organic matter content, clay content, kind of clay minerals and iron content, color, and the presence or absence of salts along with their depth in the soil horizons.

Spodosols of the present day form in a variety of climatic zones. Their features in the Morrison Formation are similar to those of the Eocene Willwood Formation in the Bighorn Basin, Wyoming (Bown and Kraus, 1981a, 1981b). The fewer calcium carbonate glaebules, fewer red beds, the sesquioxide and clay accumulation in the subsurface horizons and their higher organic carbon content indicate a wet climate with seasonal dry periods as compared to that under which the Aridisols were formed which was a dry climate, with seasonal wet periods. This is indicated by the increase of red beds, calcium carbonate and clay content in the Aridisols (Machette, 1985; McFadden and Tinsley, 1985).

It is suggested that the Morrison climate was a subhumid-subtropical climate that changed progressively towards a semiarid-arid climate. The higher organic carbon content of the Spodosols indicates that vegetation was abundant. Dinosaur tracks were found on a splay sandstone

(Figures 23a and 23b) suggesting that they lived around streams and lakes where vegetation was abundant.

The minerals present in the mudstones can also provide clues regarding the paleoclimate of the Morrison Formation. The presence of feldspars in the Aridisols and their absence from the Spodosols supports the conclusion that an increase in aridity took place towards the end of Morrison time.

Modern Spodosols are characterized only by mixed-layer clay minerals while modern Aridisols are characterized mostly by smectite clays but mixed-layer, illite and kaolinite may be present. The clay mineral analysis performed on the Morrison paleosols does not exactly agree with their occurrence in modern soils undoubtedly because they have been altered by diagenesis.

The Vertisols which are recognized in the Cloverly Formation suggest a semiarid climate (Blodgett, 1985). Alternating dry and wet periods are indicated by the desiccation features. The aridity decreased considerably from the lower to the upper part of the Little Sheep Mudstone Member. This is indicated by the aerial extent of the lacustrine deposits and the vegetation (tree trunks). During the deposition of the Himes Member the climate continued to be semiarid but the presence of crocodile and turtle bones along with the darker (gray) colors of the mudstones suggests a change to a more humid-tropical climate.

Figure 23a. Dinosaur tracks on a splay sandstone (horizon 9) at Coyote Basin

Figure 23b. Closer view of a dinosaur track of figure 23a

DEPOSITIONAL ENVIRONMENTS AND PALEOTOPOGRAPHY

The Morrison paleosols are mature paleosols which were developed on alluvial parent material. They are interpreted as vertically accreted overbank sediments which were laid down on broad interchannel floodbasin areas. The subaerial exposure of these sediments for extended periods of time resulted in situ modification by pedogenic processes.

The Spodosols were developed on the proximal areas of the floodplain which was associated with meandering channels. The Aridisols represent distal floodplain paleosols. During wet periods when sedimentation rates were increased, A horizons were truncated and incorporated into new B horizons. This may explain the great thicknesses of some of the B horizons. Floral evidence is not available for paleobotanical interpretations and a better understanding of the paleotopography. Dinosaur bones are less abundant in the Aridisols which may be indicative of sparse vegetation as compared with that of the Spodosol intervals. The soil horizons however, are laterally extensive in both the Spodosols and Aridisols. There are at least four different Aridisol profiles superimposed on top of each other which show the same degree of maturity indicating no major shifting of the channels associated with these floodplain deposits. The pedogenic carbonates present can yield information on accretion rates if studied on a regional scale (Leeder, 1975).

The Little Sheep Mudstone Member of the Cloverly Formation resembles the Eocene Green River Formation of Utah, Colorado and Wyoming which is the largest and best studied lacustrine deposit in the world (Eugster and Hardie, 1978). The lower Little Sheep Mudstone unit is interpreted as an ephemeral or playa lake deposit. Its aerial extent is considerably less than that of the upper unit. The water that fed this lake came from surface run-off indicated by thin sheet-like sandstones which are interbedded with the mudstones, and perhaps from ground water which might have been delivered by springs. During seasonal drying periods desiccation at the surface of the lake occurred (Figures 12a and 12b), vegetation started growing, and with enough time allowed the Vertisols to form. Their immaturity indicates that their development was interrupted very often by renewed flooding. The bentonite (altered volacanic ash) beds which interfinger or are totally incorporated into the mudstone matrix covered the playa lake surface filling depressions and desiccation cracks, and gave rise to the clay gel which later developed cracks as it underwent desiccation during dry periods. This resulted in the formation of the septarian nodules. During the wet seasons the desiccation cracks were filled with surface material and closed while the vegetation was probably destroyed. The different horizons and soil profiles recognized in this unit indicate that at least three major floods occurred. The first flood ended with volcanic ash covering most of the playa lake surface and the formation of septarian nodule. The second major flood deposited the horizons which show extensive desiccation polygons. The third major flood ended with the rapid sedimentation which deposited the braided stream and its associated floodplain.

The upper unit is interpreted as a perennial saline lake. A modern analog of this lake is Lake Magadi which lies in the Eastern Rift Valley, Kenya. Extensive work on Lake Magadi by Eugster (1967; 1969), Hay (1968), and Surdam and Eugster (1976) indicated that where volcanic activity leads to high concentrations of sodium carbonate and to the addition of abundant volcanic debris, complex and unusual precipitation and reactions occurs. Trona is the most abundant precipitate and occurs as thick layers at the lake floor. Chert forms from two sodium-silicate minerals, magadiite and kenyaite (Hay, 1968). Magadiite is chemically precipitated in the lake at a time of higher level and lower salinity. It contains nodules of kenyaite (product of magadiite) some of which contain cores of chert (Eugster, 1967; Hay, 1968) with the chert replacing the kenyaite. The pH is generally high ranging from 9.5 to 11 with the solubility of silica increases as the pH increases. Two types of mechanisms are suggested by Hay (1968): 1) evaporative concentration to the point of precipitation, and 2) lowering the silica solubility by decreasing the pH. Eugster (1969) has supported the second mechanism suggesting that the lowering of pH by disolution would precipitate magadiite. Other mechanisms may also be possible. Jones, Retting and Eugster (1967) indicated that precipitation of trona can raise the pH of a brine above the level determined by equilibrium with atmospheric carbon dioxide, and if additional silica were dissolved at the high pH, subsequent equilibration with atmospheric carbon dioxide would reduce the pH and might precipitate a sodium-silicate mineral or amorphous silica. Hay (1970) also suggested that alkaline, silica-rich water entering a sodium-rich brine of lower pH might yield magadiite or fermentation of organic materials in an anaerobic environment could lower the pH of a brine by increasing the partial pressure of carbon dioxide. The actual process of converting a deposit of a sodium-silicate mineral to chert is poorly understood. Magadiite-chert has been reported to occur in the Green River Formation (Eugster and Surdam, 1971; Smoot, 1983), in lakes at Trinity County, California (McAtee et al., 1968) and in Alkali Lake in Oregon (Rooney et al., 1969). Bricker (1969) using thermodynamic data indicated that magadiite appears to be a direct precipitate from alkaline brines and kenyaite is an

Figure 24. Chemical reactions during the magadiite to kenyaite to chert transformation.

$$22NaSi_7O_{13}(OH)_3.3H_2O_{(s)} + 8H^+ \rightarrow$$

(Magadiite)

$$14NaSi_{11}O_{20.5}(OH)_4.3H_2O_{(s)} + 8Na^+ + 33H_2O$$

(Kenyaite)

$$NaSi_{11}O_{20.5}(OH)_4.3H_2O_{(s)} + H^+ \rightarrow$$

(Kenyaite)

$$11SiO_{2(s)} + Na^+ + 5.5H_2O$$

(Chert)

$$NaSi_7O_{13}(OH)_3.3H_2O_{(s)} + H^+ \rightarrow$$

(Magadiite)

$$7SiO_{2(s)} + Na^+ + 5H_2O$$

(Chert)

intermediate product in the transformation of magadiite to kenyaite to chert by the action of fresher water percolating through the lake beds. The chemical transition of magadiite to kenyaite to chert involves only the removal of sodium ions and partial dehydration. The chemical reactions during the transformation are shown in Figure 24.

The Himes Member Vertisols were deposited on a floodplain associated with the Greybull channels. The dense mottling indicates that intense waterlogging of the soil horizons occurred. Translocation of clay from the absent A horizon was very intense. Ped surfaces are well differentiated and show slickensides which formed due to the shrink-swell processes of the Vertisols. The unclassified paleosols of the Himes Member were developed on a floodplain. They are aerially very extensive which suggests that a very broad floodplain existed during the deposition of the uppermost Cloverly.

SUMMARY

The paleosols of the Morrison Formation are well developed (mature) Spodosols and Aridisols which were formed on broad floodplains. Climatic conditions during the deposition of the Morrison are interpreted as being subhumid to subtropical changing to a semiarid to arid climate with the aridity increasing during the deposition of the uppermost Morrison. This is also evidenced by the increase of calcium carbonate content and decrease in plant roots and vertebrate fossils.

The paleosols of the Cloverly Formation are interpreted as weakly developed (immature) Vertisols. They formed on playa surfaces during drier periods when the lake waters were low or evaporated and soil-forming processes were active. They also formed on floodplains where waterlogging of the soil profile was very intense as is indicated by the dense mottling. Finally, the mudstones of the uppermost Himes Member are poorly developed paleosols. The paleoclimate is interpreted as being semiarid with the aridity decreasing during the deposition of the upper Cloverly.

LITERATURE CITED

Abbott, P. L., J. A. Minch, and G. L. Peterson. 1976. Pre-Eocene paleosol south of Tijuana, Baja California, Mexico. J. Sed. Petr. 46:355-361.

Allen, J. R. L. 1960. Cornstone. Geol. Mag. 97:43.

Allen, J. R. L. 1973. Compressional structures (patterned ground in Devonian pedogenic limestones. Nature (London) Phys. Sci. 243:84-86.

Allen, J. R. L. 1974a. Sedimentology of the Old Red Sandstone (Siluro-Devonian) in the Clee Hill area, Shropshire, England. Sediment. Geol. 12:73-167.

Allen, J. R. L. 1974b. Studies in fluviatile sedimentation implications of pedogenic carbonate units, Lower Old Red Sandstone, Anglo-Welsh outcrop. Geol. J. 9:181-208.

Allen, P. 1959. The Wealden environment Anglo-Paris Basin. Phil. Trans. R. Soc. (London) B242:283-346.

Allen, P. 1976. Wealden of the Weald: a new model. Proc. Geol. Assoc. 86:389-436.

Andreis, R. R. 1972. Paleosuelos de la formation Musters (Eoceno medio), Laguna del Mate, Prov. de Chubut, Rep. Argentina. Revta Asoc. Min. Petr. Sed. Argent. 3:91-98.

Andrews, C. A., W. G. Pierce, and D. H. Eargle. 1947. Geologic map of the Bighorn Basin, Wyoming and Montana, showing terrace deposits and physiographic features. U. S. Geol. Survey. Oil and Gas Inv. Prelim. Map 71.

Aguilera, N. H., and M. C. Jackson. 1953. Iron oxide removal from soils and clays. Soil Sci. Soc. Am. Proc. 17:359-364.

Avery, B. W. 1985. Argillic horizons and their significance in England and Wales. p.69-85. In J. Boardman (ed.). Soils and Quaternary landscape evolution. John Wiley & Sons, New York.

Bates, R. L., and J. A. Jackson (eds.). 1980. Glossary of Geology. 2nd edition. American Geol Inst., Falls Church, Virginia.

Bernoulli, D., and L. W. Wagner. 1971. Suaerial diagenesis and fossil caliche deposits in the Calcare Massico Formation (Lower Jurassic, central Appenines, Italy). Neues Jahrb. Geol. Paleont. Abh. 138:135-149.

Birkeland, P. W. 1984. Soils and geomorphology. Oxford Univ. Press, New York.

Black, C. A. 1968. Soil plant relationships. Wiley, New York.

Blackstone, D. L., and C. W. Sternberg. 1947. Guidebook, 2nd annual field conference in the Bighorn Basin, Wyoming. Wyo. Geol Assoc., Laramie, Wyoming.

Blades, E. L., and M. E. Bickford. 1979. Volcanic ash-flow and volcaniclastic sedimentary rocks at Johnson Shutins, Reynolds county, Missouri. Geol. Soc. Amer. Abstr. with Progr. 8:6.

Blodgett, R. H. 1985. Paleovertisols as indicators of climate. AAPG Bull. 69:239.

Bosellini, A., and D. Rossi. 1974. Triassic carbonate buildups of the dolomites, northern Italy. p. 209-233. In L. F. Laporte (ed.). Reefs in time and space. SEPM Spec. Publ. 18.

Boucot, A. J., J. F. Dewey, D. L. Dineley, R. Fletcher, W. K. Fyson, J. G. Griffin, C. F. Hickox, W. S. McKerrow, A. M. Ziegler. 1974. Geology of the Arisaig area, Antigonish county, Nova Scotia. Geol. Soc. Amer. Spec. Paper 139.

Bown, T. M., and M. J. Kraus. 1981a. Lower Eocene alluvial paleosols (Willwood Formation, NW Wyoming, USA) and their significance for paleoecology, paleoclimatology and basin analysis. Paleogeography, Paleoclimatology, Paleoecology 34:1-30.

Bown, T. M., and M. J. Kraus. 1981b. Vertibrate fossil-bearing paleosol units (Willwood Formation, Lower Eocene, NW Wyoming, USA): Implications for taphonomy, biostratigraphy, and assemblage analysis. Paleogeography, Paleoclimatology, Paleoecology 34:31-56.

Brammer, H. 1968. Decalcification of soils developed in calcareous Gangetic alluvium in East Pakistan. Pak. J. Soil Sci. 4:8-20.

Brewer, R. 1960. Cutans: their definition, recognition and interpretation. J. Soil Sci. 11:280-292.

Brewer, R. 1964. Fabric and mineral analysis of soils. John Wiley & Sons, New York.

Brewer, R. 1972. The basis of interpretation of soil micromorphological data. Geoderma 8:81-94.

Brewer, R., and J. R. Sleeman. 1963. Pedotubules: their definition, classification and interpretation. J. Soil Sci. 14:156-166.

Brewer, R., and J. R. Sleeman. 1964. Glaebules: their definition, classification and interpretation. J. Soil Sci. 15:66-78.

Bricker, O. P. 1969. Stability constants and Gibbs free energy of formation of magadiite and kenyaite. The American Mineralogists 54:1026-1033.

Brown, B. 1935. Sinclair Dinosaur Expedition, 1934. Natural History 36:3-15.

Brown, B. 1937. Excavating the great Jurassic dinosaur quarry of northern Wyoming, 140,000,000 years ago. Roengten Econimist 5:3-6.

Buol, S. W., F. D. Hole, and P. J. McCracken. 1973. Soil genesis and classification. Iowa State Univ. Press, Ames,Iowa.

Buurman, P. 1975. Possibilities in Paleopedology. Sedimentology 22:289-298.

Buurman, P. 1980. Paleosols in the Reading beds (Paleocene) of Alum Bay, Isle of Wight, U.K. Sedimentology 27:593-606.

Chalyshev, V. I. 1969. A discovery of fossil soils in the Permo-Triassic. Dokl. Acad. Sci. USSR, Earth Sci. 128:53-56.

Collins, W. H. 1925. North shore of Lake Huron. Geol. Surv. Canada, Memoir 143.

Collinson, J. D., and D. B. Thompson. 1982. Sedimentary structures. Allen and Unwin, London.

Commission for Paleopedology of INQUA. 1971. Criteria for the recognition and classification of paleosols. p. 153-158. In D. H. Yaloon (ed.). Paleopedology-origin, nature and dating of paleosols. Int. Soc. Soil Sc. and Israel Univ. Press, Jerusalem.

Coney, P. J. 1972. Cordilleran tectonics and North America plate motion. Am. J. Sci. 272:603-628.

Danilov, I. S. 1968. The nature and geochemical conditions of formation of variegated red Triassic and Permian rocks of the western Donetz Basin. Lithol. Miner. Resources 3:612-618.

Darton, N. H. 1904. Comparison of the stratigraphy of the Black Hills, Bighorn Mountains, and Rocky Mountain Front Range. Geol. Soc. Amer. Bull. 15:379-448.

Darton, N. H. 1906. Geology of the Bighorn Mountains. US Geol. Survey Prof. Paper 51.

Davidson, C. F. 1965. Geochemical aspects of atmospheric evolution. Proc. US Nat. Acad. Sci. 53:1194-1204.

Donaldson, J. A. 1969. Descriptive notes (with particular reference to the late Proterozoic Dubawnt Group) to accompany a geological map of central Thelon Plain, Districts of Keewatin and Mackenzie. Pap. Geol. Surv. Canada. 68-49.

Dregne, H. E. 1976. Soil of arid regions. Elsevier, Amsterdam.

Drosdoff, M., and C. C. Nikiforoff. 1940. Iron-manganese concretions in Dayton soils. Soil Sci. 49:333-345.

Duchaufour, P. 1977. Pedology- pedogenesis and classification. Allen and Unwin, London.

Dunham, R. J. 1969. Vadose pisolites in the Capitan reef (Permian), New Mexico and Texas. p. 182-191. In G. M. Friedman (ed.). Depositional environments in carbonate rocks. SEPM Spec. Publ. 14.

Dunoyer De Segonzac, G. 1970. The transformation of clay minerals during diagenesis and low-grade metamorphism: a review. Sedimentology 15:281-346.

Eskola, P. 1963. The Precambrian of Finland. p. 145-203. In K. Rankama (ed.). The Precambrian. Interscience, New York.

Estaban, M. 1972. The presence of fossil caliche at the base of the Eocene of the Catalonia Mountains Tarragosa, Barcelona, Spain. Acta Geol. Hisp. 7:164-168.

Estaban, H., and L. C. Pray. 1977. Origin of the pisolitic facies of the shelf crest. p. 479-484. In Upper Guadalupian Facies, Permian Reef Compex, Guadalupe Mountains, New Mexico and Texas. Publ. Permian Basin Sect. SEPM Midland 16.

Eugster, H. P. 1967. Hydrous sodium silicates from the Magadi, Kenya: precursors of bedded chert. Science 157:1177-1180.

Eugster, H. P. 1969. Inorganic bedded cherts from the Magadi area, Kenya. Contr. Mineralogy and Petrology 22:1-31.

Eugster, H. P., and R. C. Surdam. 1971. Bedded cherts in the Green River Formation. Geol. Soc. Amer. Abstr. 3:559-560.

Eugster, H. P., and L. A. Hardie. 1978. Saline lakes. In A. Lerman (ed.). Lakes: chemistry, geology, physics. Springer-Verlag, New York.

Faugeres, L., and P. Robert. 1969. Precisions nouvelles sur les alterations contenues dans le Serie du Chainon de Viglia (Kozani, Macedonie, Grece). C. R. Soc. Geol. Fr. 3:97-98.

Feofilova, A. P. 1977. Paleopedology and its significance for the reconstruction of ancient landscapes. Lithol. Miner. Resources 12:650-655.

Folk, R. L. 1968. Petrology of sedimentary rocks. Hemphills, Austin.

Frarey, M. J., and S. M. Roscoe. 1970. The Huronian Supergroup north of Lake Huron. p. 143-157. In A. J. Baer (ed.) Symposium on basins and geosynclines of the Canadian Shield. Geol. Surv. Canada Paper 40.

Fraser, J. A., J. A. Donaldson, W. F. Fahrig, and L. P. Tremblay. 1970. Helikian basins and geosynclines of the northwesern Canadian Shield. p. 213-238. In A. J. Baer (ed.) Symposium on basins and geosynclines of the Canadian Shield. Geol. Surv. Canada Paper 40.

Freytet, P. 1971. Paleosols residuels et paleosols alluviaux hydromorphes associes aux depots fluviatiles dans Cretace superieur et l'Eocene basal du Languedoc. Rev. Geogr. Phys. Geol. Dynam. 13:245-360.

Freytet, P. 1973. Petrography and paleoenvironment of continental carbonate deposits, with particular reference to the upper Cretaceous and lower Eocene of Languedoc (southern France). Sediment. Geol. 10:25-60.

Friend, P. F. 1966. Clay fractions and color of some Devonian red beds in the Catskill Mountains, USA. Geol. Soc. London Quater. J. 122:273-292.

Friend, P. F., W. B. Harland, and A. Gilbert-Smith. 1970. Reddening and fissuring associated with the Caledonian unconformity in northwest Arran. Proc. Geol. Assoc. 81:75-85.

Geological Society of America News and Information. 1986. 18:203.

Hardie, L. A., J. P. Smoot and H. P. Eugster. 1978. Saline lakes and their deposits: a sedimentological approach. In A. Matter and M. E. Tacker (eds.) Modern and ancient lake sediments, Int. Assoc. Sediment. Sp. Publ. 2.

Hares, C. J. 1917. Gastroliths in the Cloverly Formation. Proc. Geol. Soc. Washington, in J. Wash. Acad. Sci. 7:429.

Harvey, C. H. 1960. Stratigraphy, sedimentation and environment of the White River Group of northern Sioux county, Nebraska. Ph.D. Thesis. Univ. of Nebraska, Lincoln, Nebraska.

Hay, R. L. 1968. Chert and its sodium-silicate precursors in sodium-carbonate lakes of East Africa. Contr. Mineralogy and Petrology 17:255-274.

Hay, R. L. 1970. Silicate reactions in three lithofacies of a semiarid basin, Olduvai Gorgr, Tanzania. Miner. Soc. Amer. Spec. Paper 3:237-255.

Hewett, D. F. 1914. The Shoshone river section, Wyoming. US Geol. Surv. Bull. 541:89-113.

Hewett, D. F., and C. T. Lupton. 1917. Anticlines in the southern part of the Bighorn Basin, Wyoming. US Geol. Surv. Bull. 656:1-192.

Hintze, F. F., Jr. 1915. The Bighorn and Greybull oil and gas fields. Geol. Surv. Wyo. Bull. 10:1-60.

Hubert, J. F. 1960. Petrology of the Fountain and Lyons Formations, Front Range, Colorado. Colo. School Mines Q. 55:1-242.

Hubert, J. F. 1977a. Paleosol caliche in the New Haven Arkose, Newark Group, Connecticut. Paleogeography, Paleoclimatology, Paleoecology 24:151-168.

Hubert, J. F. 1977b. Paleosol caliche in the New Haven Arkose, Connecticut: record of semiaridity in late Triassic-early Jurassic time. Geology 5:302-304.

Huddle, J. W., and S. H. Patterson. 1961. Origin of Pennsylvanian underclay and related seat rocks. Geol. Soc. Amer. Bull. 72:1643-1660.

Hutton, J. T., C. R. Twidale, and A. R. Milnes. 1978. Characteristics and origin of some Australian silcretes. p. 19-39. In T. Langford-Smith (ed.). Silcrete in Australia. Univ. of New England, Armidale, Australia.

Jackson, M. L. 1958. Soil chemical analysis. Prentice-Hall, Englewood, California.

Jenkins, D. A. 1985. Chemical and mineralogical composition in the identification of paleosols. p. 23-43. In J. Boardman (ed.) Soils and Quaternary landscape evolution. John Wiley & Sons, New York.

Johnson, H. S., and S. D. Heron. 1965. Slump features in the McBean Formation and younger beds, Riley Cut, Calhoun county, South Carolina. Geol. Notes S. Carolina Devel. Board Div. Geol. 9:37-44.

Jones, B. F., S. L. Retting, and H. P. Eugster. 1967. Silicate in alcaline brines. Science 158:1310-1314.

Kalliokosi, J. 1975. Chemistry and mineralogy of Precambrian paleosols in northern Michigan. Geol. Soc. Amer. Bull. 86:371-376.

Keller, W. D., J. F. Wescott, and A. O. Bledsoe. 1954. The origin of Missouri fire clays. Clays and Clay Minerals 2:7-46.

Kidston, R., and W. H. Lang. 1921. On Old Red Sandstone plants showing structure from the Rhynie chert bed, Aberdeenshire. Part V. The thallophyta occurring in the peat bed; the succession of the plants through a vertical section of the bed and the conditions of accumulation and preservation of the deposit. Trans. R. Soc. Edinb. 52:855-892.

Kvale, E. P. 1986. Paleoenvironments and tectonic significance of the Upper Jurrasic Morrison and Lower Cretaceous Cloverly formations, Bighorn Basin, Wyoming. Ph.D Thesis. Iowa State University, Ames, Iowa.

Kvale, E. P., and C. F. Vondra. 1985. Depositional environments of the Lower Cretaceous Cloverly Formation, Bighorn Basin, Wyoming. SEPM Abtr., Annual Midyear Meeting. 2:53.

Lee, W. T. 1927. Correlation of the geologic formations between east-central Colorado, central Wyoming, and southern Montana. US Geol. Surv. Prof. Paper 149:1-80.

Leeder, M. R. 1975. Pedogenic carbonates and flood sediment accretion rates: a quantitative model for alluvial arid-zone lithofacies. Geol. Mag. 112:257-270.

Leeder, M. R. 1976. Paleogeographic significance of pedogenic carbonates in the topmost Upper Old Red Sandstone of the Scottish Border Basin. Geol. J. 11-21-28.

Lupton, C. T. 1916. Oil and gas near Basin, Bighorn county, Wyoming. US Geol. Surv. Bull. 621-I:157-190.

Machette, M. N. 1985. Calcic soils of the southwestern U. S. p. 1-21. In D. L. Weide (ed.) Soils and Quaternary geology of the southwestern US. Geol. Soc. Amer. Spec. Paper 203, Boulder, Colorado.

Maiklem, W. R. 1971. Evaporative drawdown-a mechanism for waterlevel lowering and diagenesis in the Elk Point Basin. Canad. Petr. Geol. Bull. 19:487-503.

Mantzios, C. and C. F. Vondra. 1986a. The paleosols of the upper Jurassic Morrison Formation, Bighorn Basin, Wyoming. SEPM Abstr., Midyear Annual Meeting, 3:71.

Mantzios, C. and C. F. Vondra. 1986b. Paleosols as tools for paleoclimatic interpretations. Geol. Soc. Amer. Abstr. with Progr. 18:681.

McAtee, J. L., R. House, and H. P. Eugster. 1968. Magadiite from Trinity County, California. Am. Miner. 53:2061-2069.

McBride, E. F. 1974. Significance of color in red, green, purple, olive, brown, and gray beds of the Difunta group, northeastern Mexico. J. Sed. Petr. 44:760-773.

McFadden, L. D., and J. C. Tinsley. 1985. Rate and depth of pedogenic carbonate accumulation in soils: Formation and testing of a compartment model. p. 23-41. In D. L. Weide (ed.) Soils and Quaternary geology of the south-western US. Geol. Soc. Amer. Spec. Paper 203.

McKee, E. D. 1969. Paleozoic rocks of the Grand Canyon. Guidebook Field Conf., Four Corners Geol. Soc. 5:78-90.

McPherson, J. G. 1979. Calcrete (calichi) paleosols in fluvial redbeds of the Aztec siltstone (Upper Devonian), southern Victoria Land, Antarctica. Sedim. Geol. 22:267-285.

Millot, G. 1970. Geology of clays. Springer-Verlag, New York.

Millot, G. 1977. Clay. Sci. Amer. 240:108-120.

Moberly, R., Jr. 1956. Mesozoic Morisson, Cloverly and Crooked Creek formations, Bighorn Basin, Wyoming and Montana. Ph.D. Thesis. Princeton University, Princeton, New Jersey.

Moberly, R., Jr. 1960. Morrison, Cloverly, and Sykes Mountain Formations, northern Bighorn Basin, Wyoming and Montana. Geol. Soc. Amer. Bull. 71:1137-1176.

Molenaar, N. 1984. Paleopedogenic features and their paleoclimatological significance for the Nevremont Formation (Lower Givetian), the northern Ardennes, Belgium. Paleogeography, Paleoclimatology, Paleoecology. 46:325-344.

Morad, S. 1983. Diagenesis and geochemistry of the Visingo group (Upper Proterozoic), southern Sweden: A clue to theorigin of color differantiation. J. Sed. Petr. 53:51-65.

Mountain, E. D. 1951. The origin of silcrete. South Afr. J. Sci. 48:201-204.

Munsell Color Chart. 1954. Munsell soil color charts. Munsell Color Chart, Baltimore, Maryland.

Neal, J. T., and W. S. Motts. 1967. Recent geomorphic changes in playas of western United States. J. Geol 75:511-525.

Ortlam, D. 1971. Paleosols and their significance in stratigraphy and applied geology in the Permian and Triassic of southern Germany. p. 321-328. In D. H. Yaalon (ed.). Paleopedology-origin, nature and dating of paleosols. Int. Soc. Soil Sci. and Israel Univ. Press, Jerusalem, Israel.

Ostrom, J. H. 1970. Cloverly Formation, stratigraphy and paleontology. Yale Univ. Peabody Museum of Natural Hist. Bull. 35:234.

Patel, I. M. 1977. Late Precambrian fssil soil horizon in southern New Brunswick and its stratigraphic significance. Geol. Soc. Amer. Prog. Abstr. 9:308..

Peterson G. L., and P. L. Abbott. 1979. Mid-Eocene climatic change in southwestern California and northwestern Baja, California. Paleogeography, Paleoclimatology, Paleoecology 26:73-87.

Pierce, W. G. 1948. Geologic and structure contour map of the Basin-Greybull area, Bighorn County, Wyoming. US Geol. Surv. Prelim. Map 77.

Pierce, W. G., and D. A. Andrews. 1940. Geology and oil and gas resources of the region south of Cody, Park County, Wyoming. US Geol. Surv. Bull. 921-B:99-180.

Rankama, K. 1955. Geologic evidence of chemical composition of the Precambrian atmosphere. p. 651-664. In A. Poldervaart (ed.). Crust of the earth. Geol. Soc. Amer. Spec. Paper 62.

Reeves, C. C., Jr. 1970. Origin, classification, and geologic history of calichi on the southern High Plains, Texas and eastern New Mexico. J. Geol. 78:353-362.

Retallack, G. J. 1977a. Triassic paleosols in the upper Narrabeen group of New South Wales. Part I: Features of the paleosols. J. Geol. Soc. Austr. 23:383-399.

Retallack, G. J. 1977b. Triassic paleosols in the upper Narrabeen group of New South Wales. Part II: Classification and reconstruction. J. Geol. Soc. Austr. 24:19-36.

Retallack, G. J. 1979. Early Tertiary fossil soils in the Badlands National Monument, South Dakota. Geol. Soc. Amer. Abstr. with Progr. 11:502.

Retallack, G. J. 1981. Fossil soils: indicators of ancient terrestrial environments. p. 55-102. In K. J. Niklas (ed.). Paleobotany, paleoecology and evolution. Praeger Publ., New York.

Retallack, G. J. 1983a. A paleopedological approach to the interpretation of terrestrial sedimentary rocks: the mid-Tertiary fossil soils of Badlands National Park, South Dakota. Geol. Soc. Amer. Bull. 94:823-840.

Retallack, G. J. 1983b. Late Eocene and Oligocene paleosols from Badlands National Park, South Dakota. Geol. Soc. Amer. Spec. Paper 193.

Retallack, G. J. 1984. Completeness of the rock and fossil record: some estimates using fossil soils. Paleobiology 10:59-78.

Retallack, G. J. 1986. Fossil soils as grounds for interpreting long-term control on ancient rivers. J. Sed. Petr. 56:1-18.

Retallack, G. J., D. Grandstaff, and M. Kimberley. 1984. The promise and problems of Precambrian paleosols. Episodes 7:8-12.

Rettke, R. 1981. Probable burial diagenetic and provenance effects on Dakota group clay mineralogy, Denver Basin. J. Sed. Petr. 51:541-551.

Reynolds, R. C. Jr., and J. Hower. 1970. The nature of interlayering in mixed-layer illite-montmorillonite. Clays and Clay Minerals 18:25-36.

Roeschmann, G. 1971. Problems concerning investigations of paleosols in older sedimentary rocks, demonstrated by the example of Wurzelboden of the Carboniferous system. p. 311-320. In D. H. Yaalon (ed.). Paleopedology-origin, nature and dating of paleosols. Int. Soc. Soil Sci. and Israel Univ. Press, Jerusalem, Israel.

Rogers, C. P. 1948. Geology of the Worland-Hyatville area, Bighorn and Washakie counties, Wyoming. US Geol. Surv., Oil and Gas Inv. Prelim. Map 84.

Rooney, T. P., B. F. Jones, and J. T. Neal. 1969. Magadite from Alkali lake, Oregon. Am. Miner. 54:1034-1043.

Roscoe, S. M. 1968. Huronian rocks and uraniferous conglomerates in the Canadian Shield. Geol. Surv. Canada Paper 40.

Schultz, C. B., and T. M. Stout. 1955. Classification of Oligocene sediments in Nebraska. Univ. Nebr. State Mus. Bull. 4:17-52.

Schultz, C. B., L. G. Tanner, and C. H. Harvey. 1955. Paleosols of the Oligocene of Nebraska. Univ. Nebr. State Mus. Bull. 4:1-16.

Senior, B. R. 1978. Silcrete and chemically weathered sediments in southwest Queensland. p. 41-50. In T. Langford-Smith (ed.). Silcrete in Australia. Univ. New England, Armidale, Australia.

Sharp, R. P. 1940. The Ep-Archean and Ep-Algonkian erosion surfaces, Grand Canyon, Arizona. Geol. Soc. Amer. Bull. 51:1235-1269.

Singer, A. 1980. The paleoclimatic interpretation of clay minerals in soils and weathering profiles. Earth Sci. Rev. 15:303-326.

Singer, A. 1984. The paleoclimatic interpretation of clay minerals in sediments-a review. Earth Sci. Rev. 21:251-293.

Smale, D. 1973. Silcretes and associated silica diagenesis in southern Africa and Australia. J. Sed. Petr. 43:1077-1089.

Smart, P., and N. K. Tovey. 1981. Electron microscopy of soils and sediments: examples. Clarendon Press, Oxford.

Smart, P., and N. K. Tovey. 1982. Electron microscopy of soils and sediments: techniques. Clarendon Press, Oxford.

Smoot, J. P. 1983. Depositional environments in an arid closed basin; the Wilkins Peak Member of the Green River Formation (Eocene), Wyoming. Sedimentology 30:801-827.

Soil Science Society of America. 1965. Glossary of soil science terms. Soil Sci. Soc. Amer. Proc. 29:330-351.

Soil Survey Staff. 1975. Soil taxonomy. US Dept. Agric. Handbook 436.

Spalletti, L. A., and M. M. Mazzoni. 1978. Sedimentologia del Grupo Sarmiento en un perfil ubicato al sudeste del Lago Colhue Huapi, Provincia de Chubut. Obra Centen. Mus. La Plata 4:261-283.

Steel, R. J. 1974. Cornstone (fossil calichi)- its origin, stratigraphy, and sedimentological importance in the New Red Sandstone, western Scotland. J. Geol. 82:351-369.

Summerfield, M. A. 1983. Petrography and diagenesis of silcrete from the Kalahari Basin and Cape coastal zone, southern Africa. J. Sed. Petr. 53:895-909.

Summerfield, M. A. 1984. Isovolumetric weathering and silcrete formation, southern Cape province, South Africa. Earth Sur. Proc. and Land 9:135-141.

Surdam, R. C., and H. P. Eugster. 1976. Mineral reactions in the sedimentary deposits of the Lake Magadi region, Kenya. Geol. Soc. Amer. Bull. 87:1739-1752.

Surdam, R. C., H. P. Eugster, and R. H. Mariner. 1972. Magadi-type chert in Jurassic and Eocene to Pleistocene rocks, Wyoming. Geol. Soc. Amer. Bull. 83:2261-2266.

Teruggi, M. E., and R. R. Andreis. 1971. Micromorphological recognition of paleosolic features in sediments and sedimentary rocks. p. 161-172. In D. H. Yaalon (ed.). Paleopedology- origin, nature and dating of paleosols. Int. Soc. Soil Sci. and Israel Univ. Press, Jerusalem, Israel.

Thompson, A. M. 1970. Geochemistry of color genesis in red-bed sequences, Juniata and Bald Eagle formations, Pennsylvania. Jour. Sed. Petr. 40:599-615.

Tunbridge, I. A. 1984. Facies model for a sandy ephemeral stream and clay playa complex; the Middle Devonian Trentishoe Formation of North Devon, UK. Sedimentology 31:697-715.

Turner, P. 1980. Continental red beds. Developments in Sedimentology 29. Elsevier, Amsterdam.

Turner-Peterson, C. E. 1985. Lacustrine-Humate model for Primary Uranium ore deposits, Grants Uranium Region, New Mexico. AAPG Bull. 69:1999-2020.

Van Houten, F. B. 1972. Iron and clay in tropical savanaah alluvium, a contribution to the origin of red beds. Geol. Soc. Amer. Bull. 83:2761-2772.

Walkden, G. M. 1974. Paleokarstic surfaces in upper Visean (Carboniferous) limestones of the Derbyshire block, England. J. Sed. Petr. 44:1232-1247.

Walker, T. R. 1967. Formation of red beds in ancient and modern deserts. Geol. Soc. Amer. Bull. 78:353-368.

Walls, R. A., W. B. Harris, and W. E. Nunan. 1975. Calcareous crust (caliche) profiles and early subaerial exposure of Carboniferous carbonates, northeastern Kentucky. Sedimentology 22:417-440.

Wardlaw, N. C., and G. E. Reinson. 1971. Carboate and evaporite deposition and diagenesis, Middle Devonian Winnepegosis and Prairie evaporite Formations of south central Saskatchewan. AAPG Bull. 55:1759-1789.

Washburne, C. W. 1909. Coal fields of the northeast side of the Bighorn basin, Wyoming and of Bridger Montana. US Geol. Surv. Bull. 341-B:165-199.

Watts, N. L. 1976. Paleopedogenic palygorskite from the basal Permo-Triassic of the northwest Scotland. Am. Miner. 61:299-302.

Watts, N. L. 1978. Displacive calcite: evidence from recent and ancient calcretes. Geology 6:699-703.

Weaver, C. E. 1967. The significance of clay minerals in sediments. p. 37-75. In B. Nagy and U. Colombo (eds.) Fundamental aspects of petroleum geochemistry. Elsevier, Amsterdam.

White, E. M. 1972. Soil-desiccation features in South Dakota depressions. J. Geol. 80:106-111.

Williams, G. E. 1968. Torridonian weathering and its bearing on Torrindonian paleoclimate and source. Scottish J. Geol. 4:164-184.

Wilmarth, M. G. 1925. Tentative correlation of geologic formations in Wyoming. US Geol. Surv Comm. Geol. Names, Chart.

Wilson, M. D., and E. D. Pittman. 1977. Authigenic clays in sandstones: recognition and influence on reservoir properties and paleoenvironmental analysis. J. Sed. Petr. 47:3-31.

Woodrow, D. L., F. W. Fletcher, and W. F. Ahrnsbrak. 1973. Paleogeography and paleoclimate at the deposition sites of the Devonian Catskill and Old Red facies. Geol. Soc. Amer. Bull. 84:3051-3064.

Wopfner, H. 1983. Environment of silcrete formation: a comparison of examples from Australia and the Cologne Embeyment, West Germany. p. 151-158. In R. C. L. Wilson (ed.). Residual deposits. Geol. Soc. London Sp. Publ. 11. Blackwell Sci. Publ., London.

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APPENDIX A

Glossary of terms

Soil horizon	A layer of a soil that is distinguishable from adjacent layers by characteristic physical properties such as structure, color, or texture, or by chemical composition, including content of organic matter or degree of aridity or alkalinity (Bates and Jackson, 1980).
A horizon	Mineral horizons consisting of: (i) horizons of organic-matter accumulation formed at or adjacent to the surface; (ii) horizons that have lost clay, iron, or aluminum with resultant concentration of quartz or other resistant minerals of sand or silt size; or horizons dominated by (i) or (ii) above but transitional to an underlying B or C horizon (Soil Science Society of America, 1965).
Ochric epipedon	A surface horizon of mineral soil that is too light in color, too low in organic carbon, or too thin to be either an umbric or mollic epipedon, or that is both hard and massive when dry (Soil Science Society of America, 1965).
B horizon	Subsurface horizons which show little or no evidence of the original sediment or rock structure. Several kinds are recognized, some based on the kinds of materials illuviated into them, others on residual concentrations of material (Birkeland, 1984).
Spodic horizon	A diagnostic subsurface soil horizon that is characterized by the illuvial accumulation of black or reddish amorphous materials that have a high cation-exchange capacity and consist of aluminum and organic carbon, sometimes with iron. The spodic horizon is moist or wet and has a loamy or sandy texture, a high pH-dependent exchange capacity, and few bases. It is designated as "Bs" (Soil Survey Staff, 1975).

Argillic horizon	A diagnostic subsurface soil horizon characterized by an accumulation of clay. Its thickness and clay content depend on the thickness and clay content of the overlying eluvial horizon. It usually has clay coatings on the surface of peds or pores, or as bridges between sand grains. It is designated as "Bt" (Soil Survey Staff, 1975).
Calcic horizon	A diagnostic subsurface soil horizon, at least 15cm thick, characterized by enrichment in secondary carbonate. It is designated as "Bk" (Soil Survey Staff, 1975).
Spodosols	A soil order characterized by the presence of a spodic horizon. Usually spodosols have an O horizon and an albic horizon above the spodic horizon, and a few have a fragipan or an argillic horizon below it. They generally form from sandy parent materials. In cool, humid climates they usually form under coniferous forests. In hot, humid intertropical areas they form in quartz-rich sands that have fluctuating ground water near the surface (Soil Survey Staff, 1975).
Aridisols	A soil order characterized by an ochric epipedon and other pedogenic horizons, but none of them oxic or sodic. It develops under an aridic moisture regime and has a low percentage of organic carbon (Soil Survey Staff, 1975).
Vertisols	A soil order that contains at least 30% clay, usually smectite. It is found in regions where soil moisture changes rapidly from wet to dry, and is characterized by pronounced changes in volume with changes in moisture, and by deep, wide cracks and high bulk density between the cracks when dry. Gilgai microrelief and slickensides are further evidence of the churning and self-mulching of these shrink-swell soils (Soil Survey Staff, 1975).

Ped	A unit of soil structure such as an aggregate, crumb, prism, block, or granule, formed by natural processes (Soil Science Society of America, 1965).
Cutan	A modification of the texture, structure, or fabric at natural surfaces in soil materials due to concentration of particular soil constituents or in sity modification of the plasms; cutans can be composed of any of the components substances of the soil material (Brewer, 1964).
Glaebule	A three dimensional unit within the s-matrix of the soil material, and usually approximately prolate to equant in shape; its morphology (especially size, shape, and/or internal fabric) is incompatible with its present occurrence being within a single void in the present soil material. It is recognized as a unit either because of a greater concentration of some constituent and/or a difference in fabric compared with the enclosing soil material, or because it has a distinct boundary with the enclosing soil material (Brewer, 1964).
Pedotubules	A pedological feature consisting of soil material (skeleton grains or skeleton grains plus plasma, as distinct from concentrations of the plasma) and having a tabular external form, either single tubes or branching systems of tubes; its external boundaries are relatively sharp. Tabular form, in this context, means that the feature as a unit, or its impression in the enclosing soil material, has a relatively uniform cross-sectional size and shape, most commonly circular or elliptical; that is, the impression of the pedotubule conforms to the definition of channels (Brewer, 1964).

Paleosols	Any broadly tabular rock bodies that show evidence of having being exposed at the ground surface after deposition for long enough periods to have been modified by physical and chemical weathering, including the actions of organisms (Bown and Kraus, 1981a).
Mottling	Is very common in paleosols and is considered analogous to modern soil "gleying" in which the reduction and mobilization of iron, manganese and other elements takes place at the air/groundwater interface as a result of a fluctuating water-table (McPherson, 1979).

APPENDIX B

Description of the soil horizons

The soil horizons are referenced here according to the sequence shown in Tables 2 and 3. The soil horizons 1 through 39 are from section measured at Coyote Basin (T54N, R91W, sec. 18, SE1/4, SW1/4, and T54N, R91W, sec. 19, NW1/4, SE1/4). The soil horizons 40 through 42 are from sections measured at South Nose of Sheep Mountain Anticline (T53N, R93W, sec. 35, SW1/4, NE1/4) and 43 through 46 from sections measured at Rose Dome (T54N, R93W, sec. 35, SW1/4, NE1/4).

1. Sandy siltstone; dark reddish gray (5YR4/2) with mottles of grayish olive (10Y4/2) color; massive; calcareous; slickensides; plant roots; burrows.

2. Sandy siltstone; weak red (10R4/3) with greenish gray (5GY5/1) mottles; massive; nodules at lower part of the horizon; calcareous; plant roots.

3. Siltstone to sandy siltstone; greenish gray (5GY6/1); massive; calcareous; plant roots.

4. Fine sandstone to sandy siltstone; light gray (7.5YR7/0); blocky; calcareous; organic carbon rich(?).

5. Sandy siltstone; dark grayish brown (5YR4/2) with pale olive (10Y6/2) mottles; massive; calcareous; plant roots; burrows; abundant small nodules, 1-2mm; slickensides.

6. Sandy siltstone; wqeak red (10R4/2) with grayish olive (10Y4/2) mottles; massive; calcareous; plant roots.

7. Sandy siltstone; greenish gray (5GY6/1); blocky; calcareous; plant roots.

8. Fine sandstone to sandy siltstone; light gray (7.5YR7/0); blocky; calcareous; organic carbon rich (?).

9. Sandstone; lateral accretion surfaces; ripples; quartz arenite; calcareous.

10. Mudstone; dusky red (10R3/4) with pale olive (10Y6/2) mottles; massive; ped surfaces are distinguishable; slickensides; calcareouys; plant roots are very abundant; burrows filled with carbonate; glaebules and pedotubules.

11. Mudstone; red (10R5/6) with pale olive (10Y6/2) mottles; massive; ped surfaces; slickensides; calcareous; plant roots and burrows; glaebules and pedotubules.

12. Overbank sandstone; quartz arenite; ripples; calcareous.

13. Mudstone; dark reddish brown (5YR3/2) with grayish green (10GY5/2) mottles; massive; calcareous; slickensides; plant roots and burrows; glaebules.

14. Mudstone; weak red (10R4/2) with pale olive (10Y6/2) mottles; massive; calcareous; slickensides; plant roots; glaebules.

15. Mudstone; grayish olive green (5GY3/2); massive; calcareous; some plant roots.

16. Mudstone; weak red (10R4/3) with grayish olive (10Y4/2) mottles; massive; slickensides; calcareous, plant roots; glaebules.

17. Mudstone; dark reddish brown (5YR3/2) with grayish green (10GY5/2) mottles; massive; slickensides; calcareous; plant roots; glaebules; strongly bioturbated.

18. Mudstone; grayish olive green (5GY3/2); massive; calcareous; some plant roots.

19. Mudstone; dark reddish brown (5YR3/2) with grayish green (10GY5/2) mottles; massive; peds; slickensides; calcareous; plant roots; glaebules.

20. Mudstone; pale olive (10Y6/2); massive; calcareous; some plant roots.

21. Mudstone; brownish gray (5YR4/1) with pale olive (10Y6/2) mottles; massive; noncalcareous; plant roots.

22. Mudstone; brownish gray (5YR4/1) with grayish olive (10Y4/2) mottles; massive; noncalcareous; plant roots.

23. Mudstone; Medium dark gray (7.5YR4/0); blocky to massive; slickensides are abundant; noncalcareous; plant roots.

24. Mudstone; light olive gray (5Y6/1) to medium gray (7.5YR5/0); massive; plant roots are abundant; noncalcareous; slickensides very common; septarian nodules.

25. Mudstone; olive gray (5Y5/2); massive; noncalcareous; slickensides are very abundant; desiccation cracks; chalcedony nodules.

26. Mudstone; olive gray (5Y4/1); massive; abundant slickensides; noncalcareous; plant roots.

27. Mudstone; dark gray (7.5YR3/0); massive; noncalcareous; slickensides; plant roots; burrows.

28. Mudstone; olive gray (5Y4/1); massive; noncalcareous; plant roots and burrows; slickensides.

29. Mudstone; weak red (10R4/2); massive; slightly calcareous; plant roots; slickensides; nodules.

30. Mudstone; dusky red (5R3/4) with pale olive (5Y6/4) mootles; massive; slightly calcareous; nodules; plant roots and burrows.

31. Mudstone; weak red (10R4/2) with pale olive (5Y6/4) mottles; massive; plant roots more abundant; burrows; nodules; slickensides; slightly calcareous.

32. Mudstone; dusky red (5R3/4) with pale olive (5Y6/4) mottles; massive; nodules more abundant; noncalcareous; plant roots; slickensides.

33. Mudstone; pale olive (10Y6/2); noncalcareous; massive to blocky; bone fragments very abundant.

34. Mudstone; grayish red purple (5RP4/2); massive to blocky; noncalcareous; slickensides; chert-carbonate horizons.

35. Mudstone; medium dark gray (7.5YR4/0); massive; noncalcareous; plant roots not very abundant; slickensides very common.

36. Mudstone; light olive gray (5Y6/1); massive; noncalcareous; plant roots; burrows; slickensides; chert-carbonate nodules.

37. Mudstone; olive gray (5Y4/1); massive; plant roots not very abundant; noncalcareous; slickensides very common; chert-carbonate nodules; tree trunks (?).

38. Devitrified tuff.

39. Mudstone; medium gray (7.5YR5/0); massive to blocky; noncalcareous; slickensides; plant roots and burrows.

40. Mudstone; dusky red (10R3/4) and light greenish gray (5GY8/1); massive; abundant slickensides; densely mottled; plant roots; noncalcareous.

41. Mudstone; dusky red (10R3/4) and greenish gray (5GY6/1); massive; abundant slickensides; densely mottled; plant rrots; noncalcareous.

42. Mudstone; mopderate red (5R4/6) and medium gray (7.5YR5/0); massive; abundant slickensides; densely mottled; plant roots; noncalcareous.

43. Mudstone; olive gray (5Y5/1) and dusky red (10R3/4); no pedogenic features.

44. Mudstone; light olive gray (5Y6/1); no pedogenic features; crocodile and turtle bones.

45. Mudstone; blackish red (5R2/2); no pedogenic features.

46. Mudstone; olive gray (5Y5/2); no pedogenic features; transitional with the overlying Sykes Mountain Formation.

APPENDIX C

Chemical Methods

1). Carbonate content: Samples were analyzed for carbonate content for two reasons. First to determine the carbonate content, and second to remove it for further use of the sample in other chemical analyses. The samples were first weight and then placed in a beaker. Hydrochloric acid (10%) was added slowly so to cover almost the total sample. After stirring it was left for a few hours so that the hydrochloric acid could react with the carbonate. More acid was added if it was needed. After no reaction was visible, the beaker was filled with water and the sample was washed by repeater sedimentation and decantation until the supernantant solution was nearly free from acid (Brewer, 1964).

2). Organic matter content: Sample treated with the previous method was used to determine the organic matter content. The method is after Brewer (1964). 100g of sample was mixed with 300ml of 6% hydrogen peroxide and allowed the suspension to stay overnight. 300ml of water was then added and the contents were brought to boil for 10 minutes. After cooling, the sample was weight and the difference in weight was the value of the organic carbon content.

3). Iron content: The sodium dithionite-citrate-bicarbonate method of Aguilera and Jackson (1953) was used for this determination. A summary of this method follows. A known amount of the sample is placed in a 100-ml centrifuge tube and 40ml of 0.3M sodium citrate and 5ml of 1M sodium bicarbonate are added. The temperature is brought to 75-80 degrees C in a water bath, then 1g of solid sodium dithionite is added and the mixture is stirred constantly for one minute and then occasionally for 5 minutes. A second and the a third gram of sodium dithionite is added stirring the mixture as previously. After a 15 minute digestion period, 10ml of saturated NaCl solution and 10ml acetone are added to the tube to promote flocculation. The suspension is then mixed, warmed and centrifuged for 5 minutes at 1600-2200 rpm. The clear supernatant is decanted into a 500ml volumetric flask and the solution is kept for iron determination. Further analysis by using the colorimetric thiocynate method for ferric iron and the colorimetric orthophenanthrolic method for ferrous iron yielded their percent weight present in the sample. Ferric iron was calculated as Fe2O3 weight percent while ferrous iron was calculated as FeO weight percent. Further reference to Jackson (1953).

4). Aluminum content: Using the colorimetric aluminon method of Jackson (1953) aluminum was determined as weight percent in the sample. After a series of chemical reactions, the solution is mixed with 10ml of aluminon-acetate buffer reagent. The color is red at 5300 A after one hour or longer, and referenced to the standard Al curve to obtain the concentration of Al. $\%Al_2O_3 = \%Al \times 1.89$.