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POST BURIAL ALTERATION IN PALEOSOLS AND THEIR INFLUENCE ON SURFACE SOILS, BROWN COUNTY, KANSAS

by

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Randall John Schaetzl

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ABSTRACT

A previously unstudied paleosol in Brown county, Kansas, probably Late Sangamonian/Early Wisconsinan in age, was analyzed in exhumed and buried locations. The purpose of this research is to document post-burial alterations and to determine a depth below which processes affecting alteration cease. Organic matter content, pH, texture, and quartz/feldspar ratios were determined for paleosols at various burial depths (0-550 cm). The data were fitted to a theoretical model which predicts the depth at which paleosols become isolated from surface soil-forming processes. Results indicate that paleosol pH and organic matter content rapidly reach equilibrium with buried environments, and that the values of each change logarithmically with depth. Paleosol organic matter content decreases with burial depth and exhibited little change below about 200 cm, while pH values of paleosols peak at 6.4-6.5 at 150 cm. Increased weathering of shallowly-buried and exhumed paleosols results in higher maximal clay values than are present in deeply-buried counterparts. Changes in Q/F ratios in paleosols as functions of burial depth are not significant.

Surface soils exhibit increased organic matter content over deeply- as opposed to shallowly-buried paleosols. Most paleosols are dominated by 2:1 and 2:1/1:1 mixed layer clays, and contain more kaolinite than superjacent surface soils.

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INTRODUCTION

This thesis inquires into pedogenic processes that affect buried paleosols. Exhumed, shallowly- and deeply-buried paleosols are examined, and the changing pedogenic characteristics of the paleosols are then fit to a theoretical model. The model attempts to predict a depth at which a buried paleosol becomes isolated from specific surface soil-forming processes, e.g., lessivage, pedoturbation, and decomposition of primary minerals.

To this purpose, paleosols of Late Sangamon age in Brown County, Kansas were sampled at different burial depths (0-5.5 m). Overlying loesses of Wisconsinan age and the soils developed therein were also sampled, in an attempt to relate surface soil properties to depth to a buried paleosol.

The pronounced sparcity of investigations on post-burial alteration in paleosols and the recent calls for such studies (Gerasimov 1971, Yaalon 1971, Valentine and Dalrymple 1976) prompted the present research. Numerous investigators have noted that buried paleosols are altered from their pre-burial state, but only Hallberg, Wollenhaupt, and Miller (1978) and Mausbach, Wingard, and Gamble (1982) have specifically attempted to explain the amount of alteration, or the processes involved. The present investigation 1) leads to a fuller understanding of soil-forming processes at depth, 2) explains some of the post-burial alterations that occur in paleosols buried beneath loess, and 3) introduces the term "isolation", as opposed to burial, for paleosols so deeply buried as to be below the influence of normal surface soil-forming processes.

LITERATURE REVIEW

<u>Introduction to Paleosols</u>. A paleosol, as defined by Ruhe (1965), is a soil that formed on a landscape of the past. Paleosols have been categorized into three types: relict, buried, and exhumed (Ruhe 1956, 1965, 1969, Working Group on the Origin and Nature of Paleosols 1971, Leamy 1975, Valentine and Dalrymple 1976). Paleosols which have not been covered by younger sediments and remain today at the surface are termed relict. Buried paleosols occur where a paleo-geomorphic surface, with its associated soils, has covered by sediments. If a buried paleosol loses this protective cover of sediments through erosion, it becomes exhumed.

Morrison (1965) prefered the term "geosol" to paleosol. Geosol was used for fundamental soil-stratigraphic units whose stratigraphic interval (i.e., age relationship) is known with near certainty (See also Richmond and Frye 1957). Paleosols and weathering profiles were relegated to a lower category of geosols - ones whose stratigraphic (age) relationships cannot be established based on present knowledge, with sufficient precision to qualify them as true geosols. Paleosols may also be referred to as fossil soils (Nikiforoff 1943).

Relict soils are common in the United States. Ultisols (Red-Yellow Podzolic) of the southeastern states are thought to be relict soils, as are the Mollisols (southern reddish prairie soils) of the Midwest (Thorp, Johnson, and Reed 1951, Simonson 1954, Frye and Leonard 1965, Ruhe 1965, 1974, Ruhe, Hall, and Canepa 1974).

Both soils exposed at the surface and can be traced beneath the loess and till sheets in the upper Midwest (Ruhe 1974, Ruhe _et al_. 1974), where they are the buried Sangamon soil.

Sangamon is the name given to the soil that allegedly developed during the Sangamonian Interglacial stage (Frye and Leonard 1965, Follmer 1978). There is considerable debate in the literature about the date and duration of the Sangamonian Stage. Difference of opinion also exists concerning the placement of the oft-identified Sangamon soil into the Midwestern glacial chronology. The Sangamonian Interglaciation falls between the Illinoian and Wisconsinan Glacials and was a time characterized by stable land surfaces and intensive soil formation (Caspall 1970). Estimates of the duration of the Sangamonian Stage derived from deep sea core data range from 20,000 (Emiliani 1966a, b) to approximately 210,000 years (Ericson and Wollin 1968). A commonly accepted estimate from paleosol evidence placed the end of the Sangamon well before 38,000 years B.P. (Ruhe and Scholtes 1956, Ruhe 1969). Richmond and Fullerton (no date) suggested the Sangamonian Interglaciation occurred between 130,000-120,000 years B.P. It seems unlikely that a soil as strongly developed and as widespread as the Sangamon could have formed in 10,000 years, even assuming greater intensity of soil development than that of the present. Extrapolation of Pleistocene temperatures from oceanic core data must be made with caution, because the ocean basins and the Pleistocene type areas in the Midcontinent region are so widely separated. (Ruhe 1969).

A soil commonly identified as the Sangamon is widespread throughout the Midwest, from Ohio through the type area in Illinois to southwestern Texas, a distance approaching 2000 miles (Frye and Leonard 1965). It varies in morphology from a forested profile in Illinois and other areas east of the Missouri River (Simonson 1941, Thorp *et al*. 1951, Frye and Leonard 1952, Simonson 1954, Ruhe and Scholtes 1956, Ruhe 1965, 1968, 1969, 1970, 1974, Ruhe *et al*. 1974, Ill. St. Geol. Survey 1979) to a prairie profile in the Great Plains region (Simonson 1941, Thorp *et al*. 1951, Frye and Leonard 1952, 1965, Ruhe 1965, 1974). Follmer (1978, 1979) has compiled excellent historical reviews of the development of thought about the Sangamon soil for Illinois.

In northeastern Kansas a soil believed to be the Sangamon is often seen developed in Loveland (Illinoian) loess. In all likelihood, the formation of this soil spans a period of time greater than that ascribed to the Sangamon Interglaciation. Its development probably began with the slow cessation of loess deposition during the Illinoian, continued throughout the Sangamon, and concluded sometime in the Altonian substage of the Wisconsinan Glacial, when Peorian loess deposition outpaced soil development.

The Sangamon soil in Kansas is "widespread, well-developed, and exceptionally well preserved" (Frye and Leonard 1952, p. 119). It has its maximum development in Kansas in the area adjacent to the Missouri River (Frye and Leonard 1952). Here the soil is leached of $CaCO_3$ to depths as great as 7.3m. (Frye and Leonard 1952). It has a reddish-buff color (Frye and Leonard 1949) with a deep profile (2.3m), and a thick (0.6m), clayey B horizon (Frye and Leonard 1965). It is a Reddish Prairie to Reddish Chernozem (Mollisols) soil (Frye and Leonard 1965) commonly developed in Loveland loess on the uplands and, if not exhumed, buried by Peorian or later loesses (Frye 1949, 1951, Frye and Leonard 1949, 1952, 1965, Daniels, Handy and Simonson 1960, Caspall 1970).

The Sangamon soil is usually more strongly developed than its postglacial surface analogues (Frye and Leonard 1952, Ruhe 1956, 1965, 1969, Hall 1973, Ruhe et al. 1974, Al-Barrack 1980), though exceptions occur (Mausbach et al. 1982). Strongly developed, as used here, implies several things: thicker solum (Simonson 1941, Frye 1949, Frye and Leonard 1965, Ruhe 1956, 1965, 1969, Hall 1973, Ruhe et al. 1974, Bushue, Fehrenbacher and Ray 1974, Al-Barrack 1980), thicker B horizon (Thorp et al. 1951, Ruhe 1956, 1969, Hall 1973, Ruhe et al. 1974), redder hues and stronger chromas (Thorp et al. 1951, Scholtes, Ruhe and Riecken 1951, Frye and Leonard 1952, Simonson 1954, Ruhe and Scholtes 1956, Ruhe 1956, 1965, 1968, 1969, 1970, 1974, Ruhe et al. 1974, Hall 1973, Al-Barrack 1980) and higher percentages of clay in the B horizon (Simonson 1941, Frye and Leonard 1949, Ruhe 1956, 1965, 1969, 1970, Badgley 1957, Caspall 1970, Ruhe et al. 1974, Bushue et al. 1974, Ruhe and Olson 1980) than overlying surface soils.

Documented Sangamon paleosols are exhumed on some hillsides in Kansas, Illinois, Nebraska, and Iowa (Thorp et al. 1951, Frye and Leonard 1952, Ruhe 1956, 1969, 1970, Ruhe and Scholtes 1956, Prill and Riecken 1958, Ruhe and Daniels 1958, Millet and Drew 1963, Morrison 1965, Runge 1973, Bushue et al. 1974). In Iowa and Nebraska, the exhumed paleosol is locally mapped in the Adair series (Millet and Drew 1963, Ruhe 1970) or as a variant of the Lindley series (Prill and Riecken 1958). In Illinois, Bushue et al. (1974) reported that exhumed Sangamon paleosols on hillslopes were mapped in the Hickory series (Typic Hapludalf). Eickelberry and Templin (1960) mapped the exhumed Sangamon paleosol in Brown County, Kansas as Morrill (Typic Argiudoll).

Identification of a buried soil as a paleosol is viewed by Ruhe as "no problem" (1965, p. 755), whereas Ruellan considered positive identification to be "rarely simple and irrefutable" (1971, p. 9). Investigators disagree as to what criteria to use when identifying a stratigraphic layer as a paleosol. There is general agreement, however, that pedological methods must be used (Working Group 1971). Members of the Working Group (1971) have identified pedologic criteria which are most useful in the recognition of paleosols; these are listed below (Other investigators who have noted a method's usefulness in this regard are mentioned):

1) Macroscopic (field) evidence, preferably the demonstrated presence of a B horizon. (Birkeland 1974)

2) Evidence of pedogenic accumulations, e.g., clay, carbonates, iron, silica, organic matter. (Birkeland 1974)

3) Presence of a soil fabric or structure, using micromorphological analyses. (Nikiforoff 1943, Fedoroff 1971, Dalrymple 1964, Protz, Wilding and Thorp 1973)

4) Color.

5) Mineralogy. (Ruhe 1956, Brophy 1959, Yaalon 1971, Birkeland 1974)

Ruhe (1965) and Morrison (1965) stressed stratigraphic and geomorphic means of identification which take into account lateral changes in the paleosol. Valentine and Dalrymple (1975) suggested the same when they emphasized that the crucial test of a buried soil landscape must be the existence of a "paleocatena."

Though different authors favor one method of analysis over another, most agreed that several different analyses, used in conjunction with each other, were better than one (Yaalon 1971, Working Group 1971, Protz et al. 1973, Birkeland 1974). A variety of laboratory methods applied in combination with field identification provide quantitative evidence of paleopedogenesis (Working Group 1971), something deemed necessary by Bos and Sevink (1975) in order to make identification more objective and less subjective.

Buried Paleosols and Post-Burial Alteration Processes. All buried paleosols have been altered more or less from their original state, both during and after burial. The degree and extent of paleosol alteration is a major emphasis of this thesis but, as Ruhe (1965, p. 759) noted, "unfortunately most studies of paleosols are mainly descriptive." Numerical and quantitative data are needed for detailed comparisons of surface soils and paleosols (Ruhe 1968), from which one might deduce more accurately the original properties of the paleosol. Ruhe (op. cit.) also called for these studies to be areally continuous over the landscape in order to arrive at a "better understanding of the variations in soil properties."

Before one can interpret post-burial alteration products, one must first examine changes which occur during burial. Ruellan (1971) and Simonson (1954) noted that buried soils are always modified by the process of burial itself.

Burial of soils by loess is often a slow process. Loess deposition rates have been estimated at one foot in 1400 years, to as rapidly as one foot in 121 years (Ruhe, Miller and Vreeken 1971). The first Wisconsinan loess to be deposited on the Sangamon soil in Nebraska and Kansas "seems to have collected very slowly" (Thorp *et al*. 1951, p. 8; See also Leonard 1952). During slow loess deposition, soil formation can keep pace as the soil surface migrates upward with gradual loess additions (Thorp *et al*. 1951, Stevenson 1969, Valentine and Dalrymple 1976). The result is thick, dark, granular, silty A1 horizons up to three feet thick (0.3m), according to Thorp *et al*. (1951). In some cases the A1 horizon will not thicken substantially, but rather migrate upward into the loess as the B horizon below thickens upward (Stevenson 1969). The reason for this is not completely understood, but may depend on loess depositional rates. Eventually loess deposition outpaces soil formation and the soil becomes isolated from surface soil-forming processes.

Burial of soils by loess is a time and space transgressive process (Vreeken 1975a, b). Paleosols buried slowly by loess often display this silty material intermixed in their uppermost (A, and occasionally A and B) horizons (Simonson 1941, Smith 1942, Nikiforoff 1943, Brophy 1959, Millet and Drew 1963, Bushue *et al*. Valentine and Dalrymple 1976, Ill. St. Geol. Survey 1979), making the top of the buried soil difficult to identify (Valentine and Dalrymple 1976).

Truncation of part of a solum often preceeds or alternates with loess depositional episodes. The Sangamon soil is observed to be truncated at numerous locations in the Midwest (Thorp *et al*. 1951, Frye and Leonard 1949, 1952, Hogan and Beatty 1963, Caspall 1970, Frye *et al*. 1974, Vreeken 1975a, b) and in Indiana (Mausbach *et al*. 1982). Dry conditions preceding and during loess deposition effectively reduce the humus content of the A horizon as the soil adjusts to a new equilibrium organic matter content. The lowered organic matter content of the surface soil would make it prone to deflationary processes (Stevenson 1969). Yaalon (1971), however, questioned the argument of Stevenson (1969). He suggested that "the absence of an A horizon is quite likely to be as much due to post-burial destruction of the organic matter as to oft-postulated erosion" (p. 34). Sangamon profiles have been observed which appear not to be truncated (Frye 1949, Ruhe 1956, Bushue et al. 1974, Daniels and Handy 1959).

The problem of identification of a paleosol as such is compounded where it is buried at insufficient depth to isolate it from surface soil-forming processes. Soils in which a paleosol is present within the solum of the surface soil have been called "polypedomorphic" (Bos and Sevink 1975), "multistory" (Morrison 1965), "welded" (Ruhe and Olson 1980), "polymorphic" (Simonson 1978), "polygenetic", and "polycyclic." Often, however, the paleosol is misinterpreted as a weathering zone and not a paleosol, per se, (Buntley, Daniels, Gamble, and Brown 1977), an argillic horizon of the surface soil, or a simple lithologic change, designated by the Roman numeral II in profile descriptions (Ruhe and Olson 1980). Discussions of welded profiles are found in Bryan and Albritton (1943), Williams (1945), Simonson (1954), Walker (1962), Morrison (1965), Thorp (1965), Ruellan (1971), Bos and Sevink (1975), Ruhe and Olson (1980), Thompson, Smeck and Bigham (1981), Mausbach et al. (1982) and Darmody and Foss (1982). Examples of modern soils whose lower horizons have formed in a paleosol abound, but the reader is directed to the above authors, and to Thorp et al. (1951), Prill and Riecken (1958), Hole (1976), Buntley et al. (1977), Hawando (1978), Hallberg et al. (1978), Gould, Anderson, McClellan and Gurnsey (1979) and Simonson (1982) for further descriptions.

A thin covering of overburden (till, loess, colluvium, etc.) on a paleosol affects it in several ways. The overlying material may be mixed into the paleosol during burial, or later by pedoturbation (Smith 1942), by lessivage of clays and silts in voids (Hallberg et al. 1978), or both (Hutcheson and Bailey 1965, Price, Blevins, Barnhisel and Bailey 1975). Simonson (1954) suggested that a thin loess covering over a well-developed soil can result in thickening of the A horizon during loess deposition followed by upward movement of the B horizon (See also Williams 1945 and Birkeland 1974). In deeper loess the B horizon of the surface soil might fuse with the Bb horizon, resulting in an abnormally thick total B horizon (Simonson 1954).

The difficulty of locating the lithologic discontinuity in a welded soil profile has been given much discussion. Price et al. (1975) used quartz/feldspar ratios to locate the discontinuity. Thompson et al. (1981) found that clayfree particle size profiles and total silt titanium/zircon values generally supported the field identification of a discontinuity, whereas Ruhe and Olson (1980) explored several physical and mineralogical properties of the solum to identify discontinuities in welded profiles (See also Mausbach et al. 1982).

"All these forms of postburial alteration to buried paleosols raises the question of how deep is 'buried'" (Valentine and Dalrymple 1976, p. 211), i.e. "isolated." A recurring question asks, how deep does a paleosol have to be buried to be below the zone of soil formation? Obviously the answer changes as a function of any and all of the soil-forming factors (Jenny 1941, Crocker 1952). A paleosol buried

25 meters beneath the surface in a landscape currently producing Oxisols would be well within the surface weathering profile, whereas a meter may be deep enough to effectively "isolate" a paleosol in a landscape producing Cryorthents today. One must also determine whether a buried soil needs to be below the solum (A and B horizons), soil profile (A, B, and leached C horizons) or weathering profile (soil profile plus unleached, oxidized layers beneath) in order to be truly isolated. No definitive statement is available as to whether or not buried paleosols can ever be effectively isolated from surface soil-forming processes. Deeply buried paleosols may be below the zone of surface soil-forming processes, but still be affected by diagenesis (Morrison 1965, Birkeland 1974, Ill. St. Geol. Survey 1979) which affects sediments at all depths. It is, therefore, difficult to state at what depth, under what conditions, a paleosol is totally isolated from soil-forming processes.

Bos and Sevink (1975) labeled a stratigraphic section containing a very deeply buried paleosol a composite profile. It contained the surface soil and the paleosol; in this case both were monopedomorphic soils separated by non-soil material (unaltered primary sediment). Morrison (1965) defined a "compounded profile" as one in which two or more different geosols are superposed, but the profiles do not overlap. Ruellan (1971) suggested that paleosols deeply buried beyond the present zone of direct biological action are the "only true paleosols" (p. 8).

The problem remains, however, with these definitions: a depth must be established at which one can call a soil "truly buried" (Ruellan 1971, p. 8). Wascher, Humbert and Cady (1947, p. 393) reported that a "thick cover of loess" protects a paleosol from weathering, where as Morrison (1965, p. 23) was more specific: "A few feet of covering sediment essentially removes the geosol from erosive, pedogenic, and direct biologic agencies of the surficial environment." Perhaps Valentine and Dalrymple (1976, p. 217) reasoned best:

> It is very difficult to set a minimum depth limit below which a paleosol may be regarded as truly buried. Hence, it is possible to make a case for a gradation of buried paleosols from those covered by many meters of material to those covered by only a few centimeters...

Buntley *et al*. (1977) have evidence for such a gradation of burial depths over a paleosol in Tennessee, as do Bushue *et al*. (1974) in Illinois and Prill and Riecken (1958) in Iowa.

Post-Burial Alteration Effects on Selected Soil Properties. Prior to this point, discussion has focused on depth of burial as an independent variable necessary for paleosol identification. An evaluation of the effect of burial, held at a constant depth, on certain pedogenic parameters, follows.

Alteration subsequent to burial has been noted by many investigators (Nikiforoff 1943, Simonson 1954, Ruhe 1956, 1969, Yaalon 1971, Gerasimov 1971, Valentine and Dalrymple 1976, Hallberg *et al*. 1978, Mausbach *et al*. 1982), but only Hallberg *et al*. have been able to quantify the changes, and then only over a period of 100 years. The paucity of quantitative data in the literature has prompted Yaalon (1971, p. 35) to state that "there is an urgent need for studies on the rate of alteration and preservation of pedologic features in paleosols."

The importance of post-burial alteration processes cannot be overestimated for "only from studies of buried examples of known soil and nonsoil materials and the determination of the relative stability of their features under various conditions" can one arrive at a clear understanding of the features which connote pedogenesis and which connote diagenesis (Valentine and Dalrymple 1976, p. 219). Yaalon (1971), however, cautioned that diagnosis of antecedent pedogenesis is always subject to uncertainties.

Some pedogenic properties are more lasting than others. The degree of transformation of relict paleosolic properties is most important in studies of paleosols (Gerasimov 1971). Yaalon (1971) has grouped pedogenic features into three stability classes:

1) Relatively rapidly adjusting features which approach dynamic equilibrium with their environment in less than 1000 years, e.g., Mollic, Salic, and Cambic horizons, Slickensides, and Gilgai.

2) Slowly-adjusting features which approach equilibrium in 1000-10,000 years, e.g., Umbric, Spodic, Calcic, Gypsic, and some Argillic horizons, Fragipans, and Mottles.

3) Persistent features which are irreversible, e.g., Oxic, Placic, Petrocalcic, Argillic, Natric, Albic, and Histic horizons, Plinthite, Durinodes, and Gypsic Crusts.

In addition to the characteristics of the paleosol which are affected by post-burial alteration, the effect of the paleosol itself on pedogenic transformations within the surface soil is often ignored. The paleosol usually represents a somewhat less permeable layer in the pedosphere (Hall 1973, Al-Barrack 1980, Maubach et al. 1982). An impervious substratum may impede drainage, and retard leaching and weathering processes (Smith 1942), or may promote weathering of materials above (Ruhe 1973, Al-Barrack 1980). Mausbach et al. (1982, p. 367) found little appreciable leaching of bases into a buried paleosol and cautioned that "the magnitude of change in hydraulic conductivity across an unconformity is an important factor when dealing with resaturation of buried paleosols and should be considered in future studies on alteration of buried soils." Ruhe (1969), Vreeken (1968), and Coleman and Fenton (1982) discussed the influence of paleosols on water infiltration, particularly with regard to perched water tables above the buried soil. Perched water tables may result in gleyed zones above and/or within the buried solum (Ruhe 1969, 1970) or deoxidized zones above (Al-Barrack 1980). Saturated zones above buried paleosols have been reported by Caspall (1970), Wascher et al. (1947), Ruhe (1969), and Vreeken (1975a), and may lead to slump of loess above (Caspall 1970).

Simonson (1954, p. 730) stated that whether soils are preserved when buried depends upon "the kinds of horizons in the profile, the degree of horizon differentiation, and the processes responsible for burial." One must also consider the environmental changes which may occur after burial. Faint horizons are not likely to persist (Simonson 1954), especially if the change is to a stronger weathering environment (Morrison 1965), unless deeply buried.

Post-burial changes occur with regard to color, organic matter content, carbonate content, mottling, soluble salt content, consistence, pH, clay mineralogy, base saturation, micromorphology, bulk density, etc., though the rates of change vary. The following discussion focuses primarily on those properties which are analyzed in this thesis. The author does not intend to mean that these are the sole properties appropriate for a paleopedologic study. He is well aware of the advantages of the study of other properties such as phosphorous (Smeck 1973, Runge, Walker, and Howarth 1974), amino acids (Goh 1972), micromorphology (Brewer and Sleeman 1969, Fedoroff 1971, Valentine and Dalrymple 1976, Mausbach _et al_. 1982), iron oxide content (Bronger 1974, Hogan and Beatty 1963, Yaalon 1971, Ruhe 1974), and opal phytoliths (Dormaar and Lutwick 1969), among others, in studies of paleopedology.

It was mentioned earlier that the Sangamon paleosol in the Midwest has redder hues and stronger chromas than surface soil analogues. This coloration occurs regardless of parent material (Ruhe _et al_. 1974). The red color has been attributed to the presence of hematite in substantial amounts (Ruhe 1970), which occurs as a byproduct of silicate and ferromagnesian decomposition. Accumulations of iron sesquioxides have been ascribed to either a warmer climate of formation or a long period of time under a climate similar to that of the present (Thorp _et al_. 1951, Frye and Leonard 1952, Simonson 1954, Thorp 1965, Ruhe 1965, 1969). Color, however, is not always a reliable guide for paleosol recognition or paleoclimatic reconstructions (Ruhe 1969, Birkeland 1974), for it can be inherited from the parent material (Ruhe 1970), or color variations may be produced by "deep subsurface weathering associated with ground water movement in layered sediments" (Valentine and Dalrymple 1976, p. 212). Dark color in paleosols cannot be attributed solely to organic matter (Daniels et al. 1960, Turchenek, St. Arnaud, and Christiansen 1974). Manganese has been shown to darken paleosols without the presence of even moderate amounts of organic matter (Daniels _et al_. 1960). Soil color is an unreliable tool to use in post-burial alteration studies or paleoenvironmental reconstructions. No guidelines for recognizing post-burial color changes have been accepted (Working Group 1971).

Organic matter and organic carbon content are often-used properties in paleosol identification because soils accumulate organic matter naturally and upon burial, organic oxidation occurs at varying rates depending on local conditions (Stevenson 1969). Organic carbon amounts are sometimes higher in the buried soil than in the sediment above and below (Daniels and Handy 1959, Daniels et al. 1960, Turchenek et al. 1974, Valentine and Dalrymple 1976). Buried A horizons may (if present) contain the highest amount of organic carbon of the paleosol horizons (Simonson 1941, Turchenek et al. 1974, Sanborn and Pawluk 1980), or they may contain no more than lighter colored horizons below (Bushue et al. 1974). The extent to which organic matter is preserved depends on such factors as the change in environment which preceded the new cycle of sedimentation, the circumstances under which the soil was buried, the activities of living organisms in the buried sediments (Stevenson 1969), and the depth of burial. Deeply-buried Sangamon A horizons which have become degraded by post-burial alterations have been classified as C/A horizons (Ill. St. Geol. Survey 1979).

Organic materials become highly humified upon decomposition (Brady 1974). Humification is the name given the set of processes involved in the decomposition of organic matter, leading to humus formation. These processes produce material increasingly resistant to breakdown, which Gould et al. (1974) concluded is less suitable for soil microflora and microfauna than partially humified materials. Stevenson (1969) suggested that, in part, it is the association of humus with clay which protects the former from total decomposition, though Turchenek et al. (1974) attributed low organic matter solubility to the formation of calcium humates. Nevertheless, the humus fraction of the soil, because it reaches equilibrium with the environment rapidly (See, for example, Chandler 1942 or Hallberg et al. 1978), is the soil component most susceptible (with the possible exception of pH) to change after burial (Stevenson 1969, Working Group 1971). Organic decomposition is most rapid in shallowly-buried soils (Turchenek et al. 1974). In Newgrange County, Ireland, 75% of the organic carbon in a soil buried since the Neolithic was depleted over a period of approximately 400 years, and probably much earlier (Gardiner and Walsh 1966).

Simonson (1941) found total carbon contents of horizons in three buried paleosols (1.8-3.2m) in Iowa to be nowhere greater than 0.5%, and usually less than 0.25%. Ruhe et al. (1971) examined the organic carbon content within incipient paleosols ("loess bands"), and immediately above and below the bands. They found little difference in the organic carbon contents of the incipient paleosol bands with those of the underlying and overlying loesses (See their Table 2, pp. 53-54). Organic carbon content of the "loess bands" varied from 0.15% to 0.083%. Daniels et al. (1960), on the other hand, found loess bands in Iowa to contain 10-15% more organic matter than the average of the loesses above and below the bands (0.272 ± 0.034% vs 0.259 ± 0.023%). Turchenek et al. (1974) reported organic carbon contents as high as 0.91% in buried A horizons from paleosols dated at approximately 7000 BP. They also suggested that dark color (Munsell values of 3) can be retained in paleosols with very low organic carbon contents. Hallberg et al. (1978) reported a 0.9% to 1.2% total decrease in organic carbon in a Mollisol buried by 60 cm of spoil within a 100-year span of time. They ascribed most of the decrease to rapid degradation of fibrous organic material after burial.

Studies of organic matter contents of paleosols reflect post-burial losses rather than original pedogenic characteristics (Turchenek et al. 1974). Rates of decomposition vary. Lastly, no mention is made of humus translocation into or out of a buried solum, though the former must occur in shallowly-buried soils.

Clay contents of the Sangamon paleosol B horizon, as stated earlier, are usually greater than contemporary surface soil B horizons. Ruhe et al. (1974) confirmed this statement for the Sangamon soil when formed in various parent materials and covered by soils developed in loess (See also Wascher et al. 1947, Hall, 1973, Ruhe 1974, and Ill. St. Geol. Survey 1979). It is generally agreed that higher clay contents in deeply buried soils are due to formation before burial rather than illuviation from above, though the latter has been reported (Hallberg et al. 1978, Mausbach et al. 1982). Williams' (1945) data suggested that even 0.3-0.6m of overlying loess prevent surface lessivage into a buried solum. Hallberg et al. (1978) observed illuviation of silts and clays into a buried soil in only 100 years. Runge (1973) demonstrated that clay, when moving downward in a pedon and reaching a calcareous layer, will floculate and move no farther. Therefore, clay theoretically cannot illuviate into a buried paleosol through calcareous loess, as calcium ions favor deposition of clay particles in voids (Hole 1976). The loess must first be leached of bases.

The fine clay fraction (<0.2 μ) is thought to be a better indicator of lessivation than total clay content of the Btb horizon (Brophy

1959, Sanborn and Pawluk 1980, Mausbach _et al_. 1982). Micromorphological studies aid in the identification of Btb horizons (Brewer and Sleeman 1969, Sanborn and Pawluk 1980). Simonson (1954, p. 730) suggested that distribution curves for clays are "persistent features which can be expected to last a long time" and Ruhe and Olson (1980) concurred (See also Morrison 1965, Yaalon 1971, and Mausbach _et al_. 1982). A great deal of the stability of an argillic horizon, however, is dependent upon depth of burial, horizon development, and local conditions.

Because loess and till are very often calcareous, buried paleosols are commonly enriched with bases and carbonates above preburial amounts (Simonson 1941, Ruhe 1949, 1956, 1965, 1969, 1974, Badgley 1957, Prill and Riecken 1958, Bronger 1969, 1974, Morrison 1965, David 1966, Tuchenek _et al_. 1974, Ruhe _et al_. 1974, Hawando 1978, Sanborn and Pawluk 1980, Ruhe and Olson 1980), though Mausbach _et al_. (1982) gave evidence to the contrary for the Sangamon soil in Indiana. Base saturation within the paleosol is observed even where the loess or till above are not presently calcareous (Simonson 1941, Frye and Leonard 1949). The non-calcareous (though base saturated) nature of some paleosols has been used as a criterion for field identification (Frye and Leonard 1949, David 1966, Caspall 1970). Conclusive evidence of carbonate enrichment is found where carbonate nodules descend from the overlying loess into the buried soil (Ruhe 1965).

The pH of paleosols is a very transient property (Simonson (1941, Ruhe 1956, 1965, Yaalon 1971, Ruhe _et al_. 1974, Sanborn and Pawluk 1980). Soil reaction almost invariably becomes more alkaline after burial by loess (Ruhe 1956, 1965, Badgley 1957). David (1966) demonstrated that the pH-vs-depth curve for a buried paleosol closely coincides with the carbonate-vs-depth curve, though this finding is not consistent with the data of Hawando (1978) in which a B3ca horizon with a pH of 7.8 overlies a paleosol with a pH of 7.0-7.2. Ruhe (1965, p. 757) insisted that "soil chemical analyses may be meaningless as representing the original chemical nature of the paleosol," and has stated this opinion often (1956, 1965, 1974, Ruhe and Olson 1980). Soil reaction-vs-depth curves for soils in the Midwest are usually lowest at the surface and increase in value to some equilibrium level at depth. A paleosol may or may not affect this curve.

Exhumed Paleosols and Post-Exhumation Alteration. If the paucity of post-burial alteration studies in the literature is surprising, the greater lack of post-exhumation research is even more so. Post-burial alteration of paleosols involves many detailed and poorly understood processes. Post-exhumation studies must deal with these problems, compounded by the increasing importance of pedogenic processes as the paleosol becomes exhumed.

Lack of quantitative data about exhumed paleosols is not due to a shortage of pedons for study. The Sangamon and Yarmouth/Sangamon soils crop out on many hillsides throughout the Midwest. Nor is lack of data due to disinterest in exhumed paleosols and their lateral variation (especially as compared to juxtaposed buried paleosols). Valentine and Dalrymple (1976) and the Working Group (1971) have repeatedly called for such studies.

The study of post-exhumation alterations of soils demands that one examine pedologic properties which are "self-terminating" or "irreversible" (Yaalon 1971, p. 26). Vreeken (1975a) called for the interpretation and comparison of persistent pedologic features of paleosols with surface analogues. These properties yield insight into the original characteristics of the paelosol because they have not theoretically been changed significantly by either burial or exhumation. The best properties to use in this regard are mineralogical.

Mineralogical weathering has been used as a pedodiagnostic tool by many authors (Haseman and Mashall 1945, Wascher et al. 1947, Brophy 1959, Millet and Drew 1963, Bhattacharya 1962, 1963, Ruhe 1956, 1960, 1969, Willman, Glass and Frye 1966, Hall 1973, Hawando 1978, Frye et al. 1974, Mausbach et al. 1982) and lauded as a tool by others (Nikiforoff 1943, Whiteside 1965, Yaalon 1971). Mineralogical information is useful 1) in determining the origin of the soil parent material, 2) as an index of weathering, 3) in recognizing depositional differences in horizons of the profile, and 4) as a guide in soil classification (Al-Barrack 1980). The principles behind mineralogical methods in studying paleosols are briefly outlined by Bhattacharya (1963, p. 793):

> ...under weathering conditions resistant species increase in relative abundance while less resistant ones are eliminated. A change in relative abundance of the minerals is, therefore, a measure of their resistance to weathering.

Frye et al. (1974) cautioned, however, that mineralogical weathering ratios are only meaningful when comparing soils of similar parent materials.

The weathering stability of common silicate minerals was outlined by Goldich (1938). Pettijohn (1941) and Bhattacharya (1963) formulated heavy mineral stability sequences (Table 1).

A comparison of the frequency of minerals resistant to weathering with those more susceptible, in the coarse silt or fine sand fractions of soils, gives a good indication of the relative degree of weathering therein. Comparisons can be made between paleosols, between paleosols and surface soils, or between horizons in a single profile. Ruhe (1956) used Zircon + Tourmaline/Amphibole + Pyroxene and Quartz/Feldspar ratios (See also Ruhe 1969) for soils and paleosols in Iowa. In general, he found older paleosols were more intensively weathered than younger paleosols and modern soils. Horizons nearer the surface were more weathered than deeper horizons, which was also noted for a Sangamon soil in Illinois (Frye _et al_. 1974). Millet and Drew (1963) also used Z+T/A+P weathering ratios for two soils, one formed in Peorian loess and one in a thin loess admixture within the A horizon of an exhumed paleosol. The more strongly weathered B horizon of the latter surface soil (A horizon of the paleosol) was quite evident. Brophy (1959) compared the effectiveness of Zircon + Tourmaline/Garnet, Garnet/Hornblende, and Zircon + Tourmaline/Hornblende ratios and concluded that garnet is intermediate in resistance to weathering, falling between the Zircon and Tourmaline group, and Hornblende. All ratios were however, useful in his studies of buried soils and their maturity. Hawando (1978) used Q/F ratios in comparing surface soils to buried paleosols and found differences in ratios, depending on size fraction used (coarse sand, fine sand, silt). Haseman and Marshall (1945) pioneered some

Table 1

Stability Sequences of Light and Heavy Minerals in Response to Weathering

Source:	Goldich 1938	Pettijohn 1941	Bhattarharya 1963
Least Stable	Olivine, Anorthite	Olivine	Pyrite
	Pyroxenes, Na-Ca Plagioclase	Augite	Pyroxenes
	Amphiboles, Na-Ca Plagioclase	Sphene	Amphiboles
	Biotite, Albite	Hornblende	Epidote
	K-Feldspars	Kyanite	Garnet
	Muscovite	Staurolite	Tourmaline
	Quartz	Magnetite	Magnetite
		Apatite	Zircon
		Biotite	
		Garnet	
		Tourmaline	
		Zircon	
Most		Rutile	
Stable		Muscovite	

Note: Sequences are relative. Cross-referencing from one weathering sequence to another in this table is not advised.

quantitative mineralogical techniques in studies of soil genesis and profile homogeneity. They used complete mineralogical analyses at numerous depths in soil profiles and for different size fractions, while relying heavily on the resistance of zircon to weathering as a pedogenic indicator species. Bhattacharya (1963) and Wascher et al. (1947) also used mineralogical analyses without computing ratios (See also Graham 1949, St. Arnaud and Whiteside 1963, and Frye et al. 1974).

Clay mineralogy is an excellent tool for use in studies of soil weathering, and parent material homogeneity and source. Ruhe and Olson (1980) stated that both texture and clay mineralogy in deeply buried paleosols are relatively unaffected by burial and thus represent the "true nature of the buried soil" (p. 135). Phyllosilicate clay structure is extremely stable at the earth's surface and is therefore very useful in Quaternary studies (Pedro, Jamagne, and Begon 1969). Only when the clay mineral suite has been altered by illuviation (Hutcheson and Bailey 1965, Hallberg et al. 1978) is clay mineral analysis not an excellent instrument for weathering studies of paleosols. One must also be aware of the variability of the clay mineral suite due to original parent material heterogeneity (Brophy 1959, Wascher et al. 1947, Hall 1973, Ruhe et al. 1974, Olson, Brunson, and Ruhe 1978, Mausbach et al. 1982), and even in such cases, clay mineralogy can aid in discovering the location of any lithologic discontinuities within the profile. Often, however, clay mineral differences due to lithologic changes are masked by more intense weathering of the upper horizons, or within a paleosol in the solum (Ruhe et al. 1974).

Many investigators have used clay mineralogy in comparing the weathering of a paleosol with that of a surface soil (Wascher et al. 1947, Brophy 1959, Bhattacharya 1962, 1963, Millet and Drew 1963, Hall 1973, Ruhe et al. 1974, Frye et al. 1974, Hawando 1978, Olson et al. 1978, Al-Barrack 1980, Ruhe and Olson 1980, Mausbach et al. 1982), while others have used the technique to arrive at a more complete understanding of the clay mineral suite of a paleosol (Badgley 1957, Hogan and Beatty 1963, Caspall 1970). Ruhe (1965) briefly reviewed the importance of clay mineralogy to Quaternary studies and Pedro et al. (1969) provided a more in-depth treatise on the subject.

X-ray diffraction is the standard method used to determine the clay mineralogy of a soil sample, and usually is interpreted qualitatively. Differential thermal analysis, X-ray florescence, staining, and electron microscopy have also been used. Several investigators have attempted to quantify the diffraction data (this topic is discussed in more detail later in this report). By calculating the height of the peak and the area beneath, Hall (1973), Ruhe et al. (1974) and Hawando (1978) arrived at quantitative assessments of the clay mineral composition of paleosols and surface soils (See also Glass et al. 1968, Olson et al. 1978 and Ruhe and Olson 1980). Ruhe et al. (1974), using a technique pioneered by Willman, Glass, and Frye (1966), calculated a diffraction ratio (DI) which compares the counts/second of the 10 A spacing on the diffractogram for illite with that of the 7.2 A spacing for kaolinite + chlorite. In many ways the DI is similar to Ruhe's (1956) weathering ratios, but uses a different size fraction of the soil to arrive at a more complete picture of the weathering aspects of the soil.

Weathering sequences for clay minerals vary with climate. Jackson (1964) compiled a sequence (Fig. 1) which shows many of the clay mineral reactions in soils. Kaolinite is typically the end member of the weathering sequence under humid mid-latitude climates. Therefore, it often occurs in greater abundance in strongly weathered paleosols than in the loess or surface soils above (Wascher et al. 1947, Brophy 1959, Hall 1973, Ruhe et al. 1974, Frye et al. 1974, Hawando 1978, Olson et al. 1978, Ruhe and Olson 1980), though exceptions are known (Mausbach et al. 1982). Brophy (1959) reported nearly complete illite and chlorite alteration to montmorillonite when progressing upward in a buried Sangamon till profile. Ruhe et al. (1974) and Hall (1973) noted stronger illite and chlorite depletion in the Sangamon soil in Indiana than in the surface soils above, while Olson et al. (1978) gave evidence to the contrary for

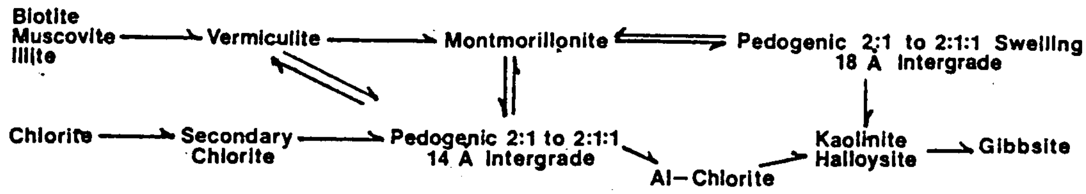

Figure 1. Clay mineral weathering sequences. (From Jackson 1964).

illite depletion. At one location, a Sangamon paleosol in Illinois exhibited pronounced illite depletion with a corresponding increase in expandable clays (Ill. St. Geol. Survey 1979), whereas at several others expandable clays were depleted in favor of kaolinite (Frye et al. 1974). Daniels et al. (1968) observed no appreciable clay mineral differences between dark colored bands (incipient paleosols) and the intervening loesses, while Badgley (1957) reported an increase in illite in a buried Sangamon profile, attributing it to leaching of potassium from loess above and redeposition within the 12-fold coordination sites of the clays. He referred to this process as a "reversal of the weathering sequence." Mausbach et al. (1982) observed a buried Sangamon B horizon that is less weathered than the B22t of the surface soil, based on clay mineralogy.

Exhumed paleosols become resubjected to surface soil-forming processes. Thus, after a period of time they become leached of bases, gain organic matter, lose clay from E horizons, etc. Bushue et al. (1974) found no great chemical differences between exhumed paleosols and modern till soils, although the exhumed soils still retained a somewhat higher base saturation below one meter. The profile with the highest concentration of bases was presumed to have been exhumed most recently. Prill and Riecken (1958) also postulated that some exhumed soils have high pH and base saturation values due to only recent exhumation. Because pH, free Fe_2O_3 content, and Ca and Mg values were all similar for an exhumed paleosol and a modern soil, Millet and Drew (1963) concluded that the paleosol had been exhumed for a very long time. Weathering ratios and texture were different, however, indicating again the usefulness of mineralogical techniques. The leaching of bases could possibly be used as a relative measure of length of exhumation.

Ruhe and Daniels (1958) described a Late Sangamon paleosol formed in a pedi-sediment layer over Kansan till in Iowa. They gave profile descriptions for the soil where it is buried beneath 310 cm of loess and where exhumed a mere 244 m away. Morphological variation between the two profiles is very slight. No chemical or mineralogical data were presented. A similar situation is described in detail by Prill and Riecken (1958), where a Late Sangamon paleosol formed in a pedi-sediment is commonly exhumed on hillsides in Iowa. The exhumed paleosol is apparently truncated, as indicated by a thinner solum (107 vs 183 cm) and A horizon (10 vs 43 cm) than its buried counterpart. The exhumed soil is finer textured, presumably due to greater length of weathering. Color hue, value and chromas vary substantially between profiles. More severely truncated exhumed soils were described for the study area, and therefore this study demonstrates that truncation is an important pedogenic change which can accompany exhumation as well as occur prior to burial. Bushue et al. (1974) found exhumed paleosols on some valley sides in Illinois to be 5-10% higher in clay content, than when they occurred deeply buried within drainage divides, due to greater length of weathering of the exhumed paleosols.

In anticipation of the chapters to follow, there is little doubt about the potential or need for research in the fields of post-burial and post-exhumation alteration of paleosols. Paton (1978, p. 106-107) perhaps summed it up best:

> The formation of soil material is the result of a complex of processes operating within boundaries determined by an interacting set of factors, the processes and boundaries being subject to variation during time...Precise statements about past changes in the factors of soil formation necessarily depend upon future research.

In the body of this thesis, these problems will be discussed and solutions advanced. Slowly, as well as rapidly, changing soil properties will be examined in order to arrive at a complete understanding of the processes and rates of pedogenesis subsequent to both burial and exhumation.

PROBLEM STATEMENT AND HYPOTHESES

By definition, paleosols have formed on landscapes of the geologic past. Therefore, they are important keys to paleoenvironments - their nature, duration, and other characteristics. The quest for answers about the past permeates the disciplines of archeology, anthropology, history, geology, and geography. But whereas an archeologist might use potsherds and bones to unlock the mysteries of the past, a paleogeographer often uses pollen and paleosols to formulate theories about previous time periods by reconstructing vegetation and soil patterns.

Paleogeography has certain innate obstacles which must be overcome. Pollen grains deteriorate at various rates, depending on grain characteristics and sediment type, rendering reconstructions of older and older climates increasingly more difficult (Hall 1978). Likewise, paleosols are altered by surface and subsurface diagenic and pedogenic processes, making their interpretation difficult. This thesis, and more specifically this chapter, deals with these changes in paleosols in a theoretical manner, and then poses several hypotheses which will be tested and evaluated.

That soils change subsequent to burial and, later, to exhumation is not questioned (See Ruellan 1971, Simonson 1954, Nikiforoff 1943, Ruhe 1956, 1969, Yaalon 1971, Gerasimov 1971, Valentine and Dalrymple 1976, Hallberg _et al_. 1978, Mausbach _et al_. 1982). Not all soil components are subject to post-burial changes, however, and rates of change vary with the soil constituent examined (Yaalon 1971). Post-burial alteration thus becomes a complex assemblage of processes and reactions to a new, subsurface environment. These processes vary spatially as functions of burial depth (vertical) and position on the landscape (horizontal). In order to fully understand such processes, a model will be explored wherein soils on a paleolandscape are buried by loess, and later exhumed.

Processes Occurring During Soil Burial. The model begins with a well-developed soil on an undulating landscape. The soil varies spatially on the landscape, but the variation is only moderate. For easy comparison with the present study, it is assumed that the soil is an Argiudoll on the uplands and a Haplaquoll in the swales.

Assume that this soil becomes slowly buried by loess. Loess deposition is typically quite slow, reported in rates of inches per century (Ruhe *et al*. 1971). Many times the A horizon of paleosols is eroded before or during episodes of loess deposition. Two possible explanations for deflation erosion of paleosols have been advanced. The first postulated that the loose granular structure of the topmost horizons makes the soil most prone to erosion. Stevenson (1969), however, explained lack of complete buried sola based on the assumption of dry conditions preceding loess deposition, which lowered the organic content of the surface horizons and made the soil more susceptible to deflation. In this model, no truncation is postulated.

Gradual loess additions are trapped by growing vegetation and accumulate at the top of the A horizon, thickening the solum. For example, conclusive evidence of very slow Peorian loess additions upon Loveland Soils in the study area is revealed by molluscan faunal assemblages (Leonard 1952). Changes in soil morphology occur as the

A horizon increases in thickness and becomes siltier in texture. Blocky structure, typical of B horizons, may become expressed on upper horizon platy (E) or granular (A1) structures. Evidence of this process may remain as dark stains when blocky ped surfaces cut across platy or granular structures (Frye et al. 1974). Because loess is often calcareous, surface pH rises. Incorporation of organic matter into the upward-building profile increases the total organic content of the solum. Deposits of slowly accumulated loess become more intensively weathered than later, more rapid, accumulations.

Now, let us assume that continued additions of loess bury the paleosol to depths of one to two meters. The effects of surface soil-forming processes on the buried soil diminish. Overall, normal chemical and biological activity within the paleosol slows. Wetting fronts from precipitation events bring carbonates and other easily-soluble weathering produces into the paleosol from overlying sediments, enriching it in Ca^{++}, Mg^{++}, Fe^{+++}, Na^{+}, and other ions. Pedoturbation slows as burial depth increases. Illuvial humus additions cannot keep pace with oxidation within the buried profile and the paleosol rapidly loses organic matter. Hydrolysis, oxidation/reduction, clay expansion and contraction, and physical weathering are slowed. The buried environment is a wetter one than that of the surface (in most climates). This change may act to promote weathering immediately above (Al-Barrack 1980) or within the paleosol (Ruhe 1969) and may lead to mass movements within or on top of the buried soil. Lastly, ped bulk density increases due to the weight of overlying strata.

It is only with very deep burial that isolation from surface soil-forming processes is approached, and even then, may not be achieved. The soil may be saturated beneath the water table, or descending vadose water may bring solutions from above into the paleosol. Organic matter becomes nearly completely oxidized. For all practical purposes, however, depths of loess five m. or more effectively isolate the paleosol from surface weathering and pedogenic processes. The paleosol is then called "dead", according to the classification of Gerasimov (1971). The paleosol remains in this state until erosion again brings it into the influence of some surface soil-forming processes, which may take thousands of years, or may "never" happen. If this happens, however, the soil systematically undergoes the processes outlined above, until complete exhumation occurs. The solum then becomes completely under the influence of new soil-forming processes, like or unlike those under which it originally formed.

Soil Burial and Isolation. At this point, it is appropriate to define "soil burial" and "soil isolation" as I will use them in this study. Most mature surface soils have some type of A-B-C horizon sequence. A subsequent deposit of alluvium, colluvium, loess, etc. buries this soil, at least temporarily. Technically, the soil has now an A-Ab-Bb-Cb sequence until the burying material is incorporated into the solum through various homogenization processes. The soil is not permanently buried because the overlying sediment is not thick enough to prevent soil-forming processes from "linking" the buried soil to the surface.

Other soils, more deeply buried (e.g., 100 cm) cannot incorporate the overlying sediment into the profile. These soils are buried and are characterized by the center example in Fig. 2. However, because certain surface pedogenic processes still function within the buried solum, it is not isolated, as in the most deeply buried soil in Fig. 2. To be isolated, a buried soil must be below the normal limit of all common vectors of surface soil-forming processes. Diagenetic processes, such as compaction, compression, and shear, are not included in the definition. Deep subsurface contamination from ground water is also excluded.

The depth to which vectors of pedogenic processes extend into the soil depends on local factors, the least important of which is time.

Paleosol properties can be expected to systematically decrease or increase with burial depth because of differing intensities of pedogenic properties with depth. Examples might be the linear decrease in weathering ratios with depth (Line A, Fig. 3), or the curvilinear increase in pH (Curve B, Fig. 3). At some depth, however, where isolation is achieved, changes in soil properties will theoretically cease. The "depth of isolation" (Fig. 3) can be defined as that depth at which variation in paleosolic properties becomes effectively zero. Put mathematically:

$$P = f(\text{Burial Depth}) \qquad \text{(Eq. 1)}$$

and:

$$\frac{dP}{dt} = 0 \text{ at or below the "Depth of Isolation"} \qquad \text{(Eq. 2)}$$

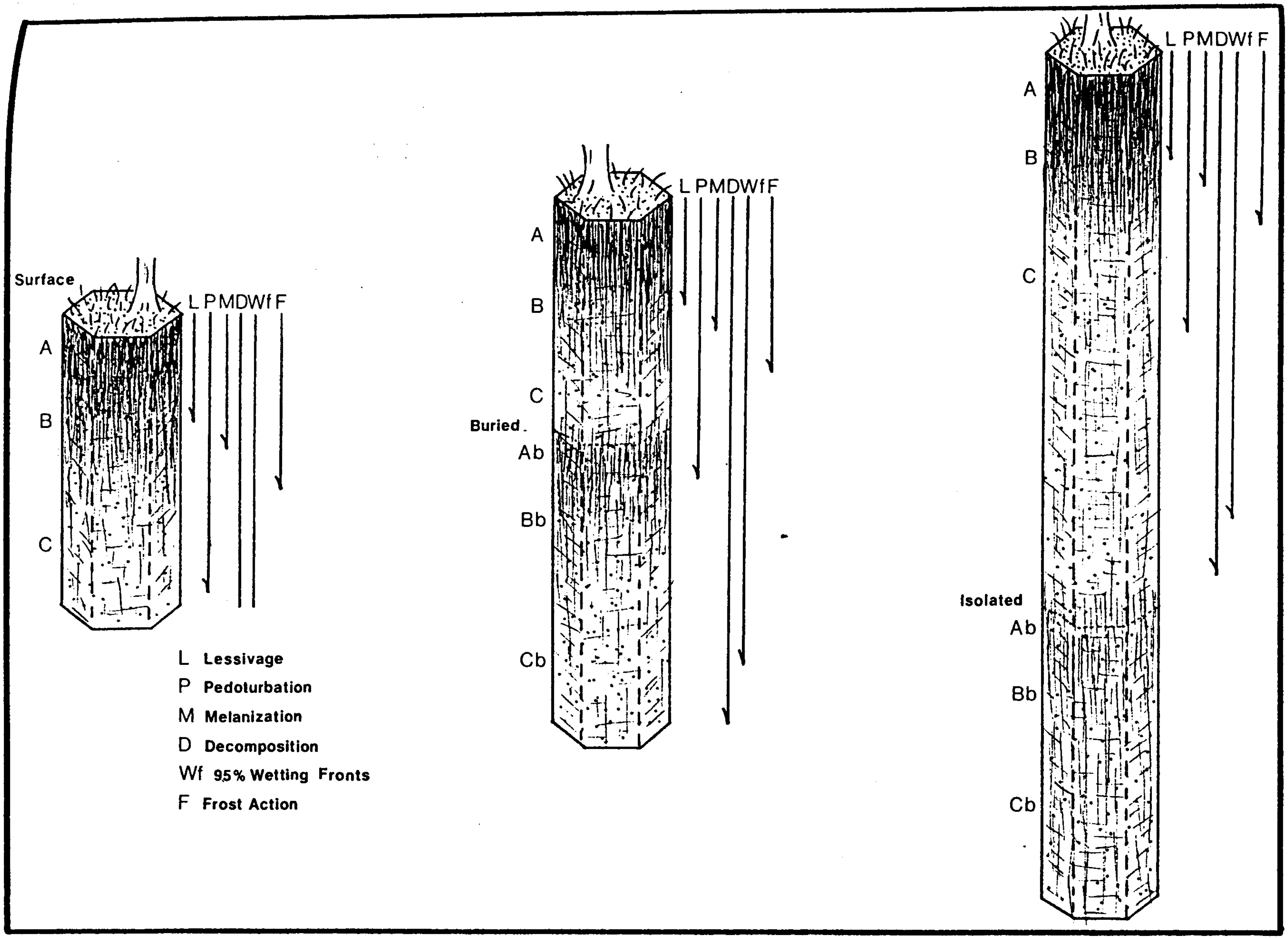

Figure 2. Examples of Pedons containing surface, buried, and isolated soils. Hypothetical depth of pedogenic process influence is shown.

where P is any soil property. The depth at which the derivative of Eq. 1 becomes zero is the depth of isolation (Fig. 3).

It is implied that Eq. 2 will be satisfied at a different depth for each soil property (Fig. 2). Does this imply that a depth at which a paleosol is isolated from surface soil-forming processes cannot be established? I suggest, as do Valentine and Dalrymple (1976, p. 217), not. Rather, paleosol properties become isolated at unique depths. The depth of isolation may be 120 cm for pH, 330 cm for organic matter, and 650 cm for weathering of the B horizon. One can discuss isolation of paleosols with regard to 1) individual soil properties or 2) soil processes.

Equations 1 and 2 assume that 1) the data set is from a single polypedon with similar soil properties throughout, and 2) contamination from subsurface water is ineffective in substantially altering the soil.

Upon exhumation soils go through the same pedogenic reactions as do initial undifferentiated parent materials. Soils gain organic matter rapidly at first, with rates of increase slowing with time (Jenny 1980). Soil pH equilibrates with the surface environment via leaching and ion transfer. Weathering proceeds at normal rates to common terminal products.

It is implicit in the above model that soil constitutents vary considerably in their resistance to change after burial or exhumation. Numerous researchers have noted this problem, but only Ruhe (1956, 1969), Yaalon (1971), Gerasimov (1971), Valentine and Dalrymple (1976), Hallberg *et al*. (1978), and Mausbach *et al*. (1982) discuss the problem at any length. A general consensus of the above articles reveals increasing resistance to post-burial alteration on the order of:

$$\text{pH} < \begin{array}{c}\text{Organic}\\ \text{Matter}\end{array} < \begin{array}{c}\text{Base Saturation}\\ \text{Carbonate Content}\\ \text{Soil Structure}\end{array} < R_2O_3 < \text{Texture} < \text{Mineralogy}$$

based on this author's interpretations and summations. Individual pedons will, of course, vary in the way they alter during and after burial. The reader is referred to Gerasimov (1971) for an excellent discussion of post-burial alterations.

In the study area, exhumed paleosols can be laterally traced to shallowly and deeply buried paleosols. Burial and exhumation act only to complicate the interpedon morphological and chemical variation of the paleolandscape. Therefore, discussion of post-burial and post-exhumation alteration must take into account the original morphometry of the landscape and the processes acting on the soils of that landscape since burial. It becomes a very deductive study.

In this thesis buried and exhumed paleosols are examined, and quantitative theories and formulae relating depth of burial to pedogenic characteristics of the paleosol are formulated. The influence of a buried pedostratigraphic layer on characteristics and genesis of the surface soils in the study area is investigated also. At the very least, it will advance questions about processes, not completely understood, acting on buried soils.

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tion from the above pattern may imply changes in the paleosol subsequent to burial and exhumation.

Lastly, parent material uniformity is assumed for both surface soils and paleosols. Though this condition is lacking in the study area, it is believed strong initial pedogenesis in the Late Pleistocene, followed by ongoing pedometamorphic processes during the Holocene, will partially negate the effect of parent material heterogeneity.

Given the model and the three assumptions outlined above, one can assume increased weathering of paleosols as burial depth decreases. Weathering of silts to clays is also favored. Therefore:

I. Maximal and total clay contents of the paleosol will increase as burial depth decreases.

and

II. Coarse silt + fine sand weathering ratios (Quartz/Feldspar) within the paleosol will increase with decreasing burial depth.

Soil pH and organic matter contents rapidly reach equilibrium with soil environments. Organic matter both oxidizes and accumulates rapidly, and pH changes as a function of soil solution concentrations. Therefore:

III. Paleosol pH's will not vary substantially from values at the same depth in loessial soils without a paleosol.

and

IV. Organic matter content within a buried paleosol will be equivalent to the organic matter content at similar depths in soils developed in deep loess only.

These four hypotheses will be tested for a paleosol in northeastern Kansas and may or may not apply to paleosols elsewhere. The investigation seeks only to begin examining some of the soil-forming processes which operate at various depths in the pedosphere through a study of paleosol alteration.

STUDY AREA DESCRIPTION

Location. The study area encompasses approximately ten square miles in the northeastern part of Brown County, which is in northeastern Kansas (Fig. 4). This area is shown on the Highland NW, Kansas 7.5 minute quadrangle. Specific sampling site locations are provided later.

Physiography. Wisconsinan and Illinoian loesses together approaching 5 m in depth cap glacial till preserved on interfluves in the study area. Major streams have incised through the loess into the till below. Weathering and erosion of the till and loess have developed a rounded terrain (Plate 1). The modern landscape consists of rolling hills having a mean relief of 30-45 m. The loess is believed to have its main source area in the Missouri River floodplain 15-25 km to the east and north, where exposures on the adjacent bluffs reveal loess depths up to 20 m (Frye and Leonard 1949, Hanna and Bidwell 1955, Caspall 1970). Lesser local sources may have included tributaries to the Missouri such as the North Fork of the Big Nemaha, and the Blue River (Brown 1977).

Slopes are variable, ranging from gentle slopes in the alluvial valleys to the steeper pediment backslopes (Ruhe 1960), which reach angles of 10-18%. Potential for soil erosion is great. Locations where several meters (up to 5 m in one location) of the friable loess soils have been eroded since cultivation began are not uncommon. Deposits believed to be post-settlement alluvium, observed in valley floodplains, often exceed 35 cm in depth. Recently, extensive terracing of tilled fields has been performed, proportionately lessening soil erosion.

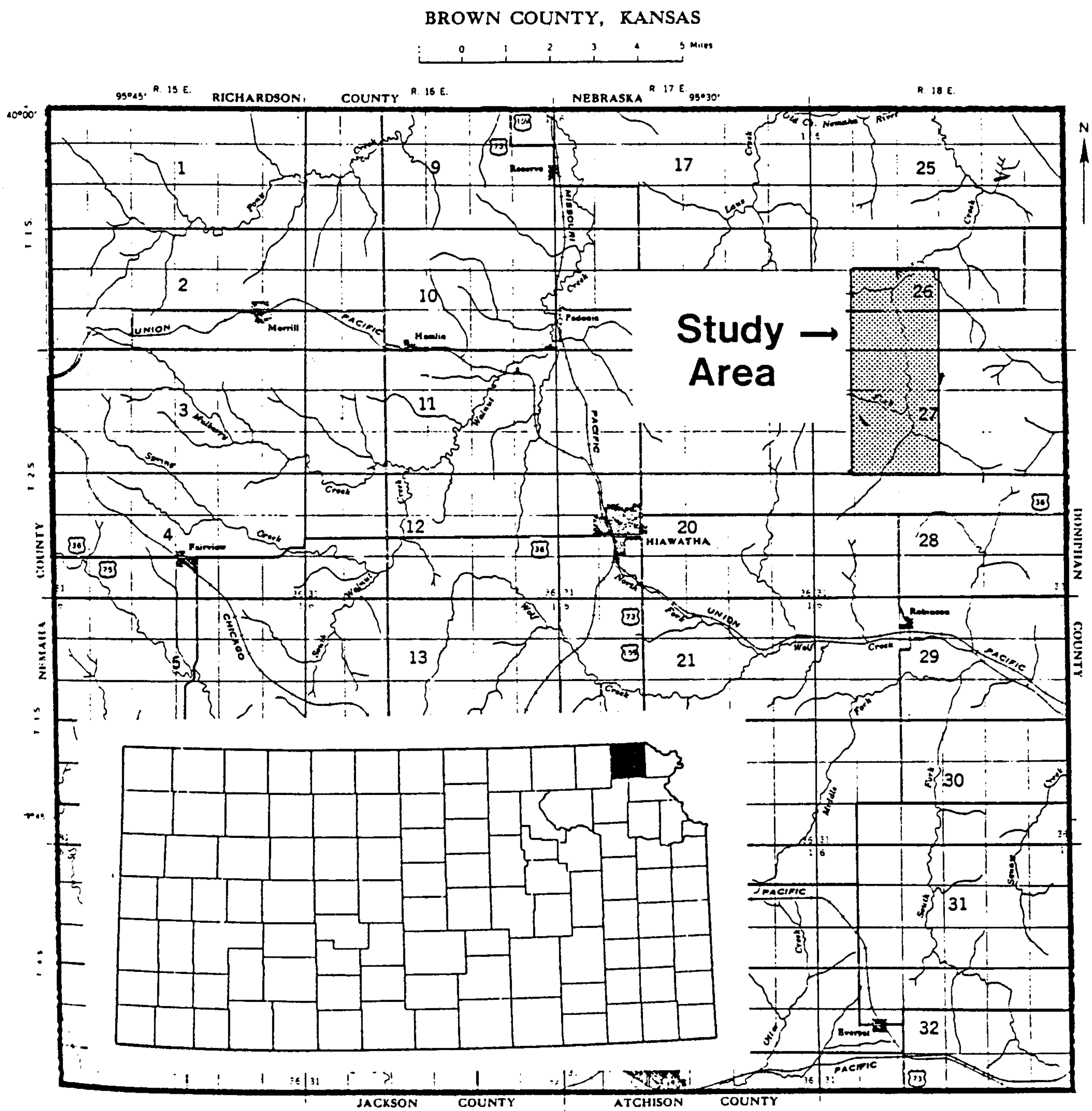


Figure 4. Study area location map. Brown county, Kansas.

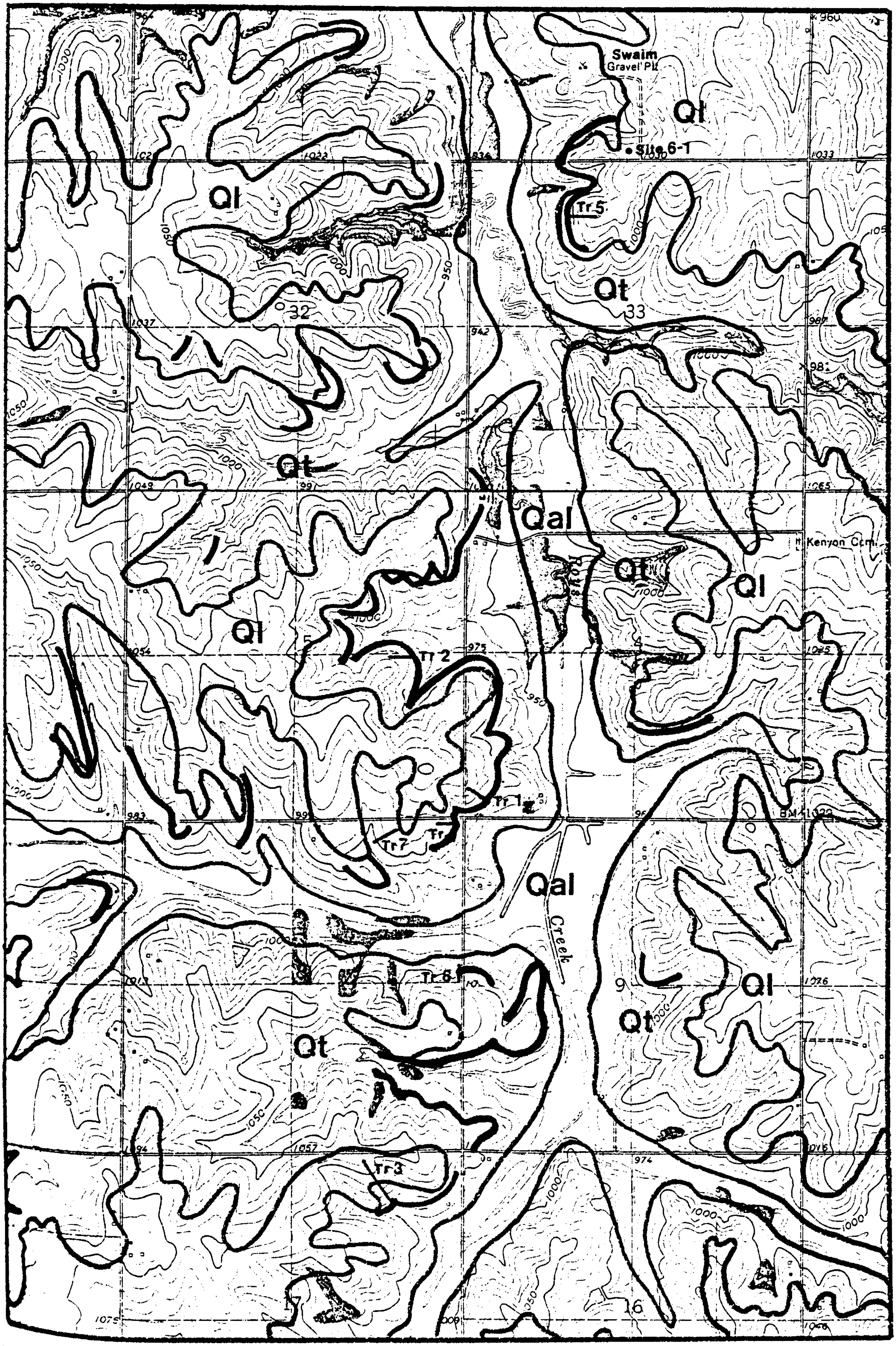
Plate 1. Landscape in the study area, northeastern Brown county, Kansas. Reddish band on hillside is exhumed paleosol.

Soils. Soils in the study area have developed in Pleistocene loesses and tills, as well as Holocene alluvium. Upland soils are predominantly the Marshall (fine-silty, montmorillonitic, mesic Typic Hapludoll) and Sharpsburg (fine, montmorillonitic, mesic, Typic Argiudoll) series. These soils have formed in thick Peorian and Bignell loesses of late Wisconsinan age. (Information in this section comes predominantly from Eickelberry and Templin 1960).

Within the study area a reddish soil band can be observed on hillslopes, often appearing about halfway up the slope from the valley bottoms (Plate 1) and consistently at or about the 1000-foot contour line (Fig. 5). This reddish band is believed to be an exhumed Sangamon or Late Sangamon paleosol, first reported in this area by Frye and Leonard (1952) and mapped as Morrill loam (fine-loamy, mixed, mesic Typic Argiudoll) by Eickelberry and Templin (1960). It is a strongly-developed paleosol according to the criteria of Birkeland (1967, 1974). This soil band is very prominent on nose slopes but is often covered by colluvial deposits within the draws (Fig. 6). Roadcuts and cores taken during the course of this research demonstrate that the red band is a pedo-stratigraphic unit which can be traced beneath younger loess as one approaches the uplands but is absent downslope owing to erosion.

Outcrops of Morrill loam are restricted to northeastern Brown county where geologically recent erosion by Roys Creek and its tributaries (Fig. 5) has stripped the loess cover from the paleosol. This is confirmed by the county geologic map (Bayne and Schoewe 1967). The map shows very well the till/loess interface which coincides closely with the paleosol outcrop locations (Fig. 5). Differences in exhumed paleosol locations from those of boundary positions on the geologic map result from the original small scale mapping errors of the latter.

West and south in Brown county, longer periods of erosion in thinner loess have removed the loess and the paleosol, exposing unweathered Kansan till below. A similar situation exists in the study area downslope from the paleosol outcrops (Fig. 6). The soils


Swaim
Gravel Pit
Ql
Site 6-1
Tr 5
Qt
32
33
Qt
Qal
Kenyon Cem.
Qt
Ql
Ql
Tr 2
Tr 1
Tr 7
Tr
Qal
Creek
Tr 6
Ql
9
Qt
Qt
Tr 3
16

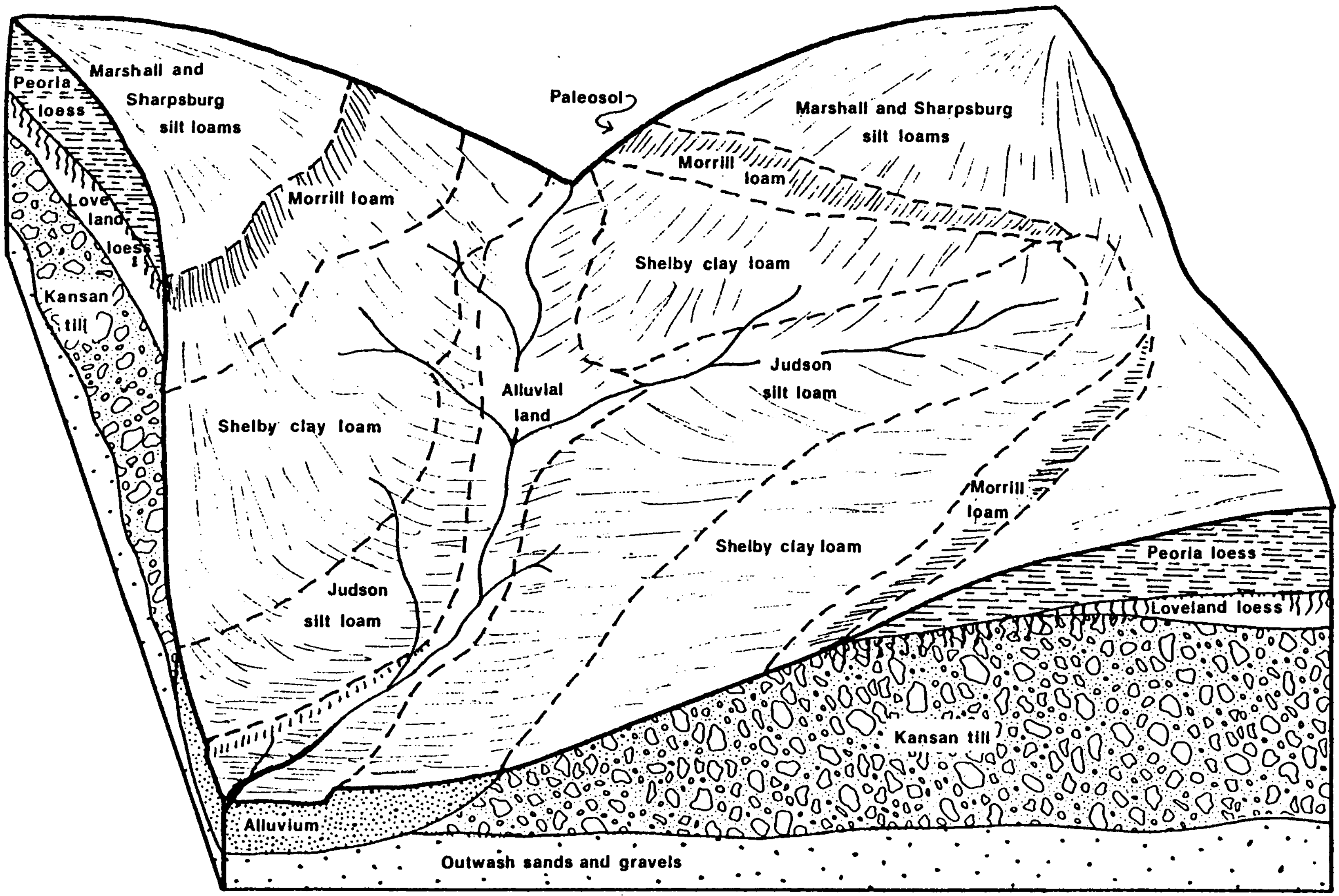


Figure 6. Near surface stratigraphy and soil associations in the study area. Locations of soil boundaries from Eickelberry & Templin (1960); subsurface information provided by the author.

developed in till without an exhumed paleosol are mapped within the Shelby series (fine-loamy, mixed, mesic Typic Argiudoll). The Morrill soil is a paleosol formed in an exhumed till (or Loveland loess) while Shelby soils are developed in relatively freshly exposed, unweathered till. At the time of mapping, however, Eickelberry and Templin (1960) did not recognize the relict properties of the Morrill.

In the valleys Judson soils (fine-silty, mixed, mesic Cumulic Hapludoll) have formed in a mixture of colluvial materials and overbank flood deposits. Alluvial soils (Entisols) are common in areas directly adjoining streams (Fig. 6).

Pleistocene History and Stratigraphy. The area was last glaciated during the classical Kansas glacial stage (Flint 1971). Present understanding is unclear, but this till is believed to date back to at least 500,000 BP (Boellstorff 1978), or to fall between 170,000-610,000 BP (Richmond no date). According to Bayne and Schoewe (1967), Kansan till underlies much of the landscape. Its presence of till in numerous exposures both within and near the study area (Caspall 1970, Frye 1951, Frye and Leonard 1949, 1952), and in soil cores taken during the course of the research, substantiate this supposition. Personal observation of stratigraphy and discussion of well logs with local farmers yielded strong evidence for a sandy/gravelly outwash deposit underlying most, if not all, of the Kansan till and subsequent loesses in the study area. This deposit is probably the Grand Island formation, outwash sands of early Kansan age (Bayne and Schoewe 1967), though definitive correlation is beyond the scope of the present study.

To the author's knowledge the pleistocene stratigraphic sequence in the study area begins with the outwash member covered with an uneven thickness of classical Kansan till (Bayne and Schoewe 1967). Loveland loess of Illinoian age was deposited on the Kansas surface. It is likely that a period of erosion preceded this depositional episode because a Yarmouth soil between the Kansan till and the Loveland loess above was never observed. In nearby southeastern Nebraska, classical Kansan till overlain by Loveland loess also lacks any evidence of a Yarmouth soil (Al-Janabi and Drew 1967). The closest suspected Yarmouth soil is in Doniphan County, Kansas. At many locations in the study area Loveland loess is conspicuously absent and Woodfordian age Peorian loess lies unconformably above Kansan till (Fig. 6).

Two types of stratigraphic sequences exist in the study area (Fig. 6). In the first, loess believed to be of Peorian age (Bayne and Schoewe 1967, Caspall 1970) directly caps Kansan till. The paleosol is then composite; the upper horizons formed in a 0-75 cm mixed zone of loess/till caused by pedoturbation (cryopedoturbation, insect burrows, windthrow, etc.) while the remaining lower horizons formed in a pedi-surface (Ruhe 1956) developed in till. Occasionally, where the till is thin, the profile may extend into the sandy outwash beneath. The mixed zone was believed formed by thickening of the A horizon at the onset of loess deposition. As loess was slowly added to the soil surface and trapped by growing vegetation, the top of the A horizon migrated upward and the solum thickened. Eventually, it is believed that loess deposition outpaced the ability of the soil to incorporate it into the profile, and the soil became buried.

Figure 6 also displays the second type of Quaternary stratigraphic succession in the study area. Here the paleosol is developed in Loveland loess only. No appreciable mixed zone is observed between the Loveland and Peorian loesses, based on textural data. Loveland loess has greater clay-free silt percentages than Peorian (84-95% vs. 75-82%). Kansan till underlies the Loveland loess, based on core data, but is not a parent material of the paleosol. At depth, outwash is probably present. This second type of stratigraphic succession is much less common than the first. A more thorough discussion of the stratigraphy in northeastern Brown county is given in Appendix I.

The study area lies at a critical point in the Pleistocene/Holocene landscape of northeastern Kansas. As one leaves the study area and approaches the Missouri River, the paleosol is buried increasingly deeper beneath thickening Peoria loess, until only 3-5 km east of the study area the paleosol is apparently completely buried. Exhumed paleosol exposures appear at progressively lower positions on the hillslopes as one approaches the presumed loess source areas. Moving westward from the river, the Peorian loess thins rapidly; paleosol exposures appear more often in upland positions. Very quickly the paleosol becomes relict, and often is absent due to geologic erosion. The study area lies at the midpoint between the deep loess area to the east and north, and the think loess/till landscape to the west and south.

Related Investigations. The study area offers an excellent opportunity to study well-preserved Pleistocene soils which have not been described previously. Hanna and Bidwell (1955) studied the upland loess soils of the area. Frye and Leonard (1949) described a Pleistocene exposure near the Missouri River and they later (1952) mentioned the presence of the Sangamon soil throughout northeastern Kansas. Millet and Drew (1963) and Al-Barrack (1980) studied soils in southeastern Nebraska which may be stratigraphically related to the paleosols of the present study. Caspall (1970) examined evidence for the presence of a Brady paleosol (Two Creekan age) between early and late Wisconsinan loesses, in numerous exposures near the Missouri River, while Frye and Leonard (1955) attempted a correlation of post-Sangamonian aged deposits in the Midwest based on Pleistocene tills, soils and loesses of (particularly) Kansas and Nebraska. Tien (1968) investigated the clay mineralogy of numerous Pleistocene deposits in Doniphan county. The loess deposits of northeastern Kansas have been studied with regard to zonation (Leonard 1951) and petrography (Swineford and Frye 1951). Studies of the Sangamon soil in Brown county are lacking in the literature.

METHODS

Field Methods and Sampling. Seven hillsides, chosen as representative of the landscape based on reconnaisance surveys, were sampled using the transect method. Transects, beginning at the exhumed paleosol, in Morrill soils, progressed upslope normal to the contour (Fig. 5). Uphill sampling sites were in soils collectively mapped as Marshall/Sharpsburg units. In order to minimize the influence of aspect, four of the seven transects were run in a cardinal direction, and three were taken along intermediate directions. Transects 1, 3, and 5 were on terraced fields.

Original clay content of the loess parent material in the study area will influence the texture within soils developed in it. Textural variations in loess deposits commonly occur as logarithmic functions of distance from the source area (Smith 1942, Hanna and Bidwell 1955, Ruhe 1969, Caspall 1970, Hall 1973, Brown 1977). Therefore, in order to minimize the inter-pedon variability in loessal texture, the sample area was delineated in a direction parallel to the Missouri River Floodplain, which is the believed source of most of the loess.

Along each transect replicate soil cores were taken with a Giddings hydraulic probe at sample sites. (Core size was 2" dia. and hence two cores were required at each sample site to insure adequate soil for laboratory analyses.) The exhumed paleosol was cored in most cases and locations of sampling points upslope were randomly determined. Paleosol burial depth increased upslope as the hillcrests were approached. Transect lines were discontinued when the depth-to-paleosol became greater than three to four meters. Core depth was sometimes limited when a substantial clay accumulation in the paleosol prevented further penetration by the coring machine. More often, the core was terminated when it appeared, based on texture and color, that the core had gone through the buried solum.

In addition to the seven linear transects, one point sample (replicated) core was taken near the crest of a hill, with the intention of continuing it as the eighth transect downslope. Mechanical problems prevented further coring, however, and the single core was used alone. Also, an exposure in the face of a gravel pit, revealing nearly four meters of loess over a paleosol developed in till and outwash gravels, was sampled. This pit will hereafter be called the Swaim Pit, after the owner, Mr. Harry Swaim. In all, thirty replicated core samples were taken for analysis.

Laboratory Methods. Laboratory analyses were conducted on 10 or 15 cm incremental core samples, taken at 0-10 cm, 15-30 cm, 45-55 cm, 70-80 cm, 95-105 cm, 120-130 cm and so on to the base of the core, resulting in four or five samples/meter. Compaction of the soil within the core was routinely less than 10% and was adjusted for appropriately.

Color proved to be a reliable indicator of the presence of a paleosol, in this study, because of its well-developed characteristics. Depth to the top of the paleosol, based on color, was recorded both in the field using moist soil and in the laboratory with dry soil. Lithologic discontinuities were documented based on fresh core appearance and, later, from cores sectioned in the laboratory.

Incremental samples were air dried and ground in a mechanical soil grinder. With the exception of the gravel pit samples, gravels were discarded. Particle size analysis was performed following the method of Bouyoucos (1936), as modified by Day (1965). Percentages of sand (2 mm-50 μ), coarse silt (50 μ-20 μ), medium plus fine silt (20 μ-2 μ) and clay (<2μ) were determined. Sample pH was measured using a 1:1 soil-water saturated paste, after three hours of equilibration, using a Croning Model 7 glass electrode pH meter. Organic matter content was determined using a modified Walkley-Black method (Black 1965). Soil laboratory data are compiled in Appendix II.

Selected samples were analyzed for clay mineralogy using a General Electric XRD-5 x-ray diffraction unit. Five sample types were selected: 1) clay from the Bt max of a representative surface soil formed in loess, 2) clay from the loess below the solum of the surface soil but above the paleosol, and clay from the Btb max of paleosols formed in 3) till, 4) loess, and 5) outwash sands. Clay was isolated using centrifugation, following dispersion with sodium hexametaphosphate and mechanical mixing in a fountain-type mixing machine. Clays were fractionated into fine (<0.2 μ) and coarse (0.2 μ-2 μ) components using centrifugation (Whittig 1965). All samples were pretreated with hydrogen peroxide before x-raying to eliminate organics. Preferred clay orientation was accomplished by allowing the clay/water suspension to air-dry on petrographic slides. Samples were x-rayed from 4-14 degrees 2 θ using 40 KVP, 15 milliamps of copper K-alpha radiation. Untreated, glycolated and heated (550^{o} C for 1 hour) samples were analyzed. Interpretation of peaks is outlined later in this report.

The coarse silt/fine sand fraction (44-62 μ) from the Btb max of all paleosols, buried and exhumed, was isolated and tested for relative quartz/feldspar content. Wet sieving segregated the grains into the appropriate size fraction. Grains were treated with hydrogen peroxide to remove organic coatings. A modified citrate/dithionite method successfully removed iron oxide coatings (Agüilera and Jackson 1953). Quartz/feldspar ratios were determined from x-ray diffraction peak heights. The height of the peaks at 20.9° 2 θ (4.2 A) and 28.0° 2 θ (3.2 A), for quartz and feldspar, respectively, were compared to arrive at a ratio. Five slides were x-rayed for each sample and the mean peak heights used.

RESULTS AND DISCUSSION

Predictably, both the surface soil and a paleosol buried below will exhibit chemical and physical changes as the vertical distance between the two varies. This section explores such changes in texture, pH and organic matter content, clay mineralogy, and weathering ratios.

Texture. In extreme northeastern Kansas, upland soils have developed in oxidized and leached Peorian loess and Kansan till. Based on particle size data from soil cores taken in the field, Peorian loess in the study area has a mean texture of 9.5% sand, 61.9% silt, and 28.6% clay: it is a silty clay loam. Peorian loess of the study area has more sand than other data on Peorian loess from the Midwest suggest (Wascher *et al*. 1947, Ulrich 1949, Hanna and Bidwell 1955, Hogan and Beatty 1963), and in most cases, less clay (Wascher *et al*. 1947, Ulrich 1949, Swineford and Frye 1951, Hogan and Beatty 1963) (Table 2). Pedogenesis and weathering in the upper 30 cm of the surface soil has resulted in an increase in mean clay content to 34.0%. Wind and water erosion have increased the surface soil sand percentage to 11.3%, both processes resulting in a decreased silt percentage of 54.7%. Hanna and Bidwell (1955) observed a clay increase of 8-10% in progressing from deep loess to the surface of upland loess soils at the study site of the present investigation. This study reports a mean Bt max clay content of 36.0%, compared to values of 41-45% observed by Hanna and Bidwell (1955).

Sangamon paleosols have formed in both Loveland loess and Kansan till, the latter having an extensive pedoturbative and/or

Table 2

Particle Size Data for Selected Peorian Loess Deposits in the Midwest

Percentages			Location	Source
Sand	Silt	Clay		
5.0	87.0	8.0	Brown Co., Kansas	Swineford and Frye (1951)
13.0	84.0	3.0	Doniphan Co., Kansas	Swineford and Frye (1951)
1.1	69.8	29.0	SE Nebraska	Al-Janabi and Drew (1967)
2.0	67.0	30.1	SE Nebraska	Al-Janabi and Drew (1967)
1.5	69.1	29.6	SE Nebraska	Al-Janabi and Drew (1967)
0.6-1.1	62.0-69.0	30.0-38.0	SE Nebraska	Al-Barrack (1980)
1.0	81.9	17.1	W. Tennessee	Wascher _et al._ (1947)
1.0	89.2	9.8	W. Tennessee	Wascher _et al._ (1947)
0.6-0.7	81.7-86.9	12.4-17.7	Miss.	Wascher _et al._ (1947)
1.9	92.6	5.5	Miss.	Wascher _et al._ (1947)
10.0-14.0	73.0-78.0	12.0-13.0	SW Wisconsin	Hogan and Beatty (1963)
2.0	67.0	31.0	NE Kansas	Hanna and Bidwell (1955)
4.0	71.0	25.0	NE Kansas	Hanna and Bidwell (1955)
7.7	69.1	23.2	SW Iowa	Ulrich (1949)
n.r.	n.r.	10.0-15.0	Doniphan Co., Kansas	Caspall (1970)
9.5	61.9	28.6	Brown Co., Kansas	This study

colluvial mixed zone where loess lies unconformably above the till. When developed in Loess, the paleosol has a silty clay loam texture (mean values: 13.1% sand, 52.1% silt, 34.8% clay). The paleosol in till is coarser textured (clay loam: 23.6% sand, 41.4% silt, 35.0% clay). Compared to a mean Bt max of 36.0% clay in the surface soils, paleosols in the study area are more clay-rich, with a mean Btb max of 39.7%, when all paleosols are examined. This finding agrees with the data of Simonson (1941), Frye and Leonard (1949), Ruhe (1956, 1965, 1969, 1970), Badgley (1957), Caspall (1970), Hall (1973), Ruhe et al. (1974), Bushue et al. (1974), and Ruhe and Olson (1980), who report that the Sangamon paleosol is finer-textured than contemporary surface soils.

Hypothesis I of this thesis supposed increasing clay in the paleosol as burial depth decreased, reflecting greater weathering nearer the modern soil surface. Data collected during the course of this research substantiate the hypothesis, though the correlation is low due to 1) variable parent materials, and 2) small sample size (n = 25). Maximal clay content decreases with depth as:

$$Y = 40.548 - 1.089X \tag{3}$$

where Y is the maximum clay content of the paleosols (Btb max) and X is the depth to the Btb max (Fig. 7). One data point, near the top of Fig. 7, was eliminated. Equation 3 explains 29.7% of the textural variation in the paleosol ($r^2 = 0.297$) and is significant at the .01 level. It is the first time these two variables have been compared in the literature. The data, in part, agree with the observations of Bushue et al. (1974) and Prill and Riecken (1958), where exhumed paleosols were finer-textured than buried counterparts. Further research in the subject area is strongly encouraged in order to either confirm or negate Eq. 3.

A comparison of Fig. 7 with Fig. 3 suggests that, in Brown county, five m. of loess is not adequate to isolate a paleosol with respect to clay formation and/or illuviation. Though some of the explanation

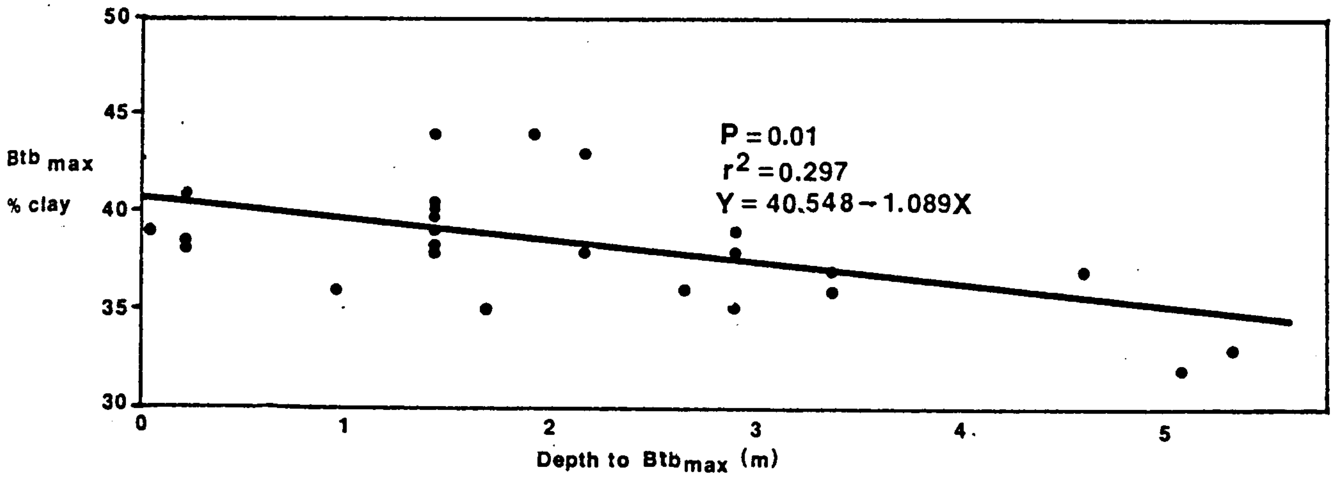


Figure 7. Relationship between maximal clay in buried paleosols and burial depth.

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of the data in Fig. 7 cannot be assigned to post-burial alteration, the pattern <u>is</u> counter to what one may believe to be the original textural variation within the paleosols. Upland paleosols (most deeply buried) here have 3-4% less clay than juxtaposed sola on sideslopes, leading one to explain the increase based on clay formation in the latter due to surface proximity. An alternative explanation might postulate increased lessivage within the sola of shallowly buried and exhumed paleosols, resulting in only larger clay peaks and not necessarily more <u>total</u> clay. Total clay in the sola of paleosols could not be calculated due to 1) the inability to clearly recognize the "top" of some paleosols, and 2) the incomplete recovery of some sola by the probing machine. The author is aware, however, that total clay rather than maximal clay would be a better dependent variable to use in Eq. 3.

A "welded soil" (Ruhe and Olson 1980) is one in which the solum of a modern surface soil merges with the solum of a buried soil because of a very thin covering of sediments over the latter. Often, these soils exhibit pronounced double clay maxima, one forming in the surface soil and one relict from the paleosol. Welded soils are commonly observed in the study area wherever the paleosol is buried at depths less than 125 cm. A thin (50-125 cm) loess blanket covering the paleosol and the subsequent lessivage in the surface soil results in the characteristic double clay maximum (Fig. 8).

Deeply buried paleosols result in "composite" profiles (Bos and Sevink 1975), where a paleosol is separated from the surface soil by nonsoil material (See Al-Barrack 1980 for examples). In the study area Peorian loess often separates the two sola. Typically, in composite profiles, double clay maxima are present yet separated by 150 cm or more of relatively unweathered loess (Fig. 9). Surface clay peaks are often observed at depths of less than 35 cm, the shallowness due not to pedogenesis but to erosion brought on by extensive cultivation. Therefore, the upper clay peak appears closer to the soil surface than can be explained by pedogenic processes alone. Occasionally, severe erosion has resulted in clay maxima at the surface; Brown (1977) observed this type of profile in the loessal soils of southeastern Nebraska.

Exhumed paleosols commonly display only one clay maximum which formed originally in the Sangamon soil and is reinforced by modern soil-forming processes (Fig. 10).

Soil pH. The stable "pH vs depth" curve for soils in eastern Kansas displays minimum values near the surface, rapidly increasing to a very slightly acidic reaction at some greater depth (See Al-Janabi

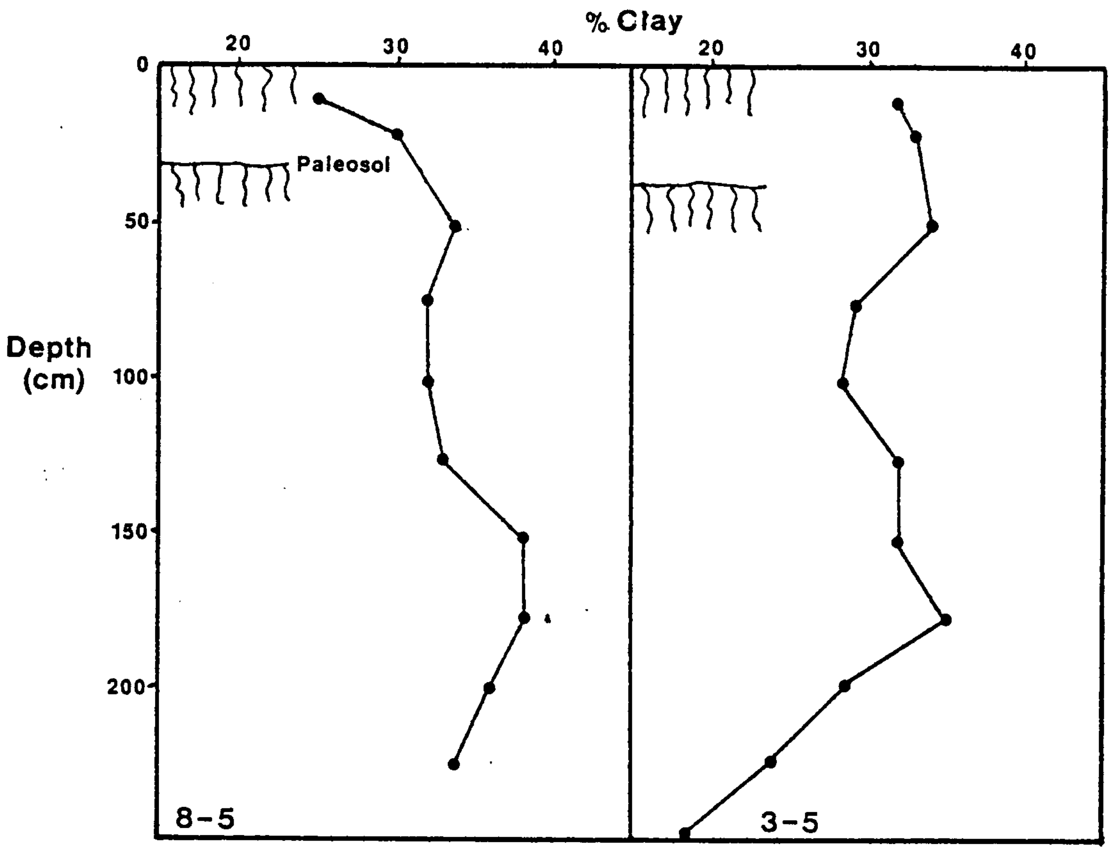

Figure 8. Welded soil profiles at sites 8-5 and 3-5.

and Drew 1967 for examples). This "depth of constant pH" varies with the parent material, topography and biota. At the study area mean surface pH in loe-s soils is 5.08. The pH increases only 0.25 units from 1.0 to 3.75 meters (Fig. 11). Where as Fig. 11 is the calculated mean value from loess and loess soils only, no significant variation is induced by the presen-e of a paleosol anywhere in the stratigraphic column. This finding is in full agreement with those of Ruhe (1956, 1965, 1974, Ruhe et al. 1974, Ruhe and Olson 1980), Al-Barrack (1980), and Yaalon (1971). That a paleosol does not significantly influence pH is conclusively demonstrated in Fig. 12, which plots depth vs. pH at the visual top of paleosols (a point where downwash of bases and carbonates might slow, proportionately increasing pH). Superimposed is the mean pH curve from

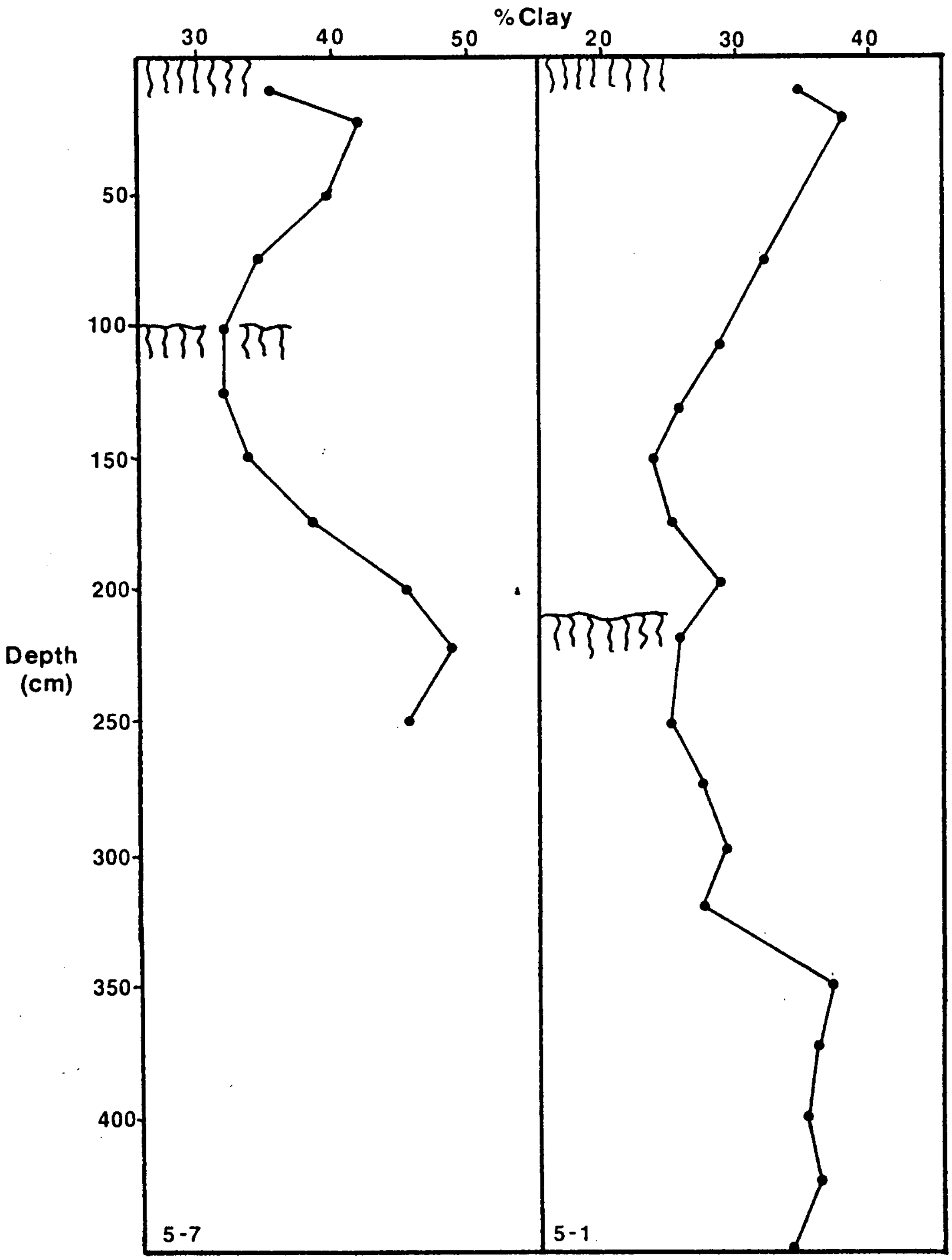

Figure 9. Composite soil profiles at sites 5-7 and 5-1.

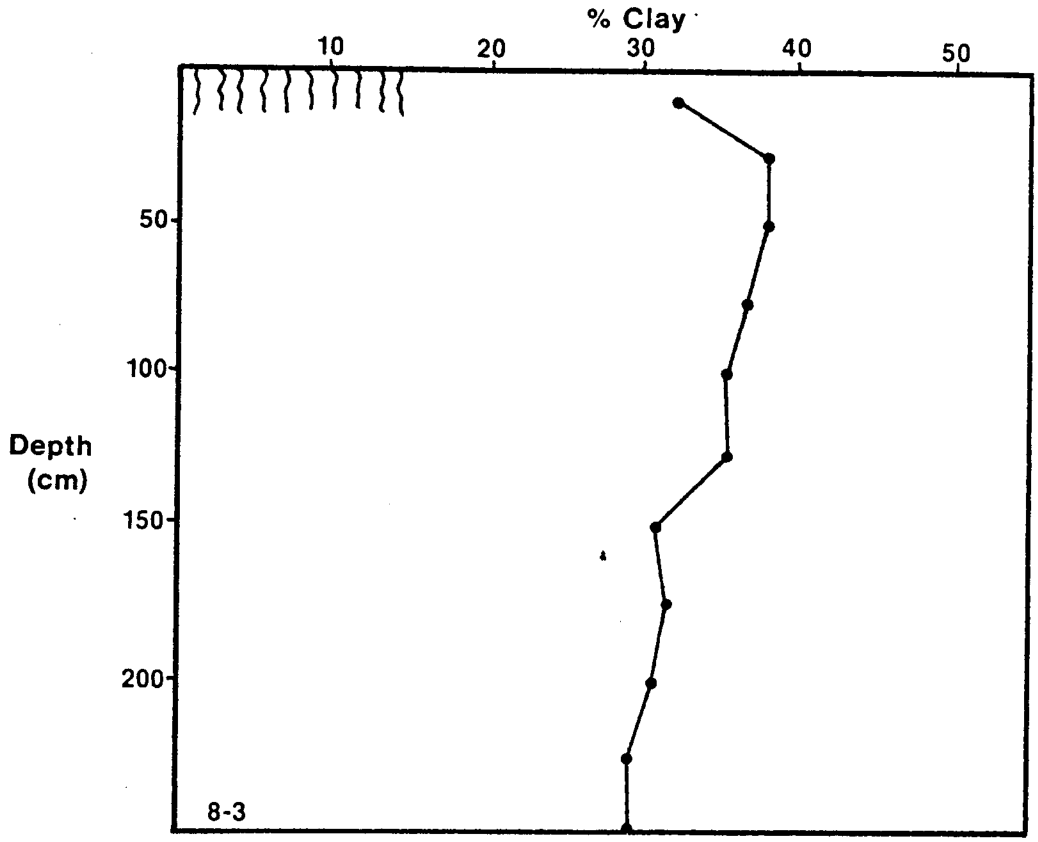

Figure 10. Single clay maximum in exhumed paleosol.

Figure 11. The pH's of the paleosols at all but the shallowest depths cluster about the mean pH curve. For paleosols in Brown county, the "depth of isolation" with regard to pH appears to be about 1.5-2.0 m (cp. Figs. 3 and 12). These findings substantiate the author's original hypothesis (III) regarding lack of influence of a paleosol on soil pH.

Exhumed paleosols have somewhat higher surface pH values than comparable depths for soils formed in loess.

Organic Matter. Organic matter both accumulates and decays quite rapidly in soils (Stevenson 1965, 1969, Valentine and Dalrymple 1976).

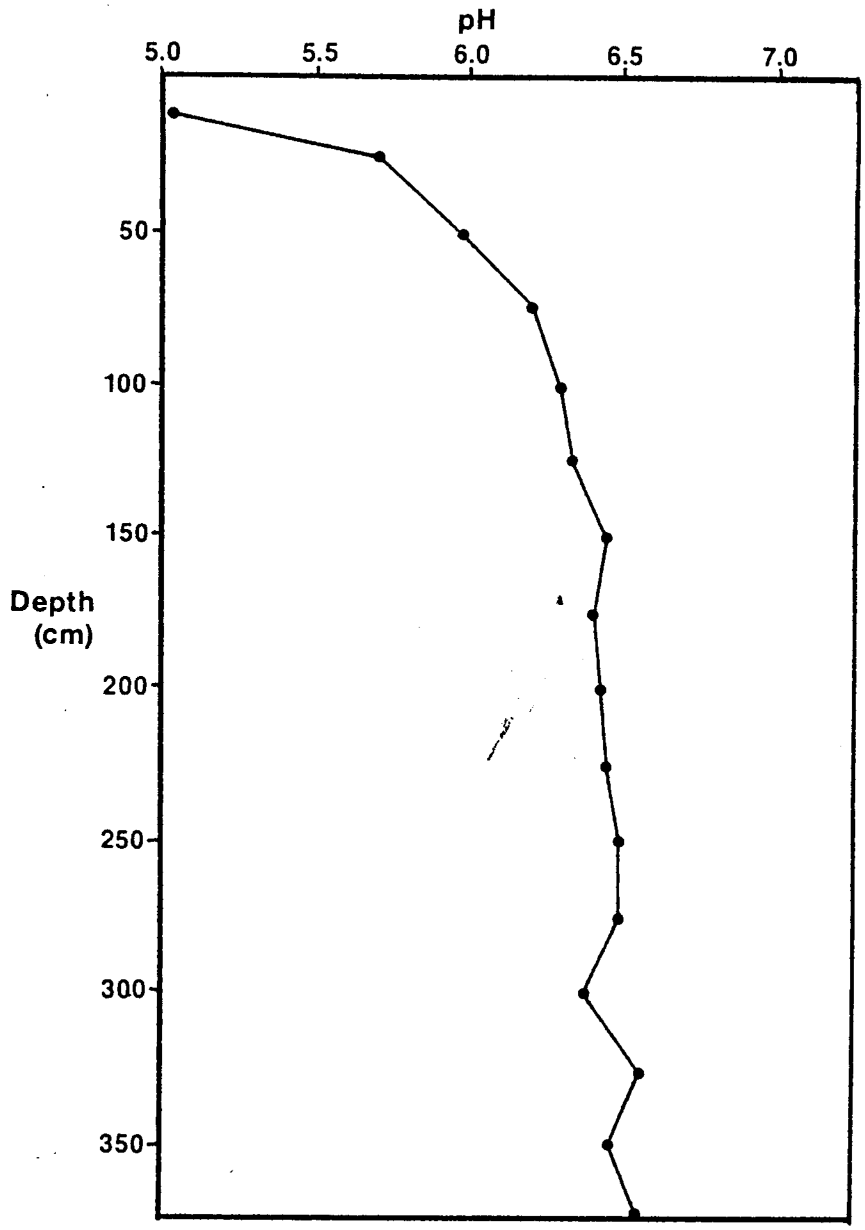

Figure 11. Mean pH values at various depths in loess and soils derived from loess in the study area. Variation at depth (> 300 cm) is due to small sample size.

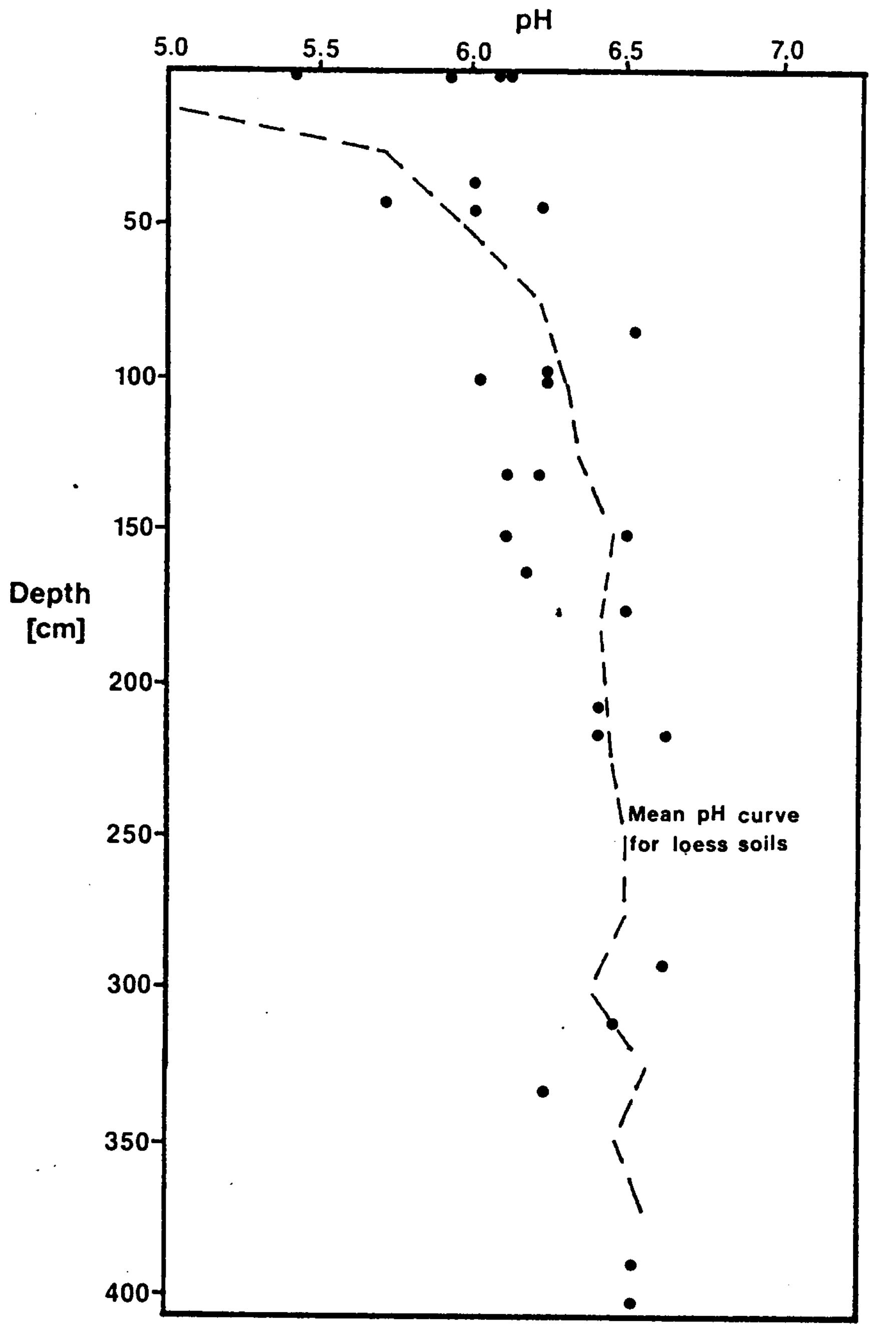

Figure 12. Spot pH values for the visual top of paleosols at various depths at sites throughout the study area. Paleosol pH closely agrees with the mean pH curve for soils formed in loess without or above a paleosol.

It is one of the most easily altered components of buried paleosol composition. Decay of organic materials after burial is initially rapid, with the fibrous organics the first to decompose (Hallberg et al. 1978). Later only the most resistant humic acids, lipids, and waxes remain (Stevenson 1969).

Mean organic matter content of modern soils (paleosol depth greater than three m.) reveals the common logarithmic decline with depth (Fig. 13). Surface organic matter averages 3.56% by weight, but rapidly declines to values around 0.38%. Exhumed paleosols display considerably less organic matter than modern soils formed in Wisconsinan loess (Fig. 13). This fact is supported by local farmers, who report lower crop yields on Morrill soils (exhumed paleosols) in normal years than on adjacent soils formed in Wisconsinan loess or Holocene colluvium.

Hypothesis IV, regarding organic matter decomposition in paleosols, was tested in the following manner. Organic matter percentages for the uppermost meter of all sampled paleosols were summed. The samples were generally spaced 25 cm apart, resulting in four values/meter. The summed values thus compiled are plotted vs depth-to-paleosol in Fig. 14. A logarithmic function is the result:

$$\log Y = 0.612 - 0.141 \log X \tag{4}$$

where Y is the additive organic matter total for the uppermost meter of the paleosol and X is the depth to the top of the paleosol. This equation explains only one-third of the organic matter variation ($r^2 = 0.33$), yet is significant at the .01 level, and does reveal the common type of organic decay function seen in soils. A

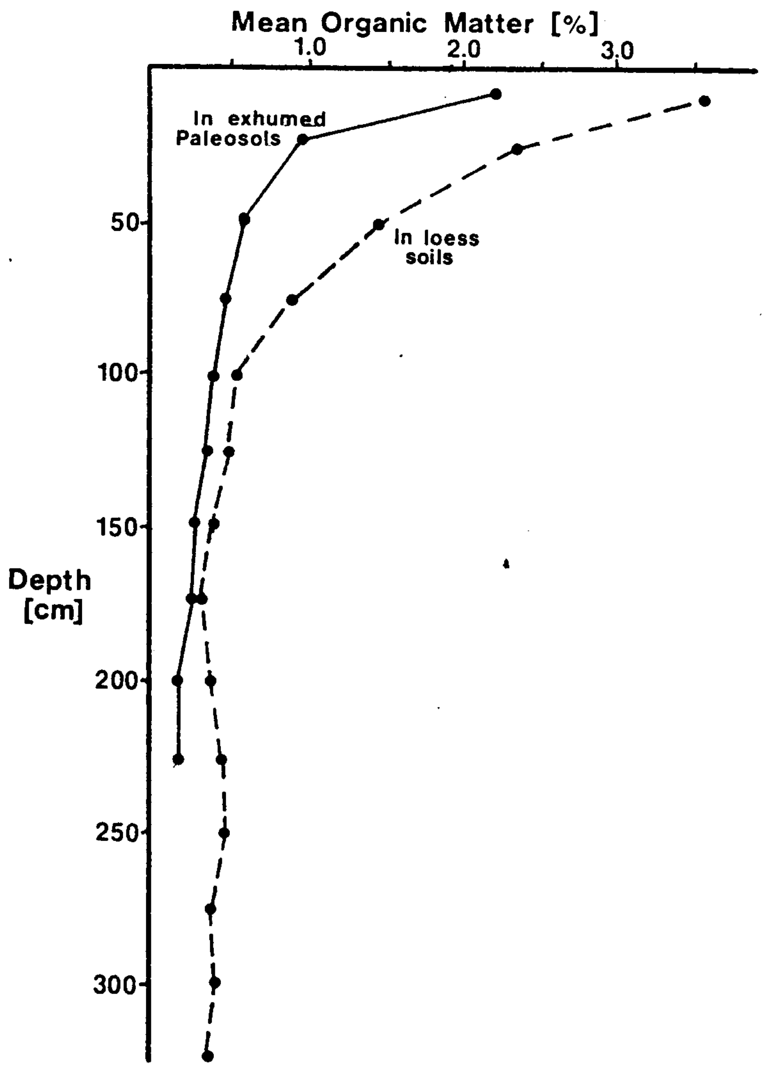

Figure 13. Mean organic matter composition of soils formed in loess, and of exhumed paleosols, in the study area.

comparison of Figs. 13 and 14 suggests that organic matter contents of paleosols reach constant levels at nearly the same depth as loessal soils (approx. 175 cm). Therefore, hypothesis IV is confirmed, in that paleosols do not retain significantly larger amounts of organic matter at great depth than do surface soils. The retention of organic matter within a paleosol is a decreasing logarithmic function and thus is similar to that of modern soils.

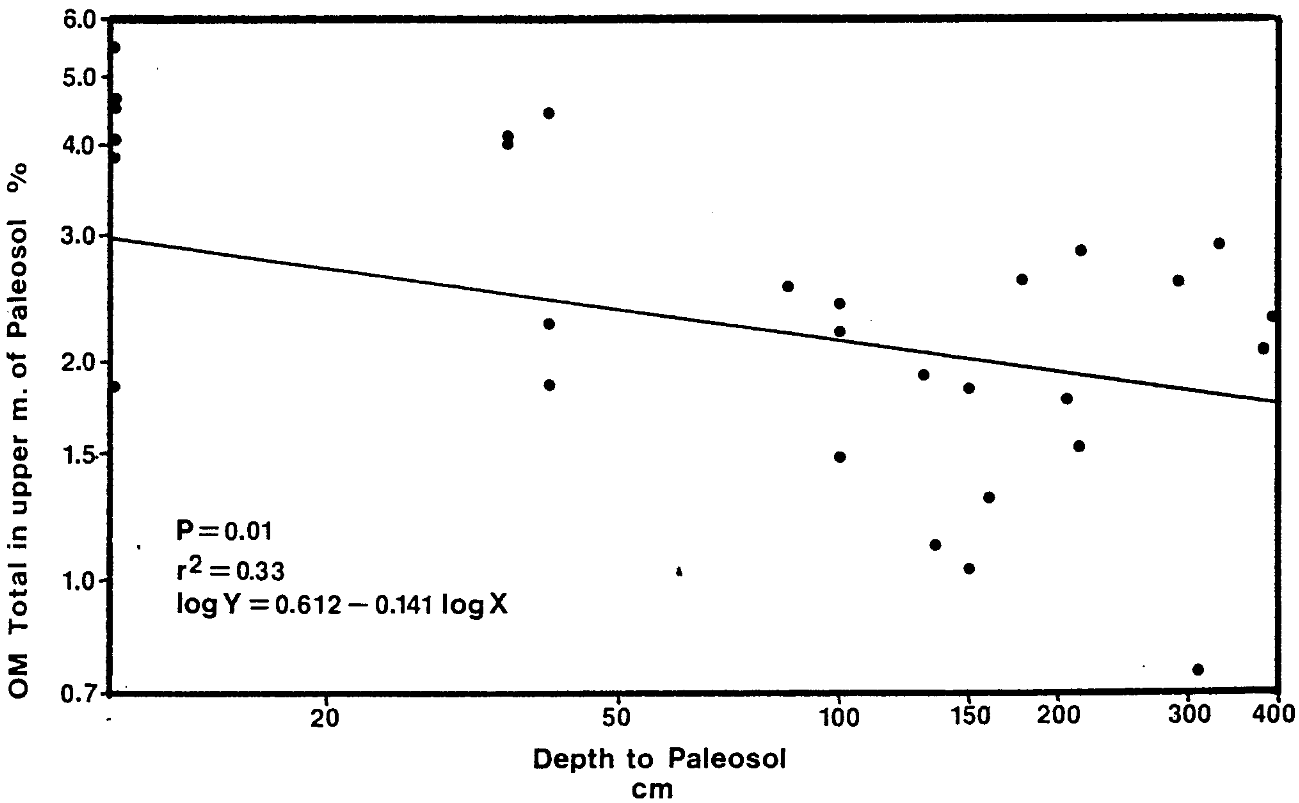

Figure 14. Relation of organic matter content in topmost meter of paleosols with burial depth.

Lastly, Fig. 14 suggests that the "depth of isolation" in paleosols buried by loess, with regard to organic matter, is approximately 175-200 cm. Paleosols buried much more deeply than 200 cm have essentially no less organic matter than those at this depth.

Standard logarithmic decline in organic matter content is often interrupted by the presence of a paleosol. Minor secondary organic peaks are observed both within and above many paleosols. Ten of fifteen stratigraphic columns with deeply buried paleosols show pronounced organic peaks from 10-65 cm above the visual tops of paleosols (mean: 38 cm). Thirteen deeply buried paleosols also contain organic maxima within the sola (Fig. 15).

Organic peaks above a buried paleosol are probably due to 1) illuvial humus, 2) decay of roots which preferentially seek out occasional perched water over the soil, or 3) relict organic matter which accumulated as very slow loess additions buried the soil. Retention of humus tightly bonded to or within clay minerals (Stevenson 1969), or protected by chemical and physical protective "spheres" (Gerasimov 1971), may be the explanation for small organic peaks in fine textured paleosols. In any event, nearly complete oxidation of humus within deeply buried paleosols renders examination of the original organic matter distribution impossible. Hence, the paleosol (or paleosols) of the study site cannot be identified as either a Mollisol or an Alfisol, as has been done by others (Thorp et al. 1951, Simonson 1954, Ruhe 1974), who report it to be a grassland Mollisol.

The presence of a buried paleosol has a rather pronounced impact on the total organic matter 1) content and 2) depth function within the surface soil, predominantly by influencing humus illuviation, rooting pattern, and pedoturbative processes. Organic matter decreases gradually in loess soils above a deeply buried paleosol as exemplified by sites 6-1 and "Pit Face" in Fig. 16. A shallowly buried paleosol, however, significantly alters this patter (Fig. 16). The organic matter curve declines rapidly with increasing depth when above a shallowly-buried paleosol, probably due to preferential rooting and pedoturbation above the buried solum. Inexplicably, the sharp decline in organic matter content with depth persists for exhumed paleosols (Fig. 16, Site 2-1).

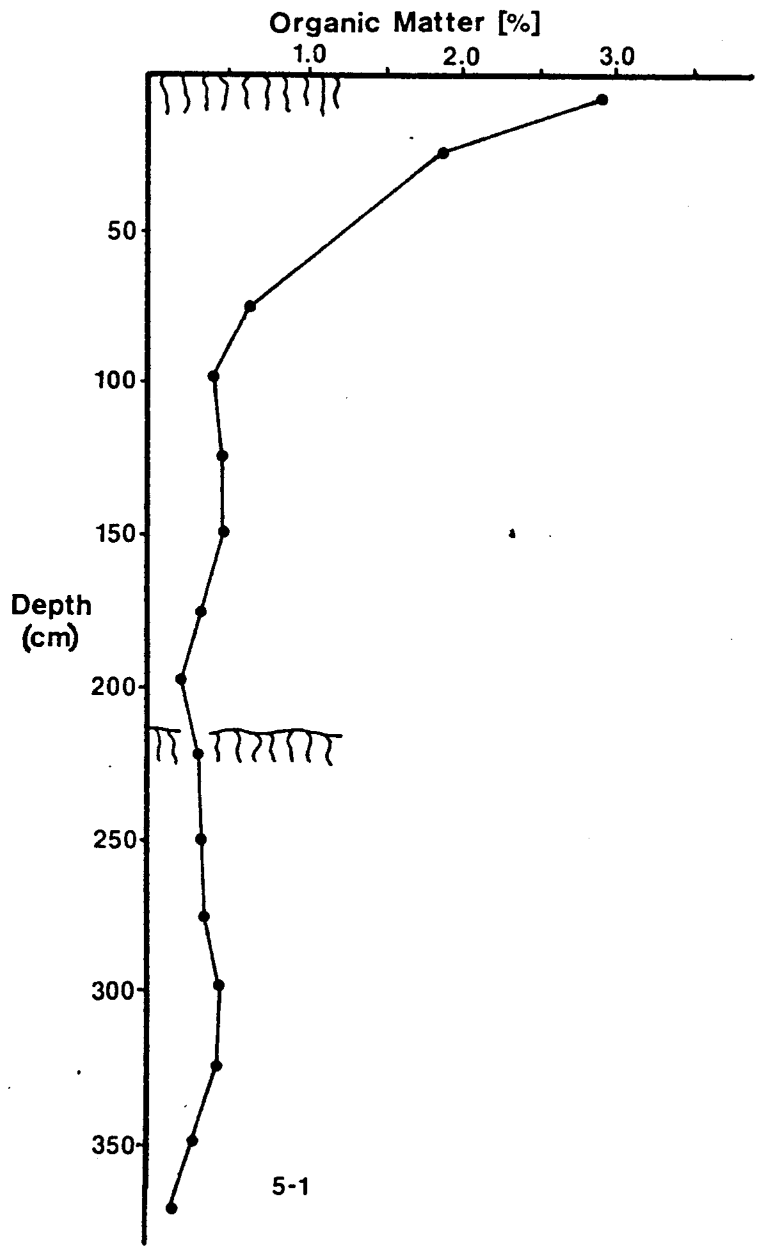

Figure 15. Stratigraphic column exhibiting organic maximum at the surface and secondary maxima above (150 cm) and within (300-320 cm) a buried paleosol.

One method of comparing organic matter contents between surface soils is by examining the 0-10 cm organic matter maximum. Because organic matter values exhibit either sharp or gradual declines with depth, depending on local parent material stratigraphy, a better way was devised. The total area to the left of the organic matter curve to a depth of one meter was calculated, using standard plane geometry (Fig. 17). This method takes into account the cumulative organic matter values for the topmost meter of soil. When surface soils are compared in this manner, it is readily observed that units of total organic matter in the upper meter of soil increase with increasing paleosol burial depth (Fig. 18). The relationship is expressed by:

$$\log Y = 2.123 + 0.19 \log X \qquad (5)$$

where Y is the units of organic matter in the topmost meter of soil and X is the depth to the Btb max. (The relationship holds nearly as well when one lets X represent the depth to the visual top of the paleosol.) Equation 5 explains 38% of the relationship between surface organic matter and paleosol burial depth ($r^2 = 0.382$). Significance at .01 was obtained. Two cores, inadvertently taken within terrace drainageways and thus having abnormally low organic matter unit totals, were not included in the sample. Three others were discarded because of lack of a clay maximum within the paleosol.

The relationship in Eq. 5 demonstrates the influence of a buried paleosol on a surface soil. Plant growth is inhibited by a buried, clay-rich paleosol. Loess has better tilth and weathers to release

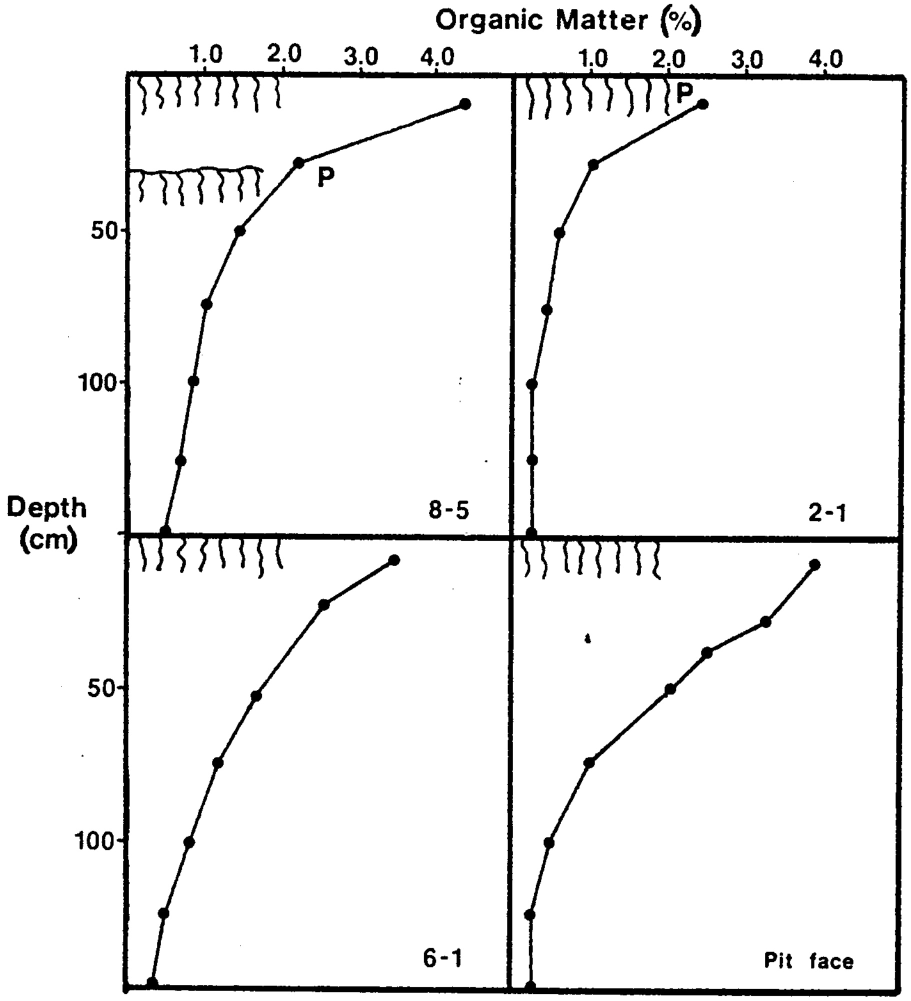

Figure 16. Organic matter depth curves in soils with deeply buried paleosols (Sites 6-1 and Pit Face Site), a shallowly buried paleosol (Site 8-5), and an exhumed paleosol (Site 2-1).

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more bases than Sangamon paleosols, promoting plant growth and, hence, organic accumulation in the surface soil.

Clay Mineralogy. As outlined earlier, samples selected for clay mineralogical analysis were representative of 1) deep, relatively unweathered Peorian loess, 2) the Bt max of a surface soil formed in the same loess, and the clay maxima of paleosols developed in 3) Kansan till, 4) Loveland loess and 5) outwash sands. The results

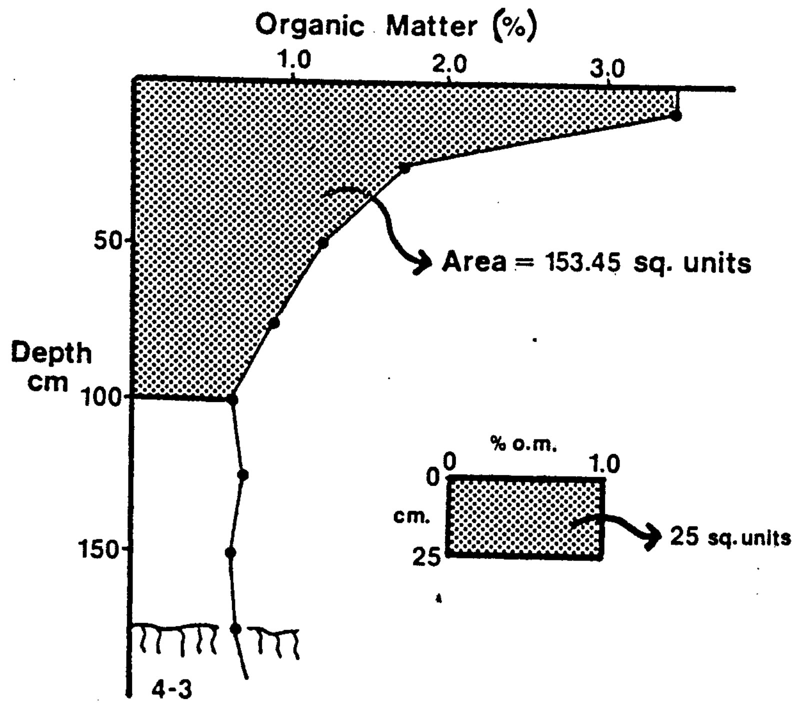

Figure 17. Computation of organic matter units within the upper meter of soil.

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are reported in "relative" or semi-quantitative form (Table 3). Estimation of actual percentages of clay minerals in a sample, based on x-ray diffraction data alone is believed not acceptable (Whittig 1965; pers. comm. Dr. G. James, Kans. Geol. Surv., 1983), though it has been attempted (Norrish and Taylor 1962, Hall 1973, Ruhe 1974, Ruhe et al. 1974, Glass, Frye, and Willman 1968, Olson et al. 1978, Ruhe and Olson 1980). Semi-quantitative, or comparative, methods based on counts per second at diffraction peaks, are the best that can be currently hoped for. These methods do not attempt to estimate the actual percentages of a clay mineral in a sample, but rather only compare the relative intensities of peaks between two or more minerals. The best known such method is the diffraction index (DI) of Frye et al. (1962), which has been used

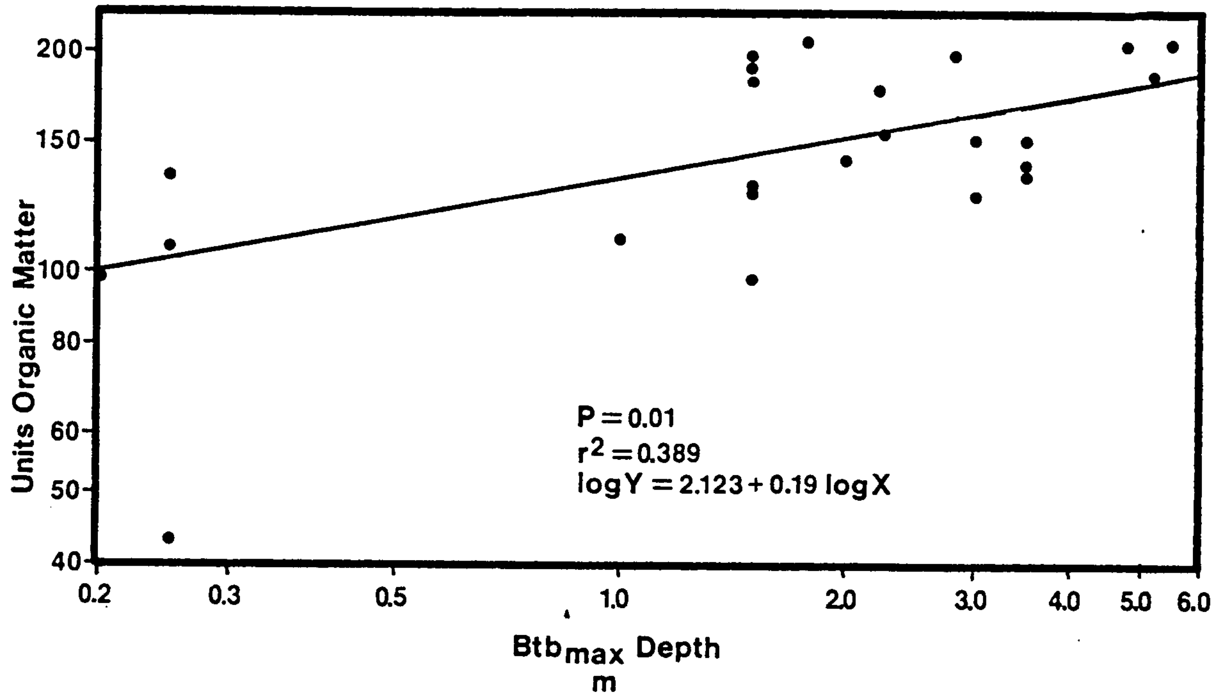

Figure 18. Relation of organic matter content of surface soils to depth above a buried paleosol.

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successfully by Tien (1968) and Hall (1973). This study, however, utilizes a ranking scheme whereby clay minerals are assigned one of five class ranks, from trace to dominant. This method was selected because it has been used frequently by previous investigator (e.g., Darmody and Foss 1982, Mausbach _et al_. 1982), and appears to give acceptable results.

Interpretation of x-ray diffractograms follows Carroll (1970) and Whittig (1965). Peaks at 7.13 Angstroms (A) not affected by glycolsolvation were interpreted as kaolinite plus chlorite. However, the absence of a secondary chlorite reflection at 7.13 A (12.4^{o} 2 θ) after heating to 550^{o} C indicated the absence (or near absence) of chlorite in all samples. Illite peaks (from glycolated samples) were observed on the plotter trace at 10.2 A (8.7^{o} 2 θ), but

Table 3

Relative Clay Mineral Abundance in Paleosols of Differing Lithology, and of Loess

Sample Site & Depth	Pedo-Stratigraphic Layer	Size Fraction	Kaol.	Mont.	Verm.	Mixed Layer	Illite
Pit Face, 545 cm	Till Paleosol	Coarse clay	**	*	*	**	***
"	"	Fine clay	**	****	***	***	****
5-7, 225 cm	Loveland Loess Paleosol	c.c.	***	*	*	**	***
"	"	f.c.	**	***	**	*****	**
3-9, 350 cm	Outwash Paleosol	c.c.	***	***	**	****	***
"	"	f.c.	***	*****	****	*****	****
3-9,	Peorian Loess	c.c.	*	***	**	***	**
175 cm	"	f.c.	*	****	***	****	**
4-7,	Surface Soil	c.c.	**	***	***	***	***
50 cm	"	f.c.	**	**	***	****	***

Key: ***** Dominant   *** Moderate   * Trace
**** Abundant   ** Small

were masked by broad swelling clay reflections on other traces. Intense, broad peaks from 5.0-7.0^{o} 2 θ (and up to 8.5^{o} 2 θ) were common especially on the fine clay traces and interpreted as follows: reflections from 5.0^{o}-6.0^{o} (17.6-14.7 A) which expanded to 18 A (4.8^{o}-5.0^{o}) upon glycolation were labelled montmorillonite; adjacent untreated reflections from 6.0^{o}-6.4^{o} (14.7-13.8 A) were interpreted as vermiculite clay, and reflections beyond 6.4^{o} (to 8.0^{o}) which collapsed to approximately 9.0^{o} (13.8-9.8 A) were grouped under "mixed layer clays." Relative amounts of montmorillonite, vermiculite and mixed layer clays were determined from the untreated trace; often, because of the masking effects of the vermiculite peak on that of the mixed layer clays, the amount of the latter was estimated after glycolation. In general, the fine clay fraction produced better and more easily interpreted reflections than the coarse clays.

Overall, mixed layer clays and montmorillonite dominate the clay mineral suite of the Pleistocene deposits in the study area. Tien (1968) determined the clay mineralogy of many lower Pleistocene glacial and glaciofluvial deposits in adjacent Doniphan county. He found abundant amounts of a "mixed layer mineral", along with lesser amounts of kaolinite and illite. Tien believed the "mixed layer mineral" to be composed of randomly interstratified layers of montmorillonite and mica (illite). However, the reflections which Tien has interpreted as mixed layer the present author has called montmorillonite. The weathered nature of the montmorillonite in the present study is evident from its broad peak reflections.

Clay from relatively unaltered Peorian loess appears to contain no, or at best only very small traces of kaolinite (Fig. 19). Montmorillonite and mixed layer clays dominate both the coarse and fine clay (Table 3). This analysis closely agrees with that of Swineford and Frye (1951), who found the loess in Doniphan county to be primarily composed of montmorillonite and illite in the clay fraction, and Glass et al. (1968), who reported near total dominance by montmorillonite in the same loess deposits. Likewise, Castellano (1961) reported that Peorian loess deposits in Nebraska contain similar clay assemblages. On a broader scale, Beavers (1957) summarized the clay mineralogy of calcareous Peorian loess in Illinois, Kansas, Iowa, Nebraska, and Missouri, and concluded that the loess is principally montmorillonitic.

Analysis of the clay from the Bt max of a Sharpsburg soil (Site 4-7: 50 cm) revealed a mixed assemblage, with slight dominance of mixed layer clays in the fine fraction. Because most of the fine clay in the pedon is probably illuvial, this study suggests that mixed layer clays, abundant in the fine clay fraction, are preferentially moving downward by lessivage. Montmorillonite appears to be stable in this regard; i.e. moving relatively slowly, if at all, because of its near absence in the fine clay sample (Table 3).

Studies of Sharpsburg soils in southeastern Nebraska by Al-Barrack (1980) and Al-Janabi and Drew (1967) make interesting comparisons with the present research, in that similar parent materials and only slightly different climates are involved. Al-Janabi and Drew reported that coarse clays contain moderate amounts of montmorillonite, illite and kaolinite, and small quantities of vermiculite. Montmorillonite dominated the fine clays; illite was sub-

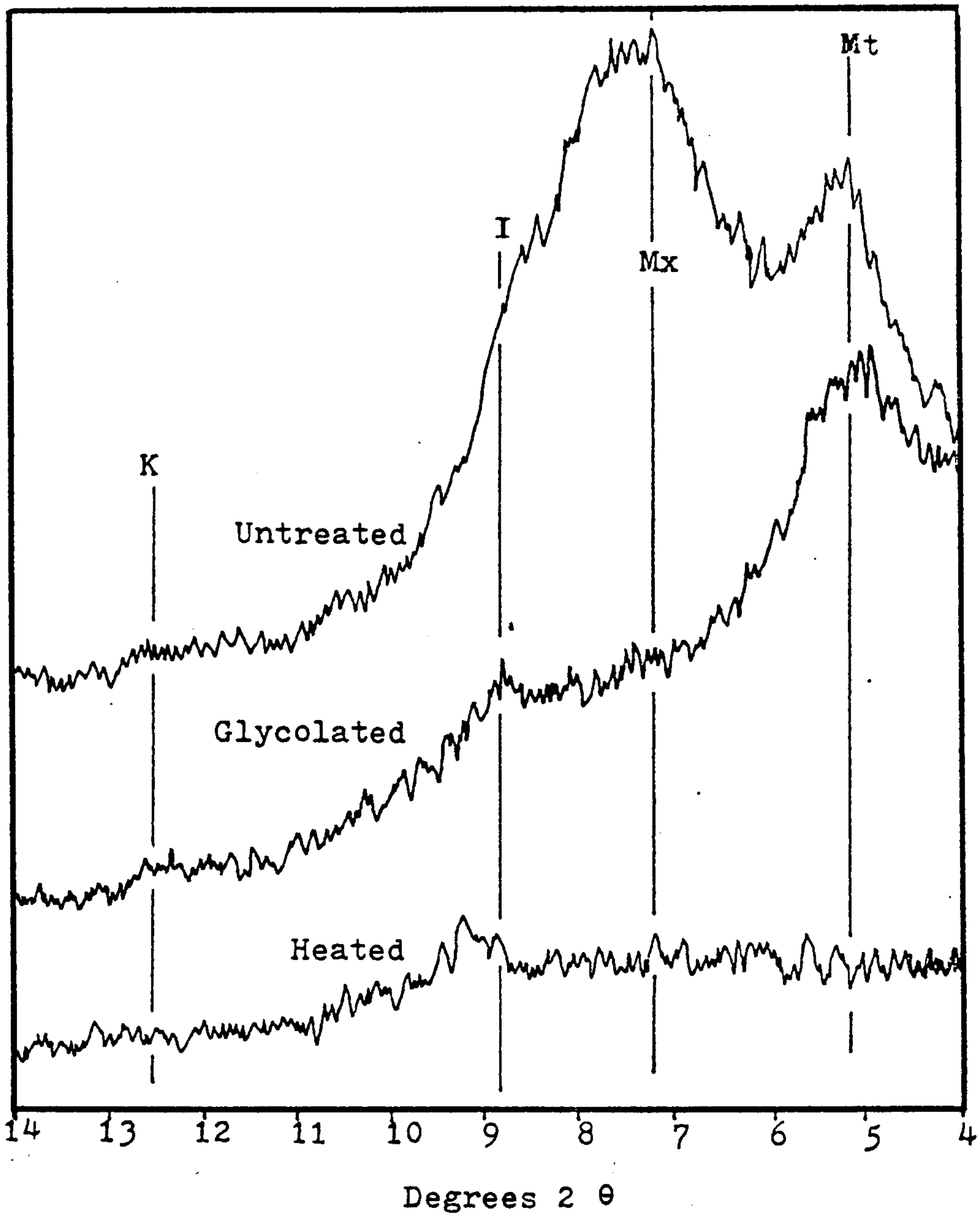

Figure 19. X-ray diffraction traces from deep Peorian loess. Fine clay sample. K:Kaolinite, I:Illite, MX:Mixed Layer, Mt:Montmorillonite.

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ordinate. Al-Barrack sampled Sharpsburg and Wymore (fine, montmorillonitic, mesic Aquic Argiudoll) soils developed in loess. Coarse clays from these soils were predominantly montmorillonitic with smaller amounts of vermiculite, illite and kaolinite. Likewise, the fine clay fraction was dominated by montmorillonite, but also contained illite and a mixed layer component. Al-Barrack noted considerable weathering of montmorillonite in the loess at a lower loess/paleosol contact. The above studies revealed similar clay mineral assemblages for surface soils in Nebraska formed in loess, compared with those of the present study. Data from this research, however, suggested smaller amounts of montmorillonite in the fine clay fraction, and more mixed layer clays than stratigraphically similar Nebraska soils.

The paleosol exposed in the Swaim pit was sampled at the Btb max (550 cm), in Kansan till. Illite and montmorillonite dominated the finer clays, but the latter is conspicuously absent from the coarser fraction (Table 3). The presence of montmorillonite and vermiculite only in the fine clay fraction suggests their origin as illuvial from the more weathered till above. Kaolinite, while present in small amounts, was not as prevalent as in the tills of Doniphan county (Tien 1968) or southeastern Nebraska (Al-Barrack 1980).

As in till paleosols, the purported Sangamon paleosol which formed in Loveland loess lacked montmorillonite and vermiculite in the coarse clay fraction (Table 3). Mixed layer clays were especially dominant in the finer size class (Fig. 20). Kaolinite was well represented (Fig. 21). Overall, the Loveland loess clay assemblage was very similar to that of till, which may suggest colluvial or pedoturbative mixing of the two on the paleolandscape.

As expected, in well-mixed outwash sediments, the clay assemblage was well complimented with all varieties (Table 3). Clays are the final sediments to settle out of water, allowing ample mixing by continued cut and fill episodes on either outwash plains or in valley trains. Also, they are capable of being carried great distances in water because only very slight amounts of turbidity can keep them in suspension. The clay suite of the outwash materials sampled in this study appears to have come from a variety of sources. All mineral species were well represented.

A very brief look at the clay mineral assemblage of several pedo-stratigraphic layers in the study area revealed several interesting generalities. The paleosols, in general, had slightly higher quantities of kaolinite than the loess or the surface soil formed in loess, which is in agreement with many of the studies of Ruhe (1974, Ruhe _et al_. 1974, Olson _et al_. 1978, Ruhe and Olson 1980). Kaolinite clay from deep loess unaffected by most surface soil-forming processes was conspicuously absent. The 2:1 clays (montmorillonite, vermiculite and mixed layers) were prevalent in finer clay fractions, especially so in paleosols where their abundance is ascribed to lessivage (all samples within paleosols were taken at depths of clay maxima). Illite was common in all samples. Overall, the _complete_ and _diversified_ clay mineral assemblage observed in most samples from the study area stands as the major inconsistency between the present study and others done in northeastern Kansas (Tien 1968) and southeastern Nebraska (Al-Janabi and Drew 1967, Castellano 1961). In the latter studies, one or two clay minerals dominated the suite, whereas in Brown county all minerals were usually represented in the Pleistocene deposits.

Weathering Ratios Within Paleosols. Quartz/feldspar weathering ratios were calculated from the fine sand + coarse silt fraction

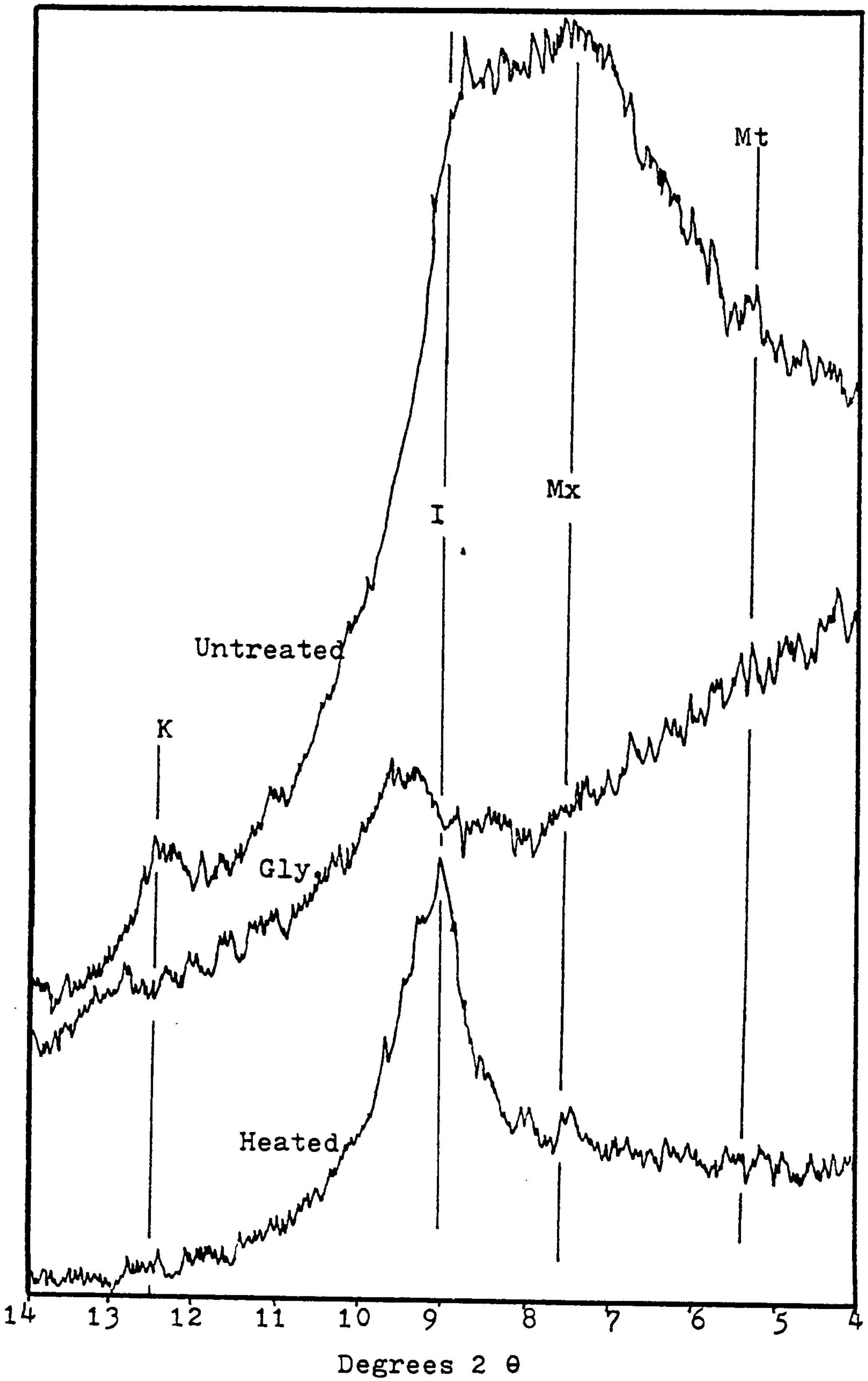

Figure 20. X-ray diffraction traces from a paleosol developed in Loveland loess. Fine clay. K:Kaolinite, I:Illite, Mx:Mixed layer, Mt:Montmorillonite.

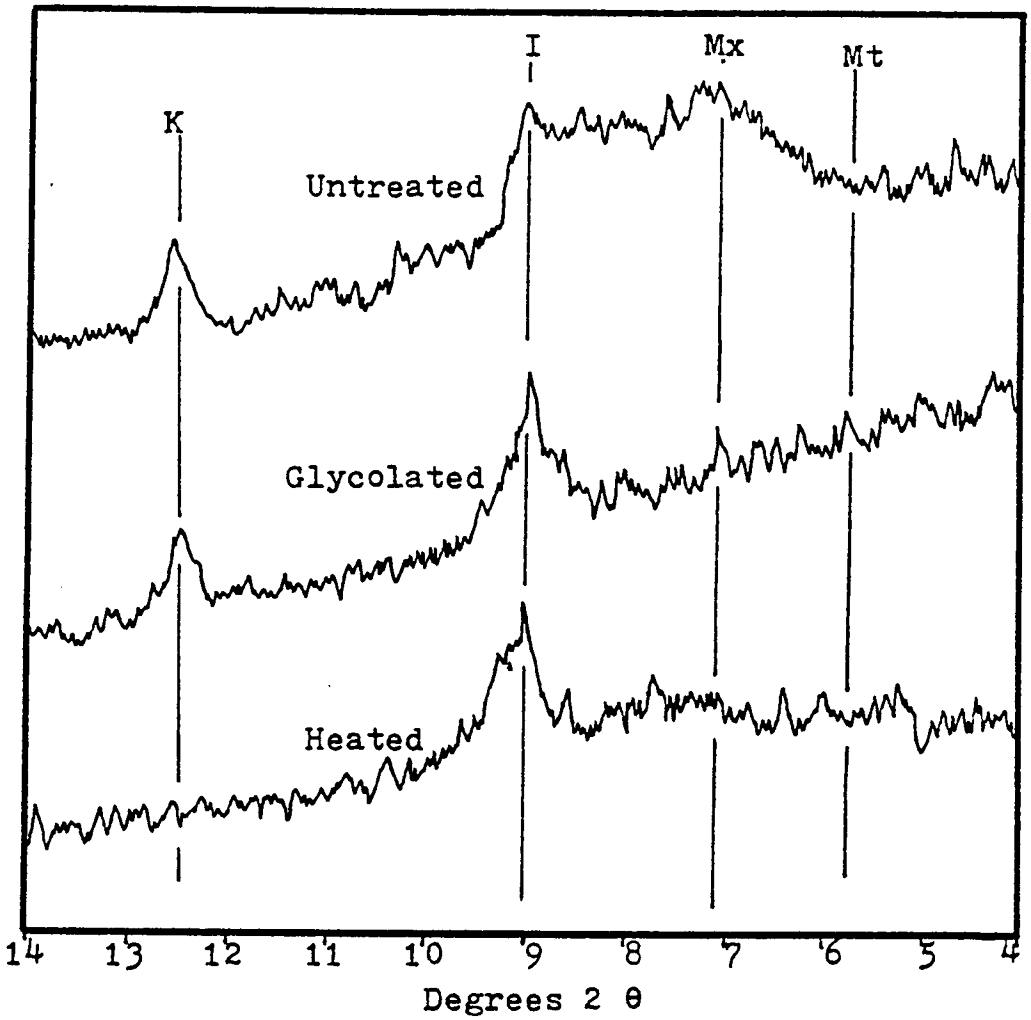

Figure 21. X-ray diffraction traces from a paleosol developed in Loveland loess. Coarse clay. K:Kaolinite, I:Illite, Mx:Mixed layer, Mt:Montmorillonite.

from the Btb horizons of all paleosols sampled and the results plotted in Fig. 22. There may be a weakly defined non-significant ($P > .05$), relationship between weathering within the paleosol and burial depth. The equation describing this relationship is:

$$Y = 1.867 - 0.115 X \tag{6}$$

where Y is the Q/F at the Btb max of paleosols and X is the sample depth. The very weak trend ($r^2 = 0.11$) is to be expected when working with paleosol samples from Kansan till, outwash sands, Loveland loess and various mixed zones between the units. Indeed, evidence of any trend at all, given the inter-pedon variation in 1) parent materials, 2) topographic position, and possibly 3) age, would have been remarkable. The data appear to indicate that exhumed and shallowly buried paleosols may be slightly more weathered than equivalent deeply buried counterparts though the relationship is not significant. Therefore, hypothesis II, which supposed increased weathering in shallowly buried paleosols, is rejected. Further research is necessary to arrive at more definitive trends, if indeed such trends exist, at other locations.

<u>Swaim Gravel Pit Exposure</u>. The best complete exposure of Late Pleistocene deposits in Brown county is located in the Swaim pit (Fig. 6). The pit was originally used as a source of gravel, but is presently abandoned.

The exposure (Plate 2) reveals three to four meters of Peorian and Bignell (undifferentiated) loess over a paleosol developed in Kansan till and gravel. Till thickness above gravel varies throughout the pit, but 2-3 m is common. Most of the gravel have been removed from the center of the pit and basal gravel units are often covered by slump deposits at most sidewall exposures. A complete profile description of the sampled face follows.

Swaim Pit Profile Description

0–18 cm	Ap	Dark brown (10 YR 3/3) silty clay loam; moderate very fine crumb structure; very friable; clear smooth boundary.
8–20	B1	Dark brown (10 YR 3/3 to 10 YR 3/4) silty clay loam; moderate very fine subangular blocky and granular structure; friable; clear smooth.

20-30	B21	Dark brown (10 YR 3/3) silty clay loam; moderately strong very fine and fine subangular blocky structure; friable; gradual, wavy.
30-46	B22	Brown (10 YR 4/4 to 10 YR 4/6) silty clay loam; moderately weak very fine subangular blocky structure; friable; gradual, wavy.
46-95	B3	Dull yellowish brown (10 YR 5/4) to brown (10 YR 4/4) silty clay loam; weak very fine subangular blocky structure; firm; diffuse.
95-132	C1	Dull yellowish brown (10 YR 5/4) silt loam; massive to very weak fine to medium prismatic structure; firm; gradual, wavy.
132-206	C2	Dull yellow orange (10 YR 6/4) silt loam; massive to weak coarse prismatic structure; abundant horizontal mottled streaks (Grayish olive 5 Y 6/2) and iron bands (Reddish brown 5 YR 4/8) intermixed and stratified; firm; gradual, smooth.
206-317	C3	Dull yellow orange (10 YR 6/3) silt loam; massive; common very dark brown (7.5 YR 2/3) iron and/or manganese stain spots (<8 mm); firm; diffuse, smooth.
317-360	C4	Mixed layer; grading from dull yellow orange (10 YR 6/4) to brown (10 YR 4/4) at base; massive, firm; gradual, smooth.
360-411	C/Ab	Brown (7.5 YR 4/4) silt loam; massive; firm; rare gravels present; gradual, smooth.
411-449	IIA1b	Dull brown (7.5 YR 5/4) silt loam; very weak very fine subangular blocky structure; firm; rare gravels present; gradual, smooth.
449-500	IIB1b	Dull brown (7.5 YR 5/4) and brown (7.5 YR 4/4) silt loam; very weak very fine subangular blocky structure; firm; common gravels; gradual, smooth.
500-590	IIB2tb	Reddish brown (5 YR 4/6) gravelly clay loam; weak fine subangular blocky structure; very firm; gradual, smooth.
590 +	IIIB3b	Covered

Colors for moist soil, from Japanese equivalent of Munsell color book. Entire profile is noncalcareous.

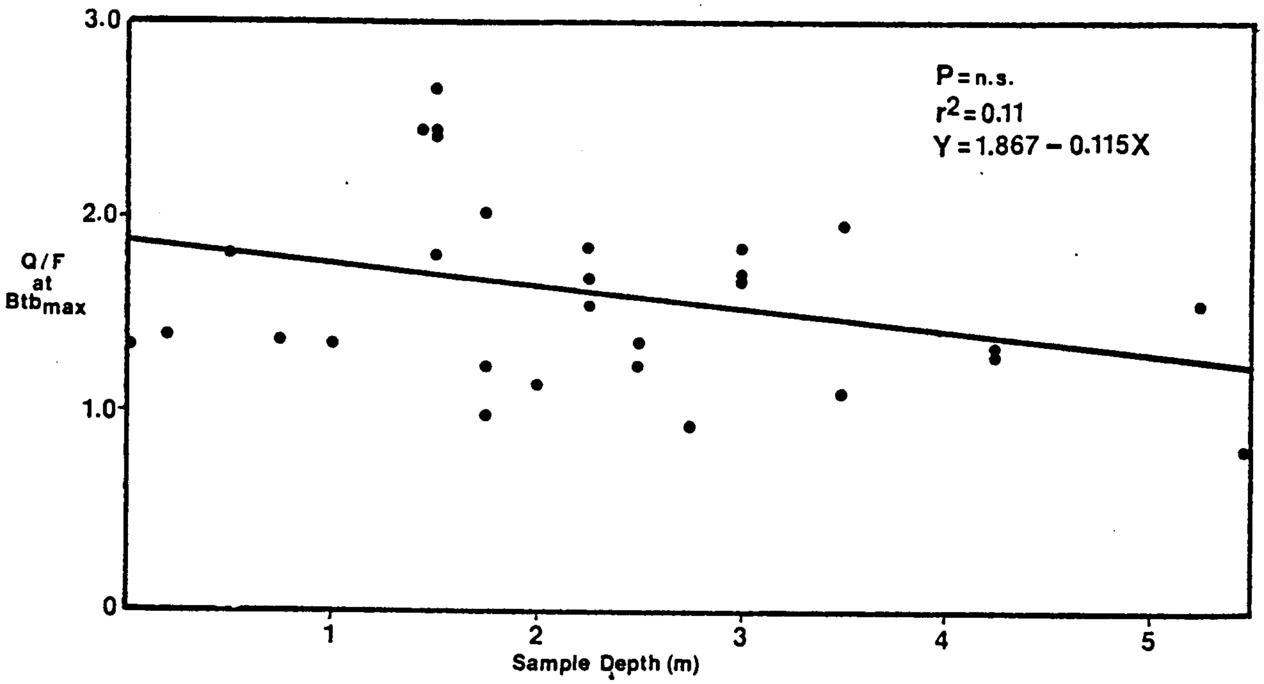

Figure 22. Relation between quartz/feldspar ratios in paleosols and burial depth.

Quartz/feldspar ratios within the paleosol in the Swaim pit ranged from 1.11 to 1.13 in the till of the upper solum (Fig. 23). Below, in the gravels, a high value of 1.41 was observed. High Q/F ratios in sands and gravels are to be expected; only the most resistant minerals are retained in such deposits. Low ratios (0.43-0.66) characterize the relatively unweathered loesses above. The Sharpsburg soil has formed in the surficial loess. It is characterized by moderate increases in weathering when compared with deeper loess (Fig. 23), though a smooth decrease with depth as might be expected in mature soils (See Ruhe 1956) is lacking.

Caspall (1970) distinguished Bignell from Peorian loess in Doniphan county on the basis of consistence. Peorian loess was very firm whereas Bignell was crumbly and friable. No Brady paleosol, which would easily separate the two loesses, could by recognized in the Swaim pit. A clear division based on consistence is, however, present at 46 cm. Sediments above 46 cm are friable; those below

Plate 2. Profile at the east central face of the Swaim Pit. NE 1/4 of SW 1/4 of Sec. 28, T1S, R18E, Brown Co., Kansas. Note buff color of Wisconsinan loesses, in contrast to the redder hues of the paleosol (below 3 m). Sampled intervals are shown.

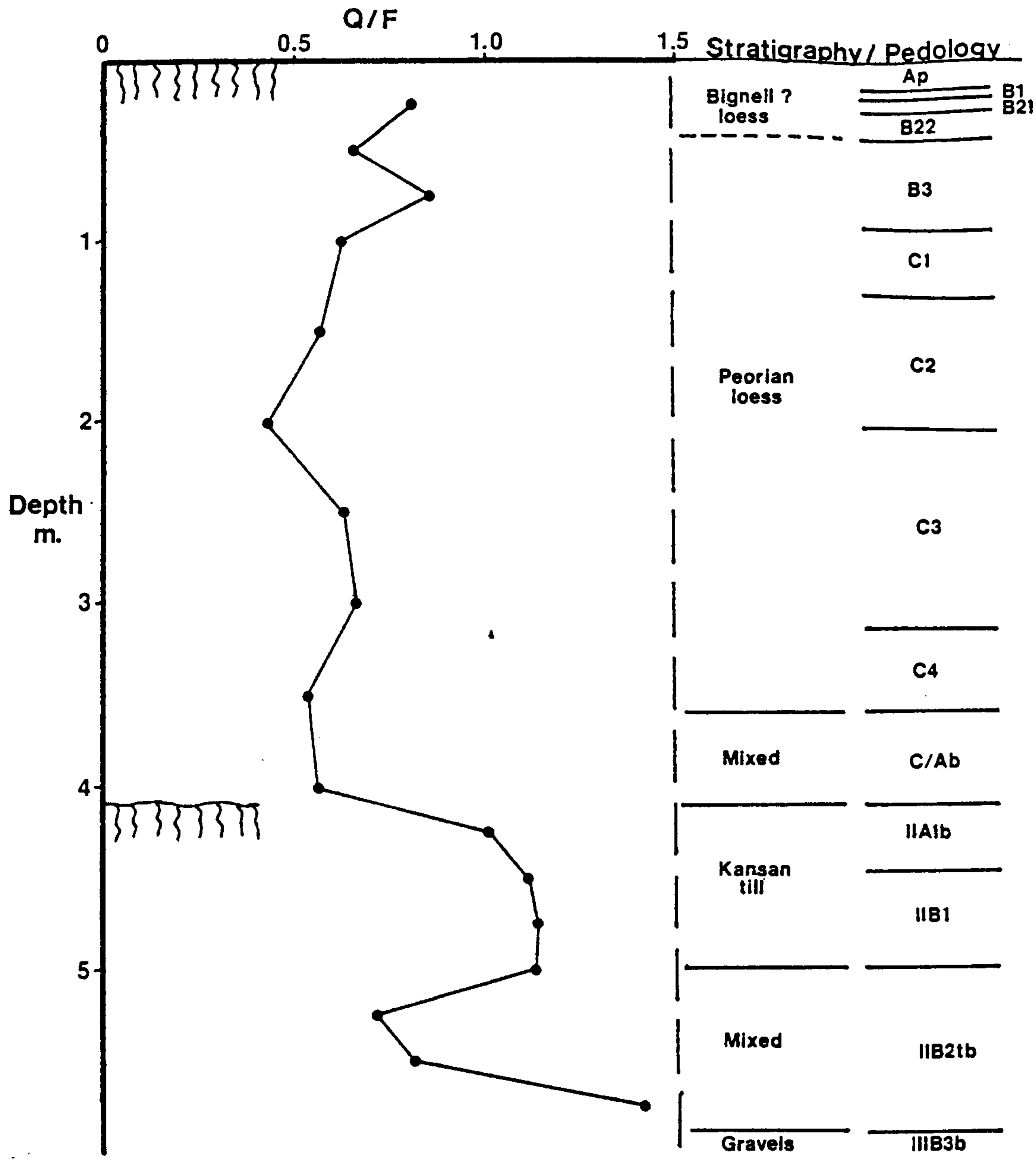

Figure 23. Quartz/feldspar ratios, stratigraphy and pedology at the Swaim Pit exposure.

are firm. Therefore, this may be a possible stratigraphic hiatus, though a definitive statement cannot be made based on the data available.

Future Research Suggestions. The need for future research on post-burial alteration is considerable, yet I believe, is overshadowed by an even greater need for studies of the Pleistocene deposits in Brown County. It is surprising that so little has been done in the county when one views the substantive research which has been carried on just 15-25 km east in Doniphan County (e.g. Frye and Leonard 1949, 1952, Tien 1968, Caspall 1970). There are several fresh gravel and borrow pits in the county which I have examined, yet have chosen not to interpret. They have not been discussed in the present report. I urge future geographers and geologists interested in the Pleistocene not to overlook Brown County. The potential for Pleistocene research in Brown County is exceptional.

SUMMARY

Paleosols crop out on hillsides in a limited area of northeastern Brown county, Kansas where loess deposits thin to reveal an otherwise buried paleosol. Upslope these paleosols are buried by Wisconsinan loesses. The paleosol is reddish-brown, well-developed and has formed in either Loveland loess (Illinoian) or Kansan till. At any location outwash sands may be present below the till and in the lower sections of the profile.

Texture of the paleosol is clay loam or silty clay loam. It is finer textured than surface soils developed in loess. Maximum clay content of the paleosol decreases approximately 1% with each increasing meter of burial. This decrease is ascribed to surficial weathering in exhumed and shallowly-buried paleosols. Quartz/feldspar ratios at the Btb max of paleosols substantiate this deduction. Ratios generally increase in shallowly buried and exhumed paleosols, indicating greater weathering.

Variable depths of burial result in many composite and welded profiles between surface soils and paleosols. Soil pH, however, was unaffected by the presence of a paleosol anywhere in the stratigraphic sequence. pH was a minimum at the surface (mean value = 5.08) and increased to a constant level of about 6.4 at 150 cm, regardless of the presence or absence of a paleosol. Exhumed paleosols exhibited somewhat higher pH values than surface soils.

Organic matter content of paleosols buried more deeply than 200 cm was very low and essentially constant with depth, indicating

near complete oxidation. Exhumed paleosols contained substantially less organic matter than modern soils. Surface soil organic matter content varied as a direct function of depth-to-paleosol. Systematic decreases in organic content with depth was much more gradual in soils over deeply-buried, as opposed to shallowly-buried paleosols. Minor secondary organic matter peaks above paleosols were common and ascribed to 1) illuvial humus, 2) decay of roots which preferentially seek out occasional perched water, or 3) relict organic matter from periods of slow burial by loess. Some deeply buried paleosols also contained small organic matter peaks within the solum. The original organic matter distribution within the paleosol could not be ascertained.

Clay mineralogy of paleosols, modern soils, and loess revealed strong dominance by 2:1 mixed layer and swelling clays. Kaolinite was present in moderate amounts only in paleosols, and was less abundant or absent in loess and modern soils. Illite was present in all deposits. The complete and diversified clay mineral assemblage of the Pleistocene deposits of the study area stands as a discrepancy from other studies in the area.

An exposure of Pleistocene deposits in a gravel pit in the study area was analyzed and described. Based on the promising results from this pit, future work on Quaternary stratigraphy is encouraged in the remaining pits in the county, which are more complex.

The major contributions of this study to the body of paleogeographic literature are twofold. Documentation and study of Sangamon paleosols in Brown county is perhaps the major gain. Ascertaining absolute depths, or the lack of, below which paleosols nearly cease to change in character, is the other.

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APPENDIX I

Preliminary Investigations of the Quaternary Stratigraphy of the Study Area.

Though the major emphasis of this thesis has been on paleosols and their alteration, some mention of the Pleistocene stratigraphy, beyond that in the text, is appropriate. The Pleistocene exposures adjacent to the Missouri River in Doniphan county are well document (Frye and Leonard 1949, 1952, Tien 1968, Caspall 1970). Pleistocene research is conspicuously lacking for Brown County. The only discussions are (to the author's knowledge) brief overviews by Frye and Leonard (1952), Eickelberry and Templin (1960), and Bayne and Schoewe (1967). Further, no definitive statements to this purpose can be advanced from the present study. Preliminary stratigraphy is included here as a guide for placing paleosols in their proper stratigraphic sequence.

The sands and gravels encountered in cores throughtout the study area may have one of two origins: 1) they are Kansan-age outwash (Grand Island or Atchison formation), 2) they are lenses within the till, which are common, according to Bayen and Schoewe (1967). Discussions with local farmers yield strong evidence for an outwash origin, especially near major stream valleys where sand and gravel depths are substantial. It is likely, however, that some of the upland cores which penetrated sands at depth are only in sand lenses in the till.

The Kansan till was probably laid down directly upon the proglacial outwash. Depths of till reach 35 m in the county. Evidence of Afton soils is absent in Brown County; the nearest Afton soil is in Doniphan county, the next county to the east, and it is morphologically very unlike the paleosols of the study area. Limited sampling, however, prevents ruling out the possibility of superposed Afton and Sangamon soils in Brown County.

Extensive and convincing evidence for a Late Sangamon erosion period in Iowa (Ruhe 1956, 1969) creates questions about the ages of paleosols in the study area. Ruhe reports truncation of paleosols on sideslopes and preservation of older soils on summits. In the study area paleosols formed in Loveland loess on the uplands can be laterally traced onto till as the sideslopes are approached (Fig. 6). The lack of Loveland loess on some sideslope positions of the paleolandscape may be ascribed to a lack of loess deposition there, or a Late Sangamon erosion episode. The latter seems more likely (Caspall 1970), though greater thickness of Loveland loess deposits on uplands is a realistic assumption. If a Late Sangamon erosion episode did occur, then all paleosols in the study area would then be of Sangamonian age and, presumably, the Afton soil should be preserved deep beneath the uplands.

Peorian and Bignell loesses were deposited on the paleosol during middle and late Wisconsinan time. No paleosol (Brady soil) is believed to have developed in northeastern Kansas between these two depositional episodes. Rather, evidence of only a slight period of leaching of the Peorian loess is presented by Caspall (1970).

During the Late Pleistocene and Holocene rejuvenated erosion episodes truncated the paleosol in the valleys, leaving Shelby soils to develop in freshly exposed oxidized Kansan tills. Upslope, a loess cover protected the paleosol from erosion. This protection continues into the present.

APPENDIX II

Physical and chemical data for soils sampled during the course of this research.

SWAIM GRAVEL PIT EXPOSURE

Depth (cm)	Percentages Sand	C.Silt	M+F Silt	Clay	CFS*	pH	%OM
5-17	9	8	45	38	85	4.8	3.78
20-30	8	8	46	38	87	5.0	3.23
30-40	10	6	48	36	84	5.3	2.43
45-55	9	7	49	35	86	5.5	2.00
70-80	8	7	53	32	88	6.0	1.01
95-105	9	8	51	32	87	6.1	0.46
120-130	9	10	55	26	88	6.4	0.26
145-155	8	8	57	27	89	6.6	0.28
170-180	9	8	58	25	88	6.7	0.16
195-205	9	9	56	26	88	6.3	0.38
220-230	7	9	56	28	90	6.5	0.42
245-255	10	9	55	26	86	6.6	0.38
270-280	9	7	60	24	88	6.7	0.23
295-305	11	8	59	22	86	6.4	0.24
320-330	8	9	63	20	90	6.7	0.26
345-355	11	10	55	24	86	6.5	0.60
370-380	8	14	53	25	89	6.5	0.48
395-405	12	11	P(390)** 50	27	84	6.5	0.62
420-430	12	11	49	28	83	6.6	0.53
445-455	15	10	47	28	79	6.6	0.56
470-480	16	10	46	28	78	6.2	0.44
495-505	20	9	41	30	71	6.4	0.39
520-530	19	9	40	32	72	6.5	0.33
545-555	22	8	37	33	67	6.5	0.34
570-580	27	10	33	30	61	6.5	0.31

* Clay-free silt ** Top of paleosol based on color

TRANSECT 1 SITE 1

Depth (cm)	Percentages Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
0-15	14	10	P (0) 37	39	77	6.1	2.19
15-30	12	12	38	38	81	5.8	1.20
30-45	8	14	41	37	87	6.0	0.91
45-55	12	11	41	36	81	6.1	0.78
70-80	14	10	41	35	78	6.0	0.59
95-105	13	12	43	32	81	6.2	0.42
120-130	14	12	42	32	79	6.0	0.31
140-155	11	16	41	32	84	6.0	0.14
170-180	10	15	43	32	85	6.1	0.27
195-205	13	13	43	31	81	6.0	0.12
220-230	13	14	42	31	81	5.9	0.13
245-255	13	13	43	31	81	6.1	0.34

TRANSECT 1 SITE 3

0-15	13	8	45	34	80	4.8	3.26
15-30	13	9	42	36	80	5.8	2.43
40-55	14	8	P (35) 44	34	79	6.0	1.36
70-80	15	8	42	35	77	6.3	0.76
90-105	16	9	37	38	74	6.2	0.86
120-130	13	10	39	38	79	5.8	1.14
140-150	12	11	38	39	80	6.3	0.47
170-180	10	12	42	36	84	6.1	0.38
195-205	14	10	42	34	79	6.1	0.38
220-230	13	11	43	33	81	6.2	0.34
245-255	14	12	42	32	79	6.2	0.28
270-280	8	12	51	29	89	6.1	0.48
295-305	13	11	47	29	82	6.4	0.41

TRANSECT 1 SITE 5

0-15	10	8	52	30	86	5.6	1.64
15-30	12	11	50	27	84	6.3	0.97
40-55	11	10	47	32	84	6.4	0.84

TRANSECT 1 SITE 5 (cont)

Depth (cm)	Percentages: Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
70-80	7	11	50	32	90	6.6	0.75
95-105	11	12	46	31	84	6.4	0.67
120-130	13	11	P(130) 45	31	81	6.2	0.55
145-155	12	12	38	38	81	6.1	0.44
170-180	14	10	38	38	77	6.3	0.48
195-205	15	9	40	36	77	6.0	0.53
220-230	no data						
245-255	11	10	46	33	84	6.1	-
270-280	12	11	44	33	82	6.5	0.30
295-305	14	10	44	32	79	6.5	0.16
320-330	13	12	42	33	81	6.4	0.23

TRANSECT 1 SITE 7

Depth (cm)	Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
0-15	11	9	45	35	83	5.5	3.96
15-30	12	6	56	26	84	5.4	2.48
45-55	10	7	49	34	85	6.0	1.21
70-80	12	7	51	30	83	6.2	0.77
95-105	11	7	53	29	85	6.3	0.46
120-130	9	7	54	30	87	6.3	0.48
145-155	11	7	52	30	84	6.4	0.23
170-180	9	9	54	28	88	6.5	0.27
195-205	12	7	53	28	83	6.4	0.14
220-230	10	12	53	25	87	6.4	0.41
245-255	10	8	56	26	86	6.5	0.44
270-280	no data						
295-305	12	9	51	28	83	6.0	1.81
310-330	10	12	P(330) 49	29	86	6.6	0.58
345-355	no data						
370-380	no data						
400-410	14	11	42	33	79	6.2	0.96
420-435	12	11	42	35	82	6.4	0.57
435-450	14	9	37	40	77	6.4	0.45

TRANSECT 1 SITE 9

Depth (cm)	Percentages Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
0-15	8	11	43	38	87	5.0	2.98
15-30	10	12	44	34	85	5.4	-
45-55	13	8	44	35	80	5.5	1.88
70-80	15	9	44	32	78	6.1	1.13
95-105	15	10 P(100)	41	34	77	6.4	0.59
120-130	16	8	37	39	74	6.0	0.83
145-155	15	9	36	40	75	6.4	0.64
170-180	13	10	40	37	79	6.3	0.45
195-205	13	9	43	35	80	6.1	0.29
220-230	13	12	41	34	80	6.2	0.31
240-250	13	11	41	35	80	6.1	0.24

TRANSECT 2 SITE 1

0-15	17	10 P(0)	36	37	73	6.1	2.47
15-30	12	13	35	40	80	6.0	1.01
40-55	11	11	42	36	83	5.9	0.65
70-80	13	11	38	38	79	6.0	0.45
95-105	12	12	40	36	81	6.1	0.28
120-130	21	10	32	37	67	6.0	0.29
145-155	20	12	28	40	67	6.3	0.30
170-180	22	8	30	40	63	6.2	0.32
195-205	26	9	27	38	58	6.4	0.29
220-230	24	7	33	36	63	6.6	0.31
245-255	27	9	33	31	61	6.5	0.09
270-280	50	6	18	26	32	6.8	0.11

TRANSECT 2 SITE 3

0-15	16	10	44	30	77	5.7	2.70
15-30	16	10	44	30	63	6.0	2.34
45-55	22	7 P(40)	39	32	68	5.7	0.71
70-80	16	13	29	42	72	6.1	0.48

TRANSECT 2 SITE 3 (cont)

Depth (cm)	Percentages Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
95-105	19	11	29	41	68	6.6	0.36
120-130	24	8	30	38	61	6.7	0.31
145-155	17	9	30	44	70	6.9	0.26
170-180	26	6	27	41	56	6.8	0.13
195-205	27	7	27	39	56	6.8	0.26
220-230	33	9	25	33	51	6.8	0.13
245-255	27	9	30	34	59	6.6	0.25
TRANSECT 2 SITE 5							
0-15	16	7	47	30	77	5.2	3.54
15-30	15	9	46	30	79	5.8	2.73
45-55	12	10	49	29	83	5.6	1.35
70-80	16	9	49	26	78	6.1	0.83
95-105	16	10	48	26	78	6.1	0.55
120-130	17	11	40	32	75	6.1	0.45
145-155	23	8 P(150)	30	39	62	6.1	0.40
170-180	23	7	27	43	60	6.0	0.29
195-205	22	7	29	42	62	6.0	0.19
220-230	23	5	29	43	60	6.0	0.16
245-255	20	8	30	42	66	6.5	0.18
270-280	19	8	32	41	68	6.4	-
295-305	21	10	31	38	66	6.4	0.04
TRANSECT 2 SITE 7							
0-15	15	9	45	31	78	6.1	3.53
15-30	8	11	53	28	89	5.8	1.52
45-55	12	11	49	28	83	6.2	1.08
70-80	16	10	47	27	78	6.5	0.63
95-105	10	14	49	27	86	6.3	0.58
120-130	17	11 P(135)	46	26	77	6.4	0.38
145-155	18	12	41	29	75	6.4	0.29

TRANSECT 2 SITE 7 (cont)

	Percentages						
Depth (cm)	Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
170-180	22	7	32	39	64	6.2	0.33
195-205	22	7	27	44	61	6.2	0.33
220-230	22	8	27	43	61	6.3	0.18
245-255	22	9	27	42	62	6.4	0.17
270-280	11	19	32	38	82	6.4	0.03
295-305	26	8	32	34	61	6.5	0.08
320-350	22	8	34	36	66	6.7	0.08

TRANSECT 3 SITE 1

0-15	23	8 P(0)	31	38	63	5.9	1.17
15-30	17	11	31	41	71	6.7	0.54
45-55	17	10	35	38	73	7.0	0.21
70-80	22	8	32	38	65	7.2	0.22
95-105	25	8	31	36	61	7.2	0.16
120-130	26	9	29	36	59	6.8	0.29
145-155	31	10	25	34	53	7.4	0.14
170-180	38	6	23	33	43	7.3	0.14
195-205	39	5	21	35	40	7.1	0.16
220-230	19	5	28	48	63	7.1	0.12
240-255	5	5	44	46	91	6.8	0.00

TRANSECT 3 SITE 3

0-15	16	7	44	33	76	5.0	3.39
15-30	18	8	42	32	74	6.1	1.51
45-55	18	9 P(40)	43	30	74	6.4	0.90
70-80	23	8	37	32	66	6.4	0.54
95-105	23	9	32	36	64	6.4	0.46
125-135	29	8	25	38	53	6.4	0.39
145-155	28	7	25	40	53	6.6	0.45
170-180	30	5	27	38	52	6.7	0.23
195-205	31	7	24	38	50	6.8	0.22
220-230	32	3	28	37	49	6.7	0.00
245-255	36	4	24	36	44	6.9	0.09

TRANSECT 3 SITE 5

	Percentages							
Depth (cm)	Sand	C. Silt		M+F Silt	Clay	CFS	pH	%OM
0-15	17	9		42	32	75	5.0	3.49
15-30	19	7		41	33	72	5.5	2.89
40-55	16	10	P(40)	40	34	76	6.0	1.75
70-80	18	10		43	29	75	6.3	1.45
95-105	25	7		40	28	65	6.1	0.68
120-125	25	10		33	32	63	6.2	0.59
145-155	35	7		26	32	49	6.2	0.39
170-180	35	5		25	35	42	6.2	0.27
195-205	56	3		13	28	22	6.2	0.20
215-230	62	1		13	24	18	6.3	0.05
240-255	71	1		10	18	13	6.3	0.08
260-280	78	3		7	12	11	6.7	0.09
TRANSECT 3 SITE 7								
0-15	4	14		45	37	94	4.9	4.18
15-30	16	6		41	37	75	5.6	2.78
45-55	14	5		47	34	79	5.7	1.67
70-80	9	7		53	31	87	6.3	0.66
95-105	10	8		54	28	86	6.2	0.89
120-130	11	11		52	26	85	6.4	0.74
145-155	17	8	P(150)	50	25	77	6.5	0.52
170-180	21	8		46	25	72	6.7	0.52
195-205	27	10		40	23	65	6.7	0.48
220-230	29	8		37	26	61	6.5	0.33
245-255	36	5		27	32	47	6.5	0.32
270-280	52	2		10	36	19	6.2	0.13
295-305	55	1		11	33	18	6.5	0.16
320-330	52	3		13	32	24	6.4	0.07
340-360	35	6		26	33	48	6.6	0.04
TRANSECT 3 SITE 9								
0-15	15	6		42	37	76	4.9	3.47
15-30	10	7		48	35	85	5.7	1.74

TRANSECT 3 SITE 9 (cont)

Depth (cm)	Sand	C. Silt		M+F Silt	Clay	CFS	pH	%OM
	Percentages							
45-55	6	10		56	28	92	6.3	0.18
70-80	9	8		51	32	87	6.3	0.77
95-105	4	9		53	34	94	6.5	0.58
120-130	9	7		55	29	87	6.4	0.66
145-155	12	8		55	25	84	6.8	0.50
170-180	3	14		57	26	96	6.7	0.40
195-205	6	13		56	25	92	6.4	0.49
220-230	14	12		49	25	81	6.5	0.54
245-255	19	9		48	24	75	6.6	0.48
270-280	21	9		45	25	72	6.2	0.74
295-305	20	12		42	26	73	6.5	0.29
320-330	33	7	P(310)	33	27	55	6.4	0.17
340-355	52	3		9	36	19	6.3	0.21
370-380	47	4		15	34	29	6.4	0.09
395-405	64	3		8	25	15	6.4	0.05
420-430	49	5		16	30	30	6.5	0.10

TRANSECT 4 SITE 1

0-15	14	10	P(0)	42	34	79	5.4	2.51
15-30	17	10		44	29	76	6.1	1.10
45-55	17	12		41	30	76	6.4	0.81
70-80	14	14		38	34	79	6.2	0.74
95-105	15	14		35	36	77	6.2	0.65
120-130	22	13		32	33	67	6.4	0.65
145-155	28	9		31	32	59	6.1	0.42
170-180	22	11		32	35	66	6.1	0.35
195-205	31	8		27	34	53	6.4	0.19
220-230	41	7		23	29	42	6.1	0.17
240-250	45	8		20	27	38	6.2	0.19

TRANSECT 4 SITE 3

Depth (cm)	Sand	C. Silt		M+F Silt	Clay	CFS	pH	%OM
	Percentages							
0-15	12	7		43	38	81	5.3	3.48
15-30	10	7		46	37	84	5.8	1.64
45-55	4	12		46	38	94	6.1	1.20
70-80	11	8		50	31	84	6.0	0.90
95-105	9	8		55	28	88	6.3	0.58
120-130	12	8		54	26	84	6.5	0.62
145-155	15	9		49	27	79	6.6	0.58
170-180	11	11	P(175)	51	27	85	6.5	0.58
195-205	15	10		45	30	79	6.5	0.71
220-230	13	13		44	30	81	6.3	0.72
245-255	16	9		45	30	77	6.4	0.62
270-280	21	10		37	32	69	6.0	0.55
290-305	27	7		31	35	58	6.3	0.50
320-335	47	5		18	30	33	6.4	0.20
340-360	60	5		11	24	21	6.1	0.15
370-390	65	4		11	20	19	6.4	0.35

TRANSECT 4 SITE 5

0-15	14	7		41	38	77	5.2	4.31
15-30	10	5		45	40	83	5.8	2.97
45-55	9	5		49	37	86	5.8	1.47
70-80	10	7		52	31	86	6.3	1.08
95-105	9	8		53	30	87	6.7	0.62
120-130	9	7		52	32	87	6.3	0.52
145-155	10	8		51	31	86	6.4	0.55
170-180	14	7		48	31	80	6.2	0.42
195-205	6	9		55	30	91	6.5	0.51
220-230	8	12		54	26	89	6.4	0.51
245-255	8	10		56	26	89	6.6	0.59
270-280	7	13		54	26	91	6.5	0.69
295-305	12	10	P(290)	49	29	83	6.6	0.79

TRANSECT 4 SITE 5 (cont)

Depth (cm)	Percentages Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
320-330	15	9	46	30	79	6.3	0.64
345-355	10	12	48	30	86	6.1	0.65
370-380	16	9	45	30	77	6.2	0.60
395-405	18	10	39	33	73	6.3	0.59
420-430	18	9	37	36	72	6.4	0.55
445-455	17	10	40	33	75	6.5	0.46
470-480	20	9	34	37	68	6.3	0.54
495-505	16	10	38	36	75	6.6	-

TRANSECT 4 SITE 7

Depth (cm)	Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
0-15	9	11	52	28	88	5.2	2.84
15-30	9	8	52	31	87	6.1	1.67
45-55	8	11	42	39	87	6.3	1.09
70-80	13	9 P(85)	50	28	82	6.5	0.69
95-105	10	14	46	30	86	6.5	0.70
120-130	9	15	45	31	87	6.3	0.70
145-155	11	15	41	33	84	6.3	0.60
170-180	20	12	34	34	70	6.3	0.57
195-205	25	10	31	34	62	6.2	0.62
220-230	18	14	30	38	71	6.2	0.40
245-255	30	8	31	31	57	6.2	0.52
270-280	28	10	35	27	62	6.5	0.53
295-305	30	10	32	28	58	6.3	0.45
320-330	21	13	33	33	69	6.3	0.41
340-360	34	5	29	32	51	6.3	0.50

TRANSECT 5 SITE 1

Depth (cm)	Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
0-15	11	7	47	35	83	4.6	2.94
15-30	4	12	46	38	94	5.3	1.86
45-55	-	-	-	-	-	6.4	-
70-80	7	8	53	32	90	6.2	0.63
95-105	6	9	56	29	92	6.3	0.43

TRANSECT 5 SITE 1 (cont)

Depth (cm)	Percentages Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
120-130	8	9	57	26	89	6.2	0.49
145-155	10	9	57	24	87	6.2	0.49
170-180	12	9	54	25	84	6.4	0.35
195-205	9	14	49	28	88	6.4	0.23
220-230	16	14 P(215)	44	26	78	6.4	0.34
245-255	8	9	48	25	88	6.1	-
270-280	22	10	41	27	70	6.5	0.36
295-305	22	10	39	29	69	6.4	0.45
320-330	19	10	44	27	74	6.2	0.45
345-355	21	12	30	37	67	6.5	0.29
370-380	30	8	26	36	53	6.4	0.17
390-405	32	5	28	35	51	6.5	0.18
420-430	29	6	29	36	55	6.6	0.14
440-450	32	6	28	34	52	6.6	0.04

TRANSECT 5 SITE 3

0-15	8	12	54	26	89	4.8	2.02
15-30	5	10	51	34	92	6.1	1.10
45-55	11	8	53	28	85	6.1	0.66
70-80	4	13	57	26	95	6.2	0.52
95-105	12	8	54	26	84	6.2	0.42
120-130	13	11	50	26	82	6.2	0.28
145-155	10	14 P(160)	47	29	86	6.4	0.52
170-180	10	14	47	29	86	6.3	0.41
195-205	18	8	47	27	75	6.4	0.40
220-230	13	14	43	30	81	6.3	0.36
245-255	25	9	34	32	63	6.1	0.14
270-280	35	4	26	35	46	6.1	0.19
295-305	29	7	25	39	52	6.3	0.19
320-330	31	5	27	37	51	6.4	0.15

TRANSECT 5 SITE 5

Depth (cm)	Percentages Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
0-15	10	10	46	34	85	4.7	2.90
15-30	9	7	50	34	86	5.1	1.85
45-55	9	9	52	30	87	5.9	0.64
70-80	10	10	50	30	86	6.1	0.43
95-105	14	11	P(100) 48	27	81	6.0	0.41
120-130	10	14	48	28	86	6.1	0.30
145-155	18	12	40	30	74	6.2	0.44
170-180	18	11	41	30	74	5.9	0.33
195-205	26	8	31	35	60	6.3	0.47
220-230	26	9	27	38	58	6.0	0.27
245-255	30	6	24	40	50	6.4	0.22
270-280	18	2	30	50	64	6.6	0.28
295-305	8	3	31	58	81	7.3	0.04

TRANSECT 5 SITE 7

0-15	14	9	41	36	78	5.3	2.44
15-30	6	12	40	42	90	6.4	1.86
45-55	8	11	41	40	87	5.7	1.48
70-80	9	10	46	35	86	5.8	1.24
95-105	13	8	P(100) 47	32	81	6.2	0.61
120-130	16	9	43	32	76	6.0	0.65
145-155	10	13	43	34	85	6.2	0.41
170-180	14	10	37	39	77	6.1	0.74
195-205	21	5	28	46	61	6.0	0.39
220-230	17	4	30	49	67	6.2	0.52
245-255	14	2	38	46	74	6.5	0.29

TRANSECT 6 SITE 1

0-15	11	8	49	32	84	4.6	3.20
15-30	6	12	45	37	90	5.2	2.44
45-55	10	9	43	38	84	5.8	1.71

TRANSECT 6 SITE 1 (cont)

	Percentages						
Depth (cm)	Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
70-80	7	11	47	35	89	6.0	1.11
95-105	4	12	48	36	94	6.3	0.78
120-130	8	10	52	30	89	5.9	0.52
145-155	15	13	44	28	79	5.9	0.35
170-180	8	10	55	27	89	6.1	0.32
195-205	9	9	56	26	88	6.3	0.35
220-230	7	13	52	28	90	6.2	0.36
245-255	12	7	53	28	83	6.1	0.62
270-280	11	10	55	24	86	6.4	0.40
295-305	9	9	58	24	88	6.6	0.53
320-330	13	9	56	22	83	6.4	0.43
345-355	10	12	57	21	87	6.4	0.46
370-380	13	13	53	21	84	6.6	0.57
395-405	12	15	P(400) 48	25	84	6.5	0.54
420-430	16	11	46	27	78	6.6	0.61
445-455	no data						
470-480	20	10	42	28	72	6.6	0.59
495-505	20	8	43	29	72	6.7	0.60
520-530	26	13	29	32	62	6.7	0.39
545-555	39	6	28	27	47	6.7	0.27

TRANSECT 7 SITE 1

0-15	11	6	45	38	82	4.8	3.24
15-30	11	7	44	38	82	5.1	1.68
40-55	8	9	49	34	88	6.3	0.95
70-80	9	9	50	32	87	6.3	0.71
95-105	4	14	51	31	94	6.3	0.57
120-130	7	12	55	26	96	6.5	0.60
145-155	13	9	53	25	83	6.6	0.63
170-180	15	10	49	26	80	6.4	0.96
195-205	12	12	47	29	83	6.5	0.68

TRANSECT 7 SITE 1 (cont)

Depth (cm)	Percentages Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
220-230	12	11	P(215) 47	30	83	6.6	0.75
245-255	9	12	49	30	87	6.6	0.80
270-280	16	9	41	34	76	6.3	0.57

TRANSECT 7 SITE 3

Depth (cm)	Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
0-15	10	7	45	38	84	4.9	3.16
10-20	6	12	42	40	93	5.2	2.34
25-35	9	9	47	35	86	5.7	1.29
45-55	6	12	50	32	91	6.2	0.74
70-80	4	11	50	35	94	6.3	0.85
95-105	8	9	57	26	89	6.4	0.38
120-130	12	8	54	26	84	6.6	0.69
145-155	11	11	51	27	85	6.4	0.76
170-180	10	12	48	30	86	6.2	0.79
195-205	12	9	P(205) 49	30	83	6.3	0.69
220-230	15	15	37	33	78	6.4	0.54
240-255	10	16	38	36	84	6.3	0.47
270-280	18	10	37	35	72	6.4	0.31
295-305	17	9	36	38	73	6.3	0.46
320-330	14	14	36	36	78	-	0.42
345-355	15	11	39	35	77	6.3	0.25
370-380	14	10	43	33	79	6.3	0.29
390-410	14	12	41	33	79	6.4	0.09
440-455	15	11	38	36	77	6.3	0.38
480-490	17	9	40	34	74	6.4	0.08

TRANSECT 8 SITE 1

Depth (cm)	Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
0-15	17	12	P(0) 44	27	77	6.3	2.86
15-30	13	9	40	38	79	5.3	1.10
40-60	15	9	41	35	77	5.4	0.71
70-80	16	11	37	36	75	5.4	0.51

Depth (cm)	Percentages					pH	%OM
	Sand	C. Silt	M+F Silt	Clay	CFS		
TRANSECT 8 SITE 1 (cont)							
95-105	10	14	41	35	85	5.5	0.51
120-130	12	12	40	36	81	5.7	0.36
145-155	9	13	44	34	86	5.7	0.36
170-180	16	12	38	34	76	5.7	0.27
195-205	24	10	33	33	64	5.7	0.75
220-230	24	8	33	35	63	5.6	0.21
235-250	18	12	37	33	73	5.8	0.15
TRANSECT 8 SITE 3							
0-15	15	11 P(5)	42	32	78	6.2	3.21
15-30	15	8	39	38	76	5.5	1.35
45-55	14	9	39	38	77	5.6	1.10
70-80	13	10	41	36	80	5.5	0.80
95-105	10	10	45	35	85	5.8	0.43
120-130	7	15	43	35	89	5.4	0.53
145-155	15	11	44	30	79	5.6	0.38
170-180	10	16	43	31	86	5.6	0.38
195-205	13	12	45	30	81	5.8	0.22
220-230	11	16	45	28	85	5.9	0.36
240-250	9	16	47	28	88	5.7	0.21
TRANSECT 8 SITE 5							
0-15	16	10	49	25	79	6.0	4.34
15-30	13	12 P(35)	45	30	81	6.1	2.18
45-55	12	10	44	34	82	5.9	1.45
70-80	14	9	45	32	79	5.5	1.04
95-105	15	11	42	32	78	5.7	0.85
120-130	11	10	46	33	84	5.9	0.74
145-155	16	9	37	38	74	6.0	0.52
170-180	13	9	40	38	79	6.0	0.53
195-205	13	9	42	36	80	6.2	0.44

TRANSECT 8 SITE 5 (cont)

	Percentages						
Depth (cm)	Sand	C. Silt	M+F Silt	Clay	CFS	pH	%OM
220-230	10	10	46	34	85	6.2	0.44
235-250	10	12	43	35	85	6.2	0.32

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0-15	14	8	48	30	80	6.5	6.12
15-30	8	11	43	38	87	6.2	2.37
45-55	12	6	55	37	84	5.1	1.27
70-80	8	8 P(65)*	50	34	88	5.2	-
95-105	8	7	53	32	88	5.5	0.65
120-130	10	10	53	27	86	5.7	0.50

*This paleosol at this site had a classic gumbotil profile. Because of its great differences in morphology from other paleosols in the area, data from the above site were not included in numerical analyses in the body of the thesis.