

THE SOILS AND GEOMORPHOLOGY OF THE EAST BRANCH WALNUT
VALLEY: CONTEXTS OF HUMAN ADAPTATION
IN THE KANSAS FLINT HILLS

by

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by

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ABSTRACT

In this study, the alluvial soils and landforms of the East Branch Walnut valley, in the southern Flint Hills of Kansas, are described and their implications for paleoenvironmental change are discussed. Using chronological and pedological evidence obtained from archaeological sites, two soil-stratigraphic units are defined, and the dominant pedogenic processes which formed each are identified. Pedogenesis of the Nuttal Soil, formed between 1500-1000 B.P. and the present, was dominated by horizonation processes, which produced a moderately-developed soil. Pedogenesis of the Snyder Soil, formed between 4000 and 2000 B.P., was dominated by haploidization processes, which maintained a weakly-developed soil. The different pedogenetic trends manifested by the soils are explained in terms of the different geomorphic contexts in which each formed. The Nuttal Soil developed beneath geomorphically stable, well-drained levee and colluvial footslopes, while the Snyder Soil formed in slowly aggrading, less well-drained floodbasin and back-water areas.

The Nuttal and Snyder Soils contribute to the formulation of an alluvial chronology for the East Branch Walnut valley. Valley-wide aggradation, in progress at ca. 4500 B.P., slowed after 4000 B.P., permitting formation of the Snyder Soil. At ca. 2000 B.P., study area streams entrenched. Levees and colluvial footslopes were deposited between 2000 and 1500-1000 B.P., locally burying the Snyder Soil. The Nuttal Soil subsequently formed in these deposits.

This sequence of geomorphic changes is interpreted as evidence for changes in flood magnitude and frequency, perhaps induced by shifts in precipitation patterns. This interpretation is supported by an analysis of the close interrelationships of geomorphic, hydrologic and climatic processes in the present-day Flint Hills. In addition, major geomorphic discontinuities like that documented at 2000 B.P. in the East Branch Walnut valley were widespread in the southeastern Central Plains between ca. 2000 and 1000 B.P., further supporting the inference that geomorphic changes resulted from regional-scale climatic shifts.

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CHAPTER 1

INTRODUCTION

Two thousand years ago, the East Branch of the Walnut River, in the southern Flint Hills of Kansas, underwent a relatively rapid shift from gradual, basin-wide aggradation to channel entrenchment. By at least 1500-1000 years ago, the entrenched channel stabilized, and has remained relatively stable to the present. This thesis presents soil-geomorphic evidence for these events, and proposes a model which accounts for the shifts between episodes in terms of climatically-induced hydrologic changes in the drainage basin.⁴ The sequence of environmental changes identified in the East Branch Walnut valley contributes to an emerging regional pattern of environmental change on the southeastern periphery of the Central Plains.

The first chapter of this thesis has two objectives: 1) to discuss the contribution of archaeological data to the central problem of the thesis, and to suggest the significance of the study as a contribution to archaeological research; and 2) to develop an analytical approach which suggests the manner in which the properties of alluvial soils and landforms reflect the influence of environment.

ARCHAEOLOGICAL CONTRIBUTIONS

Although this thesis deals primarily with pedologic, geomorphic, hydrologic and climatic phenomena, archaeology both contributes to and benefits from the study. Archaeology provides methodologies for the stratigraphic recovery and chronological ordering of artifacts. Temporally-

diagnostic artifacts (e.g., ceramics and projectile points) and radio-carbon dates from archaeological sites are used to define and correlate stratigraphic units in the East Branch Walnut valley. In addition, vertical and horizontal patterns observed in artifact distributions at several sites are used to infer geomorphic and pedologic features not readily discerned in fine-grained alluvial deposits, such as the configuration of former land surfaces, rates of deposition and the extent of pedoturbation.

In return for the largely technical contributions noted above, this study provides information relevant to a variety of archaeological problems. For instance, the discussion of soil phosphorus in Chapter 4 suggests implications for bone preservation at sites 14BU9 and 14BU25. Discussion of surface drainage in relation to landforms in Chapter 5 has implications for prehistoric site selection. However, these and similar results, summarized in Chapter 7, are largely "spin-offs" from the study's main goal of environmental reconstruction.

The primary contribution of this thesis to the archaeology of the East Branch Walnut valley is the model of environmental change developed in Chapter 6. During the 4500 year time span embraced by this model, the East Branch Walnut valley was inhabited by human groups who subsisted primarily by hunting and gathering. It is assumed that these groups were adapted to prevailing environmental conditions and that a knowledge of environmental change is required to properly understand these adaptations.

Butzer (1971:3-10) emphasizes the importance of interdisciplinary research in studying the interactions of prehistoric humans and their

environments. The objective of this research, he states, is "environmental reconstruction as applied to an understanding of the ecological setting to prehistory" (Butzer 1971:10). As stated by Clarke (1957:20):

What the prehistoric archaeologist has to study is the history of human settlement in relation to the history of the climate, topography, vegetation and fauna of the territory in question...it is only by observing the human cultures of antiquity as elements in a changing ecological situation that it is possible to form a clear idea of even the economic basis of early settlement....

The holistic, ecological approach to prehistory envisioned by Clarke requires interdisciplinary cooperation, integrating archaeological concepts and techniques with those of the biological and earth sciences (Butzer 1971:401). "Geoarchaeology", the study of the geomorphic context of archaeological sites and artifacts (Renfrew 1976; Galdfelter 1977) is a "vital facet" of this ecological approach (Butzer 1977:126).

The development of a model of the prehistoric human ecology of the East Branch Walnut valley is beyond the scope of this thesis. However, the study provides a first step toward such a model, by attempting to infer the changing environmental context within which prehistoric adaptations occurred.

ANALYTICAL APPROACH

The remainder of this chapter develops an analytical approach for recognizing the paleoenvironmental implications of the alluvial soils and landforms of the East Branch Walnut valley. The initial step in this approach is to realize, along with Bryson and Wendland (1967:271), that:

Evidences of past climate do not appear as readings of a thermometer, a rain-gauge, or a wind-vane, but rather as natural phenomena which represent, usually, complexes of climatic parameters acting on soils, topography or biota....

The study of climatic change therefore involves the identification and interpretation of "proxies" or "surrogates", preserved manifestations of phenomena which are influenced by climate. Soils and landforms are the principle surrogates used in this study, although studies of past vegetation and land snails are cited in support of these lines of evidence. Since all these phenomena are influenced by climate and leave preserved evidence of that influence, all are suitable surrogates. However, it is imperative to realize that the records of these various phenomena do not directly reflect climatic change, but rather reflect responses of each phenomenon to climatic change.

This point is especially significant when using soils or landforms as climatic indicators, since each of these phenomena responds to a variety of environmental systems in addition to climate. An attempt to derive the paleoenvironmental implications of soils and landforms must carefully consider the interaction of numerous environmental processes and parameters. It is necessary to determine two things: 1) the processes by which each phenomenon responds to environment; and 2) the manner in which this response is reflected in preserved records of the phenomenon.

The following sections discuss how soils, in general, and alluvial soils, in particular, are influenced by environment and how this influence is reflected in soil properties. The influence of environment on soil is considered from three perspectives:

1. Pedologic: Soil properties are viewed as products of the complex interaction of pedogenic processes. These processes are responsive to environmental parameters, which are commonly summarized as the five "fundamental soil-forming factors" of climate, organisms, parent material, relief and time (Jenny 1941; Crocker 1952).

2. Soil-stratigraphic: For soil formation to occur, surface stability and time are required. The presence of a soil, therefore, indicates a period of relative landscape stability, and is useful in stratigraphic correlation and in reconstructing the geomorphic history of a locality.

3. Soil-geomorphic: This perspective combines pedologic and soil-stratigraphic evidence to define lateral variation in soil properties across a landscape, and to account for this variation in terms of geomorphic processes.

The following sections discuss each of these perspectives.

Pedologic Perspective

Pedogenesis is that combination of biological, chemical and physical processes by which mineral and organic materials are transformed into the distinct layers, or horizons, of soil. These processes create the morphological, chemical and physical properties of soil, which can be described and measured.

According to Yaalon (1971:29):

A soil-forming processes can best be described as a combination of a multitude of individual reactions, taking place in a soil or parent material, acting together or consecutively, and producing a given soil or soil feature.

Pedogenic processes act on soil by adding, removing, transforming and transferring soil substances (Simonson 1959:153). The specific processes operating in the soil, and the intensity with which they act, vary among soils, and with time in the same soil (Simonson 1959:155; Nikiforoff 1949:220).

Environmental variables, external to the soil itself, importantly influence the kinds of pedogenic processes at work in a soil, and the rates and intensities at which those processes operate. Pedologists often refer to these environmental variables as "factors" of soil formation. As defined by Buol et al. (1980:105);

A factor of soil formation is an agent, force, condition, or relationship, or combination of these, which influences, has influenced or may influence a parent material of a soil, with the potential of changing it.

Pedologists commonly group the innumerable factors influencing pedogenesis into a set of five fundamental soil forming factors: climate, organisms, parent material, relief and time (cf. Jenny 1941; Crocker 1952).

Soil properties are not the direct products of external environmental factors. Rather, environmental factors influence the kinds, rates and intensities of pedogenic processes. Pedogenic processes, in turn, produce the observable and measurable properties of soils. The paleopedologist, interested in inferring the characteristics of past environments from soil properties, must first relate soil properties to the pedogenic processes responsible for their formation, and then must infer from this set of processes the environmental factors to which the processes were responsive.

However, the relationships of soil properties, pedogenic processes and environmental factors are rarely, if ever, direct and linear. Soils with similar profiles may develop from entirely different combinations of processes and environmental variables. Changes over time in environmental factors inducing changes in pedogenic processes, may alter properties from their original state. The following examples illustrate these considerations.

Different combinations of pedogenic processes may produce similar soil properties. For instance, the formation of textural B horizons (B horizons characterized by distinctly clayier textures, relative to overlying and underlying horizons) is conceivably attributed to 3 processes: 1) differential rates of weathering, so that clays form faster in the B horizon; 2) translocation of clay from overlying horizons, accumulating in the B horizon; or 3) non-uniform textures of sedimentary layers, so that clayey B horizon textures are inherited from characteristics of the original parent material (Chittleborough and Oades 1979:4). These processes are often difficult to distinguish from pedological evidence alone, as discussed in Chapter 4 with reference to the B horizons of certain East Branch Walnut soils.

Environmental factors influence not only the kind, but also the duration and intensity of pedogenic processes, affecting the degree of expression of soil properties. For instance, a major debate in the interpretation of intensely weathered, Late Pleistocene paleosols of the midwestern United States involves the difficulty of distinguishing the effects of intense weathering under a warmer or more humid climate from weathering due to prolonged exposure under less intense climate

(Frye and Leonard 1952:23; Simonson 1954:721-723; Morrison 1967:44-46; Ruhe 1974). As stated by Simonson (1954:722), "The product of a long interval of time and a low climate intensity could be the same as the product of a short interval of time and high climatic intensity."

Polygenetic soils, or soils formed under the influence of changing environments (Birkeland 1974:129), reflect both past and present processes. Present-day processes impose new properties on the soil and alter properties inherited from an earlier set of processes. For example, according to Gile et al. (1966:357-358), K horizons (horizons cemented by calcium carbonate) formed during the moist Late Pleistocene in New Mexico and mark the depth to which soil water normally infiltrated before precipitating dissolved carbonates. At present, these horizons persist, while pedogenesis under drier Holocene climates (with a correspondingly decreased depth of leaching) has created carbonate accumulations in the overlying horizons.

Burial of soils also alters pedogenic processes and properties (Nikiforoff 1949:227; Gerasimov 1971:19-20). Many properties of buried soils, especially chemical ones, are subject to "secondary enrichment" from the overlying soil (Ruhe 1956; Ruhe and Olson 1980). For example, calcium carbonate concretions in the upper portions of Yarmouth-Sangamon paleosols in Iowa probably represent leaching from the overlying, calcareous Wisconsin loess (Ruhe 1956). Given sufficient time, the properties of a buried paleosol may be altered to the extent that it is morphologically indistinguishable from the overlying soil (Ruhe and Olson 1980).

Because of the complexity of the relationships among soil properties, pedogenic processes and environmental factors, several alternative "genetic pathways" can generally be devised to account for the properties of a single soil (Arnold 1965). The problem is accentuated when dealing with polygenetic soils formed under several different environmental regimes. It is generally the case that ancillary lines of evidence are required to evaluate the alternatives suggested by the evidence of soils (Birkeland 1974).

The relationships of soil properties, pedogenic processes and environmental factors are summarized as follows:

1. The morphological, physical and chemical properties of a soil are the products of the operation and interaction of pedogenic processes.
2. Pedogenic processes operate in response to environmental factors including climate, organisms, parent material, relief and time.
3. The relationships of property, process and the external environment are rarely, if ever, direct and linear. A number of alternative processes, operating with varying duration and intensity over time, may provide equally plausible explanations for properties observed in a single soil. In attempting to infer past environmental parameters from soil properties, the effect of polygenetic modifications of properties must be considered.

Despite the complex interaction of pedogenic processes, most processes tend to promote one or two generalized trends, which Buol et al. (1980:95) term "horizonation" and "haploidization". Horizonation processes promote the differentiation of parent material into soil horizons. Haploidization processes inhibit horizonation, maintaining simple ("haploid") soil profiles (Buol et al. 1980:95). The properties of a haploid soil vary little with depth and are thus characterized as "isotropic".

Horizonation differentiates properties among horizons, creating "anisotropic" distributions (Hole 1961). In actuality, horizonation and haploidization are overlapping trends (Buol et al. 1980) which form a continuum from extreme anisotropism to extreme isotropism.

Birkeland (1974:23-24) offers a qualitative scheme for describing the degree of horizonation observed in a soil. A weakly developed soil exhibits an A horizon, indicating the incorporation of organic matter, but lacks a B horizon. Moderately developed soils exhibit B horizons. In sufficiently intensive leaching environments, A2 horizons may form. A2 and B horizons form by the chemical and physical weathering and translocation of soil substances. In strongly developed soils, the maximum differentiation of horizons in terms of color, thickness, texture and structure is achieved.

Soil-Stratigraphic Perspective

Soils serve as stratigraphic markers, delimiting boundaries between sedimentary deposits and indicating the position of erosional and depositional surfaces. In bodies of fine-grained, homogeneous sediments, soils may provide one of the few means of subdividing the stratigraphic record into discrete units of sediment and intervals of time.

To operate with sufficient intensity to form a soil, pedogenic processes must dominate, or at least balance out, geomorphic processes of erosion and deposition (Ruhe 1969a:398). Pedogenesis is inhibited if sediment is eroded or deposited more rapidly than it can be incorporated into, or maintained as, soil by pedogenic processes (Bos and Sevink 1975:223). A time factor is also involved, since pedogenic processes

must maintain dominance or equilibrium relative to geomorphic processes for a period sufficient for soil formation to occur (Butler 1959:5).

Landscapes develop through the alternation of episodes of deposition, erosion and stability. Evidence of depositional and erosional events is preserved in the stratigraphic record as accumulations and truncations, respectively, of sedimentary deposits. These events create surfaces. A surface, defined as a two-dimensional form having length and width but no thickness, may be depositional, representing the top of a depositional unit, or erosional, representing the removal of sediment from a previously deposited unit. Many surfaces possess both depositional and erosional segments (Daniels et al. 1971:56). For example, the erosion of a hillslope may produce an erosion surface on the slope itself and a colluvial depositional surface at the foot of the slope. Soils form beneath surfaces that are not subject to active deposition or erosion.

Butler (1959, 1967) characterized the process of landscape development as an alternation between unstable phases, characterized by erosion and deposition, and stable phases, characterized by soil formation. He defined the combinations of an unstable phase and the subsequent stable phase as a "soil-cycle" or "K-cycle" (from the Greek khronos, meaning "time"; Butler 1959:5). Specifically, a cycle consists of:

The interval of time covering the formation, by erosion and/or deposition of a new landscape surface, the period of development of the soils on that surface, and ending with the renewal of erosion and/or deposition on that surface (Butler 1959:7).

Although tectonic factors, base level changes and the exceeding of internal geomorphic thresholds may account for the destabilization of

landscape systems, much recent work deals with the importance of climatically-induced geomorphic and hydrologic change as a primary impetus for this phenomenon (Knox 1972, 1977; Baker and P.-Orellana 1977). This subject is dealt with in Chapter 6. For present purposes, it is sufficient to note that soils provide a line of evidence for establishing the existence and relative chronological order of events which reflect the episodic stabilization and destabilization of the landscape.

Soil-Geomorphic Perspective

Pedogenic processes promote the differentiation of soil properties in a vertical dimension, creating the horizons of a soil profile. Equally important, however, are lateral variations of soil properties across the landscape. A landscape is comprised of contiguous landforms, which are the products of erosion and deposition (Ruhe 1969b:4-5). A portion of the landscape which can be shown to have formed during a specified period of time and which has mappable geographic boundaries is termed a "geomorphic surface" (Ruhe 1969b:546; Daniels et al. 1971:461). Examples of geomorphic surfaces in an alluvial landscape are stream terraces. Each terrace, representing the remnant of a former floodplain surface of the stream (Leopold et al. 1964:461), is mappable and the time of its formation can be specified, at least relative to terrace surfaces above or below it.

Studies of geomorphic surfaces and the soils beneath them suggest that profile anisotropism increases with the relative age of the surface. Ruhe (1956), Parsons et al. (1970), Gile (1977) and Daniels and Gamble

(1978), to name only a few major studies, demonstrate relationships between age of surface and degree of soil development.

In addition to surface age, pedogenesis is also influenced by the geomorphic stability of surfaces. Walker and Coventry (1976) combine the concepts of profile anisotropy, surface stability and surface age in discussing pedogenesis in Australian alluvial landscapes. The resultant five-stage model (Table 1) clearly shows the interaction of pedogenic and geomorphic processes. Geomorphic processes dominate the stratic stage, which consists of stratified deposits, relatively unmodified by pedogenesis. These deposits form low benches immediately adjacent to stream channels, and are subject to erosion and deposition during floods which occur two to three times per year. In the cumulic stage, geomorphic and pedogenic processes are in relative equilibrium. Soils of this stage occupy floodplains which are flooded once every one to three years. Deposition and erosion occur in sufficiently small increments and at sufficiently infrequent intervals that pedogenic processes, although inhibited, are not totally interrupted. The cumulative A horizons (Riecken and Poetsch 1960; Nikiforoff 1949) developed under these conditions are adjusted to periodic increments of sediment, and tend toward haploidization processes.

In the latter three stages of the model, pedogenic processes dominate geomorphic processes. Soils of these stages occur on successively higher terraces above the floodplain. The lowest terraces, on which soils of the low-contrast solum stage occur, are infrequently flooded, while the higher terraces, occupied by the high-contrast solum and extended subsolum stages, are seldom, if ever, flooded. Horizonation

Table 1. Five-stage model of pedogenesis in alluvial landscapes (adapted from Walker and Coventry 1976:307-309).

Profile Stage	Geomorphic Context	Soil Properties
Stratic stage	Low alluvial benches, flooded 2 to 3 times per year.	Stratified alluvium; biotic and pedogenic influences minimal.
Cumulic stage	Floodplains; floods occur once every 1 to 3 years.	Thick, cumulative A horizons; dominated by biotic churning.
Low-contrast solum stage	Low terraces; floods infrequent, recurring at intervals greater	A horizons; minimal cambic or argillic B horizons; clay illuviation minor.
High-contrast solum stage	High terraces; not (or seldom) flooded; may be in erosional phase.	A, B and C horizons clearly defined; mineral weathering pronounced; clay illuviation often evident.
Extended sub-solum stage	High terraces; not (or seldom) flooded; often erosional landscapes.	A, B and C horizons strongly differentiated; intensely weathered solum and subsolum; clay illuviation often evident; subsolum cemented with Fe and Si.

processes represent the dominant trend in pedogenesis at these stages. The extent of horizonation is controlled primarily by the relative age of the surfaces (Walker and Coventry 1976:313).

Walker and Coventry (1976:312) view flood magnitude and frequency as critical parameters influencing pedogenesis in alluvial landscapes, noting that, "The higher the level of the alluvial surface, the greater the magnitude of the flood required to over top it and the lower the flood frequency of such an event." The high frequency of flooding at the lower levels of the landscape inhibits pedogenesis. At higher levels, flooding is not a limiting factor and pedogenesis proceeds unimpeded. Given sustained geomorphic stability, the extent of horizonation is controlled primarily by surface age.

While Walker and Coventry (1976) discuss differences in soil properties among geomorphic surfaces of different ages and landscape positions, they do not consider parameters which promote differential pedogenesis on a single surface, such as moisture regime.

Soil moisture regime, although ultimately related to climate and vegetation parameters such as precipitation and evapo-transpiration, is also influenced by geomorphic parameters. The geometry of a landform (whether concave or convex, level or sloping) influences the rate at which water runs off or infiltrates the soil. Textural and structural properties of the soil beneath the landform further influence permeability (Soil Survey Staff 1951:158-172). The internal and external drainage conditions of a soil influence the kind and intensity of pedogenic processes. Rapid runoff, given proper vegetation and soil-textural conditions, promotes erosion, and inhibits pedogenesis. Portions of the

soil profile subject to frequent or continual saturation in the presence of high water tables, exhibit the characteristic gray or bluish colors or mottling associated with such conditions (Birkeland 1974:118; Buol et al. 1980:129). In addition, internal drainage conditions influence the mobilization and translocation of soil substances. Substances which can be dissolved or suspended in soil water, such as chemical constituents and fine clays, are available for translocation, both vertically and laterally in the soil. Translocation is one of the major processes by which horizonation develops in soil (Simonson 1959); conditions which impede the movement of soil water may thus impede horizonation (Runge 1973; Smeck and Runge 1973).

Internal and external drainage of soils on a given geomorphic surface may be inherited from topographic features formed at the time the surface developed. Daniels and Gamble (1978), Parsons and Balster (1966) and Parsons (1979) discuss the influence of paleotopography and soil-stratigraphy on textural properties and drainage parameters. Similar variation in drainage characteristics are also inherent in the nature of alluvial deposition.

A geomorphic surface formed by fluvial processes, such as a floodplain, represents a "depositional sedimentary environment," defined by Reineck and Singh (1976:4) as a "geomorphic unit in which deposition takes place." Depositional sedimentary environments are subdivided into "subenvironments." The deposits of each subenvironment are characterized by a particular set of sedimentological and morphological properties, reflecting the hydrodynamic conditions of deposition (Reineck and Singh

1976:230). The subenvironments of the fluvial environment include channels, levees and floodbasins (Reineck and Singh 1976:230). The surface geometry and drainage characteristics of each subenvironment are often sufficiently distinct that soils developed within them exhibit differential profile development. For example, channel deposits, such as point bars, may retain evidence of sedimentary structures (the stratic stage of pedogenesis; see Table 1), although pedogenesis may eventually obscure such features. Levees, developed at the edge of channels, are higher in elevation and more convex than the lower-lying, more level floodbasins. The better surface drainage of levee surfaces may produce slightly more anisotropic soils, relative to the more poorly drained floodbasins. Pedogenic differences among the various subenvironments often persist after the floodplain has been abandoned and is no longer subject to erosion or deposition.

To summarize, numerous geomorphic parameters and processes influence pedogenesis in alluvial landscapes. Flooding and poor internal drainage inhibit horizonation. In the absence of these constraints, horizonation proceeds and, if not influenced by subsequent erosion or deposition, will be limited primarily by the age of the surface. Lateral variability in soil properties beneath a single geomorphic surface often reflects moisture regimes associated with different landforms or depositional subenvironments. In an alluvial landscape, soil properties may aid in the identification of relict floodplain subenvironments preserved on terrace surfaces.

Analytical Approach

Having examined pedogenesis from pedologic, soil-stratigraphic and soil-geomorphic perspectives, it is concluded that alluvial soils are profoundly influenced by geomorphic variables, among which are surface stability, surface drainage and surface age. Surface stability is prerequisite to soil formation. Pedogenesis will not occur unless the geomorphic processes of deposition and erosion are at least balanced by pedogenic processes. Poor surface drainage, reflecting landform geometry and sedimentological characteristics, also impedes pedogenesis. On geomorphically stable, well-drained surfaces, pedogenesis is limited primarily by surface age.

The influence of geomorphic variables on pedogenic processes is reflected by soil properties. The dominant trend in pedogenesis at a given landscape position is suggested by the relative anisotropy (i.e., differentiation with depth) of soil properties. In addition, soils in alluvial landscapes serve as stratigraphic markers indicating episodes of relative stability which separate episodes of erosion and/or deposition. The duration of the stable episode and the degree of stability maintained during it, may be revealed by the relative anisotropy of soil properties.

Given the above statements, the following approach to the analysis of the alluvial soils of the East Branch Walnut valley is suggested. The analysis focuses on the soils of a single geomorphic surface, the first terrace. In Chapter 2, along with general discussions of local physiography, vegetation, climate and prehistory, this geomorphic surface and

adjoining surfaces are described.

Chapter 3 describes the soil-stratigraphy of the first terrace. Two soil-stratigraphic units, representing two soil forming intervals of specified duration, are defined. These units are recognized at five archaeological sites, which provide the chronological data for time-stratigraphic correlation. In Chapter 4, the soil properties of each soil-stratigraphic unit are discussed to determine the dominant trends of pedogenesis during each soil forming interval.

Chapter 5 considers the geomorphic contexts of the soil-stratigraphic units. Landforms and depositional environments associated with each soil are identified and the influence of flooding and surface drainage are discussed. The soil-stratigraphic units and their associated surfaces and sediments are arranged into an alluvial chronology which accounts for the development of the East Branch Walnut valley landscape from ca. 4500 B.P. to the present. In Chapter 6, a model is developed which views the events in this chronology as responses to climatically-induced hydrologic change. Partial confirmation of this model is demonstrated by comparing the East Branch Walnut valley sequence to an emerging pattern of environmental change from the southeastern periphery of the Central Plains. Finally, Chapter 7 summarizes the results of the study and suggests its implications for human prehistory in the East Branch Walnut valley and southern Flint Hills.

CHAPTER 2

STUDY AREA

The Walnut River, a tributary of the Arkansas River, drains an area of approximately 5064 km², primarily from Butler and Cowley Counties, Kansas (Kansas Water Resources Board 1960:25). At the city of El Dorado, in Butler County, the Walnut divides into East and West Branches. This thesis focuses on the East Branch, dealing specifically with the 40.5 km² area now inundated by El Dorado Lake, a U.S. Army Corps of Engineers flood control project. The dam, located 2.4 km northeast of El Dorado, was closed in 1981. Archaeological and geomorphological studies, including those reported in this thesis, were undertaken prior to inundation to mitigate the impact of construction and inundation on prehistoric and historic archaeological sites. Approximately 24 km of the East Branch Walnut valley and the lower 3 to 4 km of Cole, Satchel and Durechen Creek valleys are included in the study area, which is referred to in this thesis as the East Branch Walnut valley.

This chapter describes the study area. The first section describes the southern Flint Hills, which are drained by the Walnut River; the second section focuses on the East Branch Walnut valley in particular; the third section summarizes the human prehistory of the study area.

FLINT HILLS

The Flint Hills comprise the westernmost physiographic province of the Osage Plains of eastern Kansas (Fig. 1). As defined and

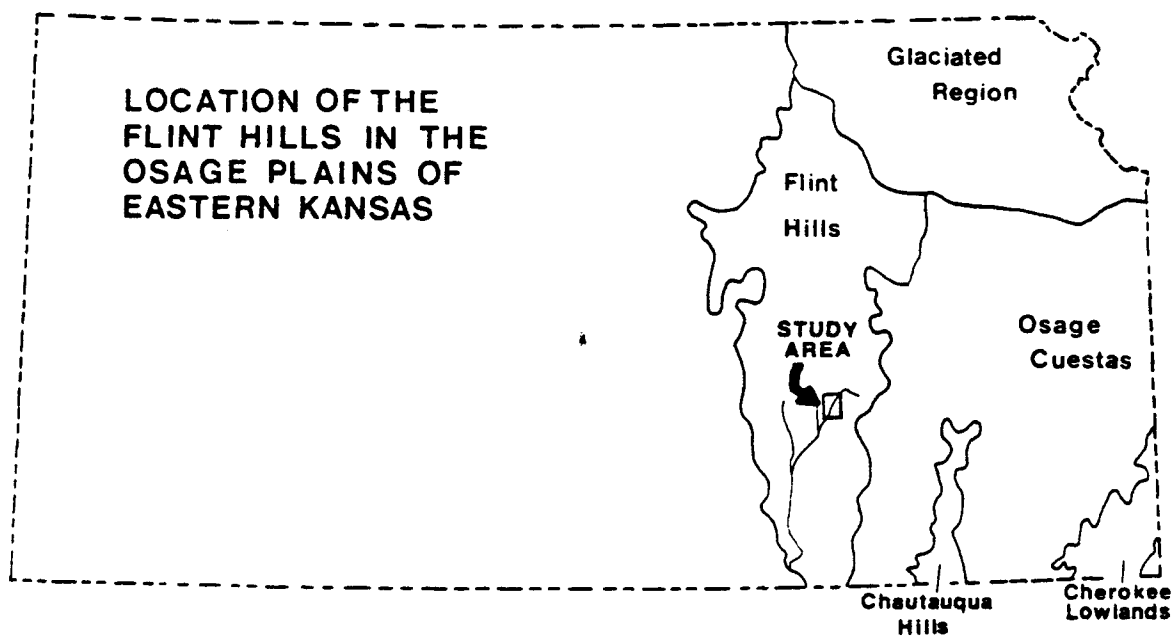


Fig. 1. Location of the study area in the Flint Hills of Kansas (after K.G.S. 1979).

mapped by Schoewe (1949:276,286), the boundaries of the Flint Hills are approximately coincident with the extent of surface outcrops of lower Permian limestones and shales of the Council Grove and Sumner groups (Kansas Geological Survey 1964). The northern boundary of the Flint Hills is that of the maximum southern extent of the Kansas glaciation (Fig. 1). The Flint Hills continue southward into Kay and Osage Counties, Oklahoma, where they are included in the "Northern Limestone Cuesta Plains" geomorphic province of Curtis and Ham (1979).

As is characteristic of the Osage Plains of eastern Kansas, the relief of the Flint Hills is a product of differential erosion of alternating beds of resistant limestone and erodible shale, creating gently rolling upland surfaces interrupted by steep, limestone-capped escarpments. Because the rock layers dip gradually to the west, the major escarpments face eastward and are backed by long, gentle, west-facing slopes (Schoewe 1949:282).

The natural vegetation of the Flint Hills is tall-grass prairie, dominated by big and little bluestem, switchgrass and Indian grass (Kuchler 1974:595-596). Except for occasional lone trees or scrubby thickets, generally restricted to steep hillsides and ravines (Barker 1969:539), the uplands are treeless. Upland soils, which range from heavy clays to thin, rocky lithosols (Bidwell 1966) impose edaphic restrictions which largely exclude trees (Kollmorgen and Simonett 1965). The frequency of fire may also be a factor restricting trees from the upland (Wells 1970).

Gallery forests dominated by hackberry, cottonwood, willow and elm grow in narrow bands alongside major streams. Low-lying areas

of the valleys support stands of wet-prairie and freshwater marsh dominated by prairie cordgrass and various sedges (Kuchler 1964: 600-601). Steep hillsides and ravines at the edge of the valleys commonly support woodlands of scrub oak, hickory, elm and hackberry (Barker 1969:538).

U.S. Weather Bureau records for the city of El Dorado, located near the study area, indicate a mean annual temperature of 13.8°C for the period 1903 to 1980. Average monthly temperatures range from 0.1°C in January to 26.8°C in July (Penner et al. 1975:57). Average annual precipitation at El Dorado (1903-1980) is 83.8 cm. The wettest year on record is 1951, with 130.1 cm; the driest year was 1936, with 44.2 cm. Total precipitation varies markedly from year to year (Fig. 2). For example, the wettest year on record (1951) was followed by five years of drought (1952-1956), during which precipitation averaged 25 cm below normal. These years were followed by six years (1957-1962) of above average rainfall.

Fig. 3 shows the average monthly precipitation at El Dorado, based on the period 1903-1960 (Penner et al. 1975:Table 10). Approximately 70% of the total annual precipitation falls between April and September. The maximum amounts are generally received in May and June. July and August are drier, but most years show slight increases in September. Winters are dry; January is the driest month (Fig. 3).

The principle source of atmospheric moisture in Kansas is the Gulf of Mexico (Bark 1961). Warm, moist air flowing northward from the Gulf of Mexico collides with outbreaks of cold air from northern

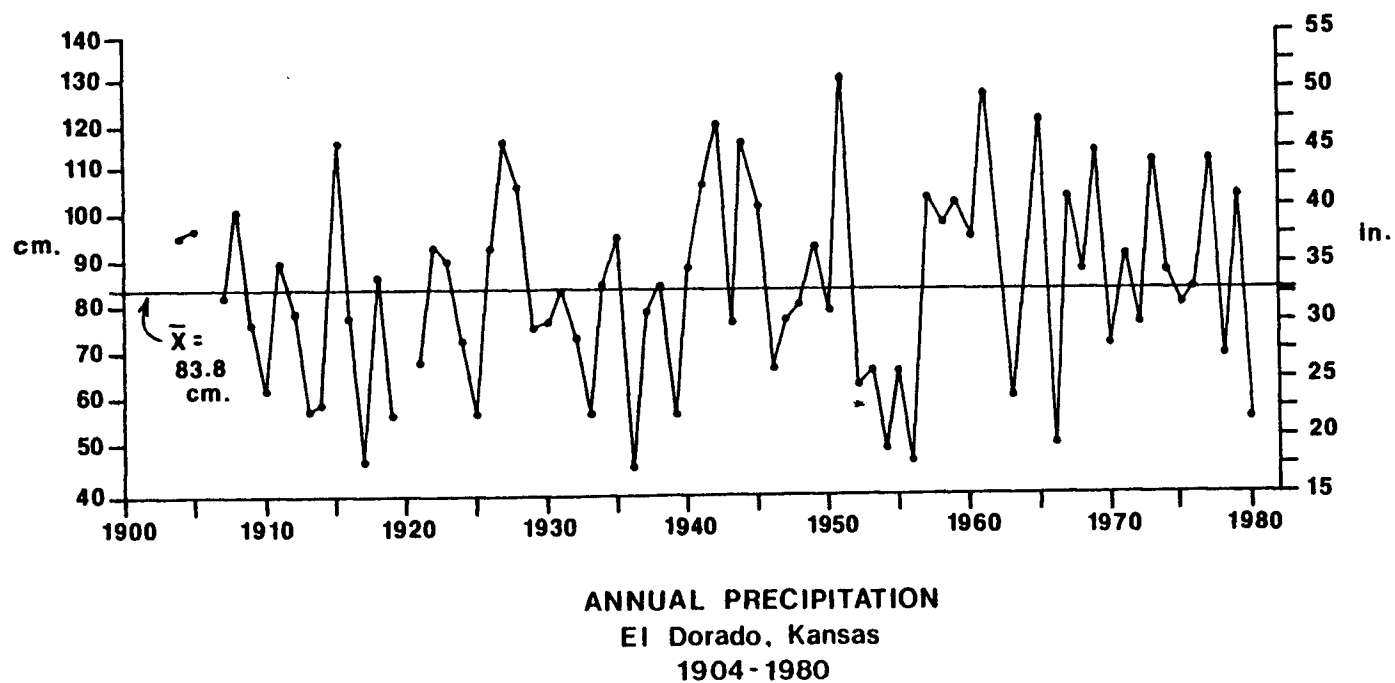


Fig. 2. Annual precipitation at El Dorado, Kansas (1903-1980).

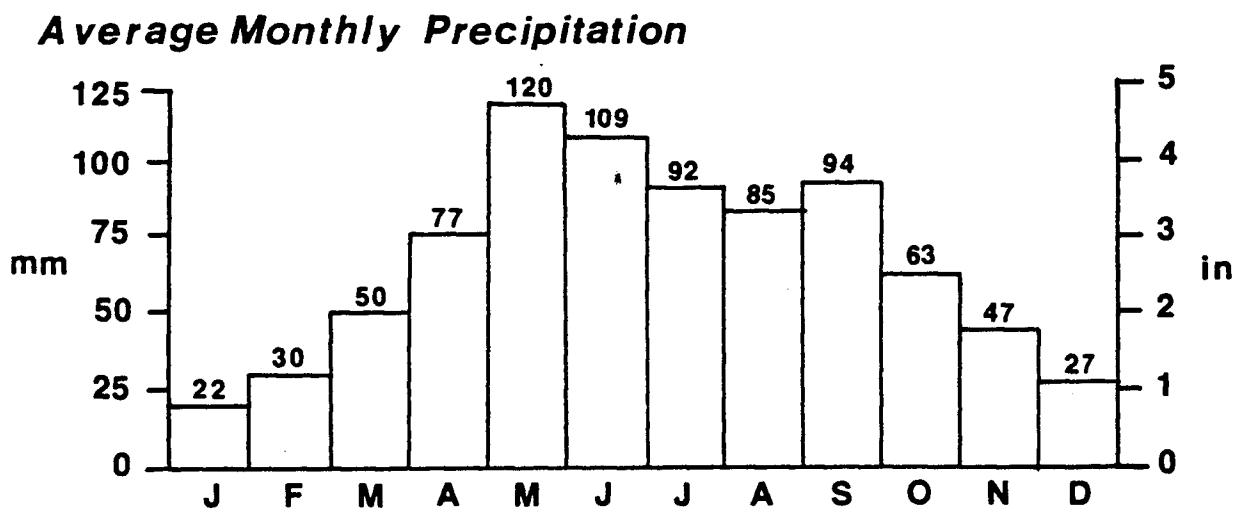


Fig. 3. Average monthly precipitation at El Dorado, Kansas (1903-1960).

latitudes to generate the rains which fall over the Flint Hills in the spring and summer (Penner et al. 1975:56). In winter, the maritime tropical (Gulf) and continental polar air masses are separated by a wedge of mild, dry Pacific air which flows westward from the base of the western Cordillera. This wedge is produced by an intensification of westerly flow in the upper atmosphere; its presence has the effect of blocking the northward penetration of moist Gulf air into the mid-continent, thus accounting for the typical dryness of the winters (Borchert 1950; Bryson and Hare 1974:4). Drought occurs when the strong westerlies of winter persist into spring and summer (Borchert 1950; Bark 1961).

EAST BRANCH WALNUT VALLEY

Of the two branches of the Walnut River which join at El Dorado, the East Branch is the larger. The East Branch drains an area of ca. 694 km², while the West Branch drains 202 km² (Leonard 1972:41,42). The five named tributaries of the East Branch Walnut, listed in order from south to north, are Bird, Bemis, Satchel, Durechen and Cole Creeks (Fig. 4). The first four tributaries join the Walnut from the east, while Cole Creek joins from the north. As shown in Table 2, the creeks drain areas ranging from 60 to 132 km². Average stream gradients vary from 1.3 m/km on the mainstem river to 2.7 m/km on Bird Creek. Prior to completion of El Dorado Lake, the city of El Dorado maintained two small impoundments on Satchel and Bemis Creeks (Fig. 4). Both Lakes El Dorado and Bluestem are included in the larger Federal reservoir.

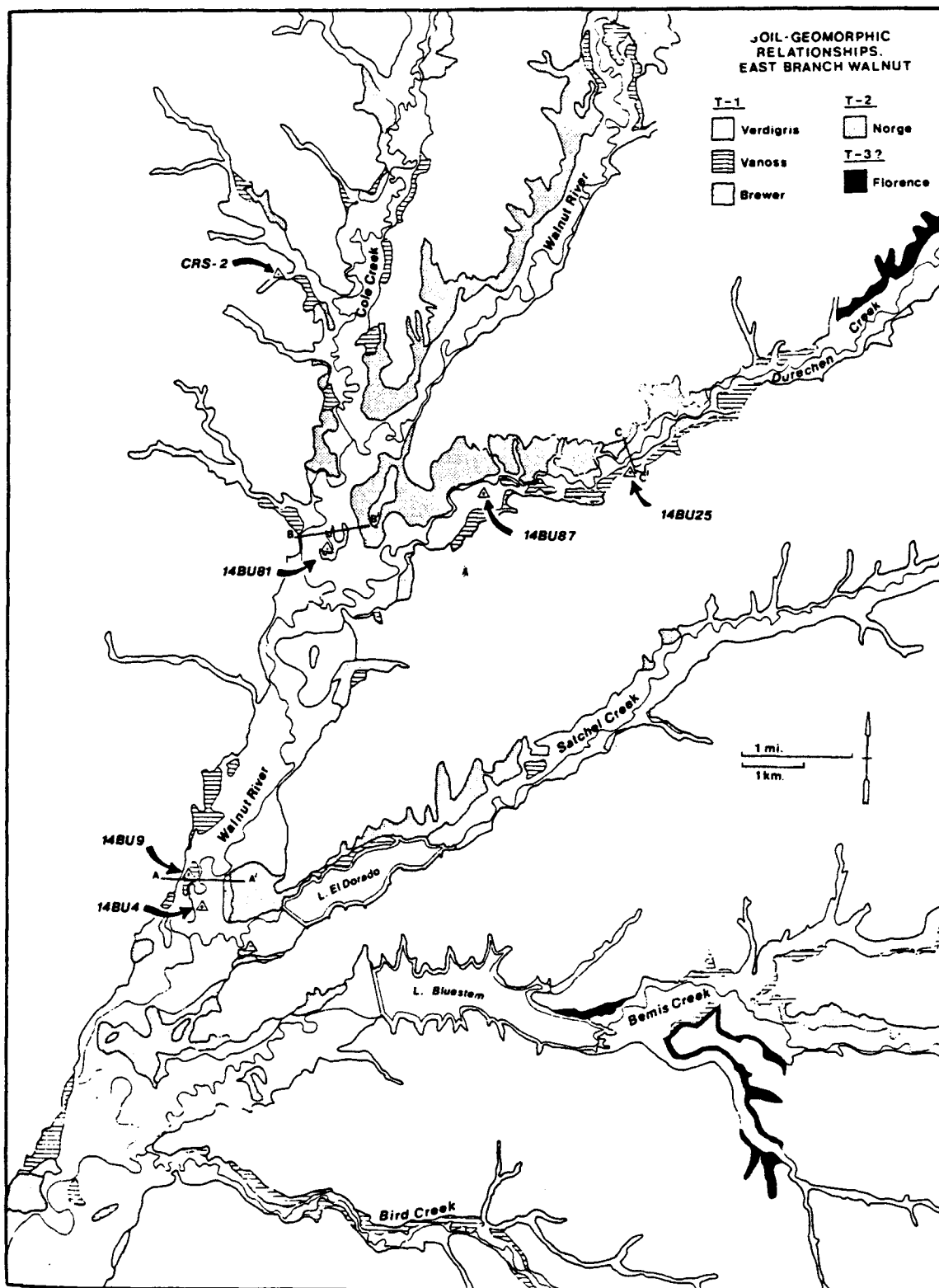


Fig. 4. Alluvial scils of the East Branch Walnut valley, showing sites and valley cross-sections discussed in text (after Penner et al. 1975).

Table 2. Morphometric data for study area streams.

Stream	Drainage Area ^a (km ²)	Stream Length ^b (km)	Vertical Fall ^b (m)	Overall Gradient ^b (m/km)
East Branch Walnut	694	72.5	91	1.3
Bird Creek	60	26.9	73	2.7
Bemis Creek	132	36.5	76	2.1
Satchel Creek	96	36.3	73	2.0
Durechen Creek	98	42.4	75	1.8
Cole Creek	93	34.3	47	1.4

^a measured planimetrically from U.S.G.S., 1:250,000 topographic maps, except for East Branch Walnut, which is quoted from Leonard (1972).
^b determined from U.S.G.S., 1:24,000 topographic maps. Stream length measured with dividers set to 100 feet on map scale.

In Fig. 4, the margins of the East Branch Walnut alluvial valley and its tributaries represent the contact of upland soils, developed over limestone and shale, with alluvial soils, as mapped at 1:24,000 scale by Penner et al. (1975). Upland soils include the Dwight-Labette association, developed primarily over limestone, and the Irwin-Ladysmith association, developed primarily over shale (Penner et al. 1975:2-3). Alluvial soils of the Verdigris-Brewer-Norge association (Penner et al. 1975:3), include the Verdigris, Vanoss, Brewer and Norge series. These alluvial soils are discussed below; detailed descriptions of each series are given in Appendix A.

Observations on the terraces and geomorphic surfaces of the study area were made by the author between 1977 and 1981, largely in the course of archaeological investigations. Field observations were

supplemented by studies of 1:24,000 soil maps (Penner et al. 1975) and 1:200 topographic maps of the El Dorado Lake area prepared by the U.S. Army Corps of Engineers. Geomorphological studies by Drew (1981, 1983) provided information on the thickness of the valley fill.

Two distinct alluvial terraces are present in the study area. Each surface is occupied by a distinct soil series or group of series. The terraces are designated T-1 and T-2, in order of ascending elevation from T-0, the present floodplain.

As shown in valley cross sections prepared from 1:200 scale maps (Fig. 5), the East Branch Walnut and its tributaries are entrenched in narrow, deep channels. In the mainstem valley, the trenches are 4.5-6.0 m deep (Fig. 5a,b). On the tributaries, the depth of entrenchment decreases with distance upstream. In the middle reaches of Durechen Creek (Fig. 2.5c), the trench is 2.5-3.0 m deep. Stream beds of both the mainstem and its tributaries are composed of riffles of limestone rubble and chert gravel, alternating with deep, silt-bottomed pools, similar to the pattern of other Flint Hills streams (Cross 1954). The channel banks are heavily timbered.

Immediately adjacent to the channels are narrow surfaces, generally less than 100 m wide, which are designated T-0. These surfaces form sloping shoulders (e.g., the west side of the channel in Fig. 5a and the east side in Fig. 5b) or level benches (e.g., the east side of the channel in Fig. 5a and the north side in Fig. 5c). In places, two such benches, separated by 30 to 60 cm high scarplettes, are present (Fig. 5a). The sediments underlying these surfaces were not investigated by the author; it is not known whether they represent depositional

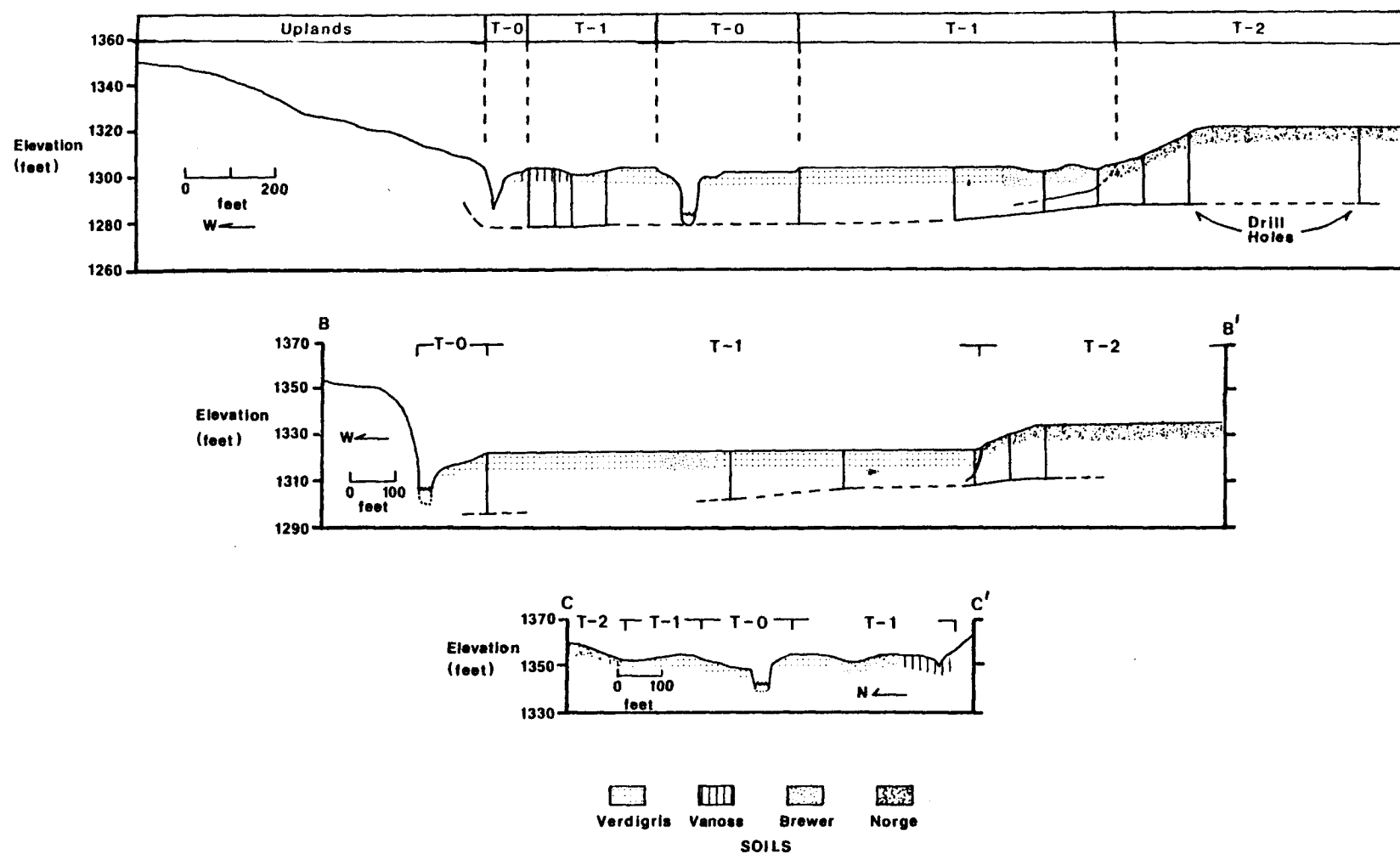


Fig. 5. Topographic cross sections of the East Branch Walnut valley:
a) East Branch Walnut near 14BU4, 14BU9; b) East Branch
Walnut near 14BU81; c) Durechen Creek near 14BU25.

or erosional surfaces. Low levees are developed at the top of channel banks in many localities. Penner et al. (1975) map T-0 surfaces as the frequently flooded phase of the Verdigris series. T-0 surfaces are too narrow to be mapped at the scale of Fig. 4.

The first terrace (T-1) is separated from T-0 surfaces by scarps of 1-2 m. In the mainstem valley, T-1 surfaces comprise the largest area of the valley floor. Penner et al. (1975) map Verdigris, Brewer and Vanoss soils on these surfaces (Figs. 4-5). Vanoss soils (Udic Argiustolls) are mapped on slightly higher, better-drained surfaces, while the Brewer series (Pachic Argiustolls) are mapped on low-lying, concave surfaces (e.g., the Brewer soil surfaces shown in Fig. 5a,c). Verdigris series soils (Cumulic Hapludolls) occupy remaining portions of T-1 surfaces, which are level to gently sloping and moderately well drained. The relationship of soils to landscape position on the T-1 surface is discussed further in Chapter 5.

In the mainstem valley, second terrace (T-2) surfaces are separated from adjacent T-1 surfaces by scarps of 3.5-6.0 m (Fig. 5a,b). Penner et al. (1975) map the Norge series (Udic Paleustolls) on these surfaces. As shown in Fig. 4, T-2 surfaces in the mainstem valley are most extensive near the confluences of the Walnut and each of its five tributaries. At these localities, the terrace surface projects as an interfluvial lobe into the valley, extending nearly across the valley in places (e.g., upstream from the confluence of Bemis and Bird Creeks, Fig. 4).

In general, T-1 and T-2 are less well-defined on the tributaries than on the mainstem. T-1 surfaces on the tributaries are considerably

dissected by abandoned stream channels, which often give the surface an undulating appearance (e.g., Fig. 5c). T-2 surfaces, identified by the distribution of the Norge series, are extensive along the north sides of the valleys of Satchel and Durechen Creeks (Fig. 4). These surfaces are not separated from T-1 by scarps, but instead slope gradually to the valley floor (Fig. 5c), suggesting that the surfaces are erosional rather than depositional.

The T-1 and T-2 surfaces are each underlain by a distinctive valley fill unit. Drew (1981, 1983) identified these two units and provided data on their stratigraphic relationships, including the drill hole data used in Fig. 5a to show the depth to bedrock. T-2 surfaces are underlain by a strongly-weathered, reddish-hued sediment, which is up to 9 m thick (Fig. 5a,b). The advanced weathering of this unit is reflected by characteristics of the Norge soil (see Appendix A). The B horizon of this soil, over 1.6 m thick, has blocky structure, silty clay loam textures and 5YR (moist) soil colors which increase in chroma and value from 3/4 in the B1 to 5/6 in the B3 horizons. As revealed in observations in stream banks, borrow pit exposures and drill holes (author's observations; Drew 1981, 1983), 5YR soil colors characterize the entire 9 m thickness of the deposit. The T-1 surface, by contrast, is underlain by less intensely weathered sediment which ranges in color from dark gray to grayish brown (10YR and 7.5YR hues). As demonstrated by deep holes drilled by Drew (1983), with assistance from the Kansas Geological Survey, T-1 deposits are inset against T-2 deposits (Fig. 5a,b).

The valley is underlain by a relatively smooth bedrock surface (Fig. 5a,b) which tends to be slightly higher beneath T-2, relative to T-1 surfaces (Drew 1983). In Fig. 5(a,b), the bedrock surface is ca. 1 m higher beneath T-2, relative to its position beneath T-1.

A third terrace (T-3) is tentatively identified on Bemis and Durechen Creeks (Fig. 4). Penner et al. (1975) map the soils of this surface as Florence cherty loams. Where observed by the author on Bemis Creek, this terrace consists of deposits of coarse chert rubble, which mantle the valley edges to unknown depths. The chert is derived from the Florence formation, which outcrops along the length of these streams (Haury, in press). Both angular and rounded pieces of chert occur in the deposits on Bemis Creek.

PREHISTORY OF THE EAST BRANCH WALNUT VALLEY

The University of Kansas Museum of Anthropology conducted archaeological investigations in the East Branch Walnut valley from 1967 through 1981. Investigations from 1967 through 1975 were funded by the National Park Service, the Butler County Historical Society and the University of Kansas. The 1976-1981 investigations were funded by the U.S. Army Corps of Engineers. These investigations, spanning 14 years and including the work of over 20 individual investigators, are summarized by Leaf (1979:5-15) and Johnson (1983). A total of 66 prehistoric sites were recorded. Test excavations were conducted on 18 of the 66 sites; large scale excavations were conducted on an additional 7 sites. Sixty-six historic Euro-American sites were recorded, 10 of which were excavated (Johnson 1983). Studies of the

Table 3. Radiocarbon dates from archaeological sites in the East Branch Walnut valley (updated from Leaf 1979a:10).

Lab Number	Site	Context	Radiocarbon Years B.P.	Calendar Date (Uncorrected)	Cultural Historical Period
* UGa-3947 ^a	14BU5	Feature 1	Modern	-	
* UGa-3043 ^b	14BU32	Feature 301 (charcoal concentration)	Modern	-	
* UGa-2502 ^c	14BU55	Feature 1 (charcoal concentration)	325 ± 140	A.D. 1625	
* UGa-2561 ^d	14BU9	Composite sample from flotation	520 ± 55	A.D. 1430	
* UGa-3044 ^b	14BU32	XU103/XU104 (composite sample)	720 ± 70	A.D. 1230	
UGa-1347 ^e	14BU71	Feature 2 (limestone-filled pit)	740 ± 75	A.D. 1210	Plains Village
UGa-3137 ^f	14BU57	Excavation Unit 5 (composite sample)	755 ± 235	A.D. 1195	Village/Woodland?
UGa-1345 ^e	14BU55	House 1	890 ± 60	A.D. 1060	
UGa-2504 ^c	14BU55	1 m ² excavation unit (composite sample)	950 ± 135	A.D. 1000	
UGa-1346 ^e	14BU55	House 1	970 ± 80	A.D. 980	
UGa-2501 ^c	14BU55	Feature 18 (charcoal concentration)	984 ± 45	A.D. 965	
UGa-2500 ^c	14BU55	Feature 6 (charcoal concentration)	1065 ± 65	A.D. 885	Plains Woodland
UGa-2503 ^c	14BU55	Feature 19 (post mold)	1090 ± 55	A.D. 860	
UGa-2560 ^g	14BU31	Feature 302 (limestone-filled pit)	1215 ± 55	A.D. 735	
UGa-2807 ^h	14BU31	Feature 304 (limestone-filled pit)	1240 ± 75	A.D. 710	
N-769 ⁱ	14BU9	40-55 cm below surface (no other provenience)	1970 ± 110	20 B.C. /	L Walnut phase
N-1276 ⁱ	14BU9	45 cm below surface (no other provenience)	2060 ± 80	110 B.C. ✓	A T E
N-1552 ⁱ	14BU9	57-88 cm below surface (no other provenience)	3100 ± 165	1150 B.C.	
N-1277 ⁱ	14BU9	84 cm below surface (no other provenience)	3240 ± 85	1290 B.C. ✓	A R C El
N-770 ⁱ	14BU9	100-125 cm below surface (basin-shaped refuse pit)	3650 ± 140	1700 B.C. ✓	H Derado phase
N-771 ⁱ	14BU9	125-150 cm below surface (no other provenience)	3910 ± 160	1960 B.C. ✓	A I C
N-1278 ⁱ	14BU9	128 cm below surface (no other provenience)	3980 ± 100	2030 B.C. ✓	
* N-1551 ⁱ	14BU9	250 cm below surface (limestone concentration)	4150 ± 110	2200 B.C.	Chelsea phase
UGa-2806 ^j	14BU25	Feature 5 (limestone concentration)	4435 ± 100	2485 B.C.	
* N-1280 ⁱ	14BU9	250 cm below surface (limestone concentration)	4600 ± 125	2650 B.C.	
* N-1279 ⁱ	14BU9	178 cm below surface (boundary of soil horizons)	4830 ± 105	2880 B.C.	

* Dates questionable due to recency or stratigraphic inversion.

^a Parisi 1981; ^b Leaf and Root 1981; ^c Adair and Brown 1981; ^d Leaf 1981; ^e Fulmer 1977; ^f Brown 1982; ^g Root 1981c; ^h previously unpublished; ⁱ Yamasaki et al. 1972, 1974, 1977; ^j Root 1981b.

geomorphology (Drew 1981, 1983), palynology (King 1979; Egbert 1982) and chert resources (Haury, in press) of the East Branch Walnut valley were also undertaken.

As a result of archaeological investigations, it is possible to outline a prehistoric cultural sequence for the East Branch Walnut valley which extends from approximately 700 to 4500 years ago. Twenty-four radiocarbon dates from seven archaeological sites, summarized in Table 3, provide a framework for this sequence. Sites of three broadly defined culture-historical periods are represented. As defined by Johnson (1983), these are Plains Village (250-950 B.P.), Plains Woodland (950-1950 B.P.) and Late Archaic (1950-5450 B.P.). This division reflects widely used taxonomies in Central Plains prehistory, full discussion of which is beyond the scope of this thesis. For general overviews of Central Plains prehistory, the reader is referred to Wedel (1959, 1961, 1978) and Chapman (1975, 1980). The Archaic period is treated by Johnson (1980); the Woodland period by Johnson and Johnson (1982) and Reynolds (1979:84-106); and the Plains Village period by Blakeslee (1978). The following sections discuss the manifestation of each period in the archaeological record of the East Branch Walnut valley.

Plains Village

The Plains Village period is poorly represented in the East Branch Walnut valley. The one single-component site of this period identified in the study area is 14BU71 (Fulmer 1977:40-57). Excavation and surface collections at this site recovered cordmarked and smooth-

surface sherds (untempered or tempered with shell or indurated clay), small, side-notched and unnotched, isosceles triangular projectile points and double-beveled knives of the Harahey type. The lithic artifacts in particular are reasonably good indicators of a Plains Village affiliation. The site was radiocarbon dated at ca. 750 B.P. (Table 3).

Other sites yielded small, unnotched and side-notched projectile points: for example, 14BU57 (Leaf 1979b; Brown 1983), 14BU87 (Root 1979a) and 14BU4 (Artz 1983). However, these sites also possess Woodland components which, on the basis of excavation, appear to represent the major occupation of these sites. It is probable that Plains Village utilization of the study area was not intensive.

Plains Woodland

According to Johnson's (1983:9) tabulation, 24 of the total 75 components identified in the study area are Plains Woodland manifestations. Sixteen Woodland components have been excavated. Ceramics from Woodland components are cordmarked, tempered predominantly with limestone or indurated clay. At 14BU55, crushed bone was the predominant tempering material (Adair and Brown 1981:268). Projectile points include small, stemmed or corner-notched forms (Scallorn type) and a variety of forms of larger stemmed and corner-notched points. The former were presumably hafted on arrows, while the latter were hafted on atlatl darts, or possibly served as hafted knives.

Woodland sites occur both as relatively sedentary "hamlet or village" occupations and as smaller, limited-function occupations

(Johnson 1983:23-24,26-27). Sedentism is suggested by the presence of oval to rectangular, wattle-and-daub structures and by evidence of horticulture in the form of maize, pumpkin and sunflower seeds at 14BU55 (Adair and Brown 1983:254-261, 324). Maize has also been recovered from 14BU57 (Brown 1983). Less intensive, less permanent Woodland occupations are documented at 14BU31 (Leaf 1979b), 14BU16 (Artz 1981), 14BU32 (Leaf and Root 1981) and 14BU4 (Artz 1983). These sites probably represent temporary, perhaps seasonal, occupations by small groups.

The major Plains Woodland occupation of the East Branch Walnut apparently occurred late in the Woodland period. Radiocarbon dates from Woodland components (Table 3) range from 1240 ± 75 B.P. at 14BU31 to 890 ± 60 B.P. at 14BU55. Dates from 14BU9 and 14BU32 were considered anomalously recent by the investigators of these sites. A date of 755 ± 235 B.P. from 14BU57 suggested to Brown (1983) that the site was transitional between Plains Woodland and Village periods, an argument which she supported by artifactual evidence from the site, which includes both stemmed, Scallorn-like and side-notched, isosceles triangular projectile points (Brown 1983; Leaf 1979a). However, the large sigma of the date cautions against its use in determining whether a single, transitional occupation, or a mixed, multiple component occupation is represented.

The only substantive evidence of occupation during the first half of the Woodland period (ca. 1500 to 2000 B.P.) consists of 2 dentate-stamped rimsherds from 14BU9 (Grosser 1970:94-95). Dentate stamping

is generally associated with Middle Woodland, Hopewellian manifestations, which have been recognized between 1500 and 2000 B.P. in the Kansas City area (Johnson 1979), in the Kansas River drainage of north-central Kansas (O'Brien et al. 1979) and northeastern Oklahoma (Bell and Baerreis 1951).

Late Archaic

The Late Archaic period has been investigated in excavations at seven sites in the East Branch Walnut basin. One of these, the Snyder site (14BU9), possessed stratified deposits spanning the interval 4500 to 2000 B.P. (Grosser 1970, 1973, 1977). Deeper and presumably older components were detected (Haury and Leaf 1981), but their excavation was unfortunately never funded (Johnson 1983:16). Grosser divided the Late Archaic sequence at 14BU9 into three phases, which he termed Walnut, El Dorado and Chelsea. These units were originally identified on the basis of changes in projectile point form upward through the cultural deposits (Grosser 1970). These changes, which are important for stratigraphic correlation of the East Branch Walnut soils, are detailed in Chapter 3. Grosser (1973, 1977) and Bradley (1973) subsequently suggested that the three phase sequence was characterized by changes from phase to phase in subsistence strategies. These changes involved a shift from "diffuse adaptation", focusing on a wide range of resources in the Chelsea and El Dorado phases to a "specialized adaptation", focusing on a select subset of subsistence resources in the Walnut phase.

The Walnut phase, the most recent, was associated at 14BU9 with radiocarbon dates of 1970 ± 110 and 2060 ± 80 B.P. (Table 3). A series of triangular, corner-notched projectile points small enough to be used as arrow points were recovered from Walnut phase levels, along with larger dart points. These artifacts represent the earliest appearance in the East Branch Walnut of the bow and arrow (Johnson 1983: 21). Other Walnut phase components are tentatively identified on the basis of comparisons of projectile points and soil stratigraphy (see Chapter 3) at 14BU81 and 14BU87, but little is known of these occupations, due to the limited extent of test excavations (Artz 1981).

El Dorado phase radiocarbon dates from 14BU9 range from 3240 ± 85 to 3980 ± 100 B.P. A date of 3100 ± 165 B.P. from an El Dorado phase component at 14BU50 (Bradley 1973) overlaps this range (Table 3). Sites 14BU81 and 14BU89, tested in 1978, were also assigned to the El Dorado phase on the basis of projectile point morphology (Artz 1981). The four sites represent a diversity of site situations, possibly suggesting the outlines of a functionally interrelated settlement system. Sites 14BU9 and 14BU81 are situated on the first terrace of the Walnut River; 14BU50 is located on a talus slope at the edge of Bird Creek valley, while 14BU89 is situated in the uplands overlooking the Walnut valley.

No radiocarbon dates were directly associated with the Chelsea phase at 14BU9. Grosser (1973:235-236) considered the three dates from levels underlying that of the Chelsea phase to be stratigraphically inverted and therefore unacceptable. The rejection of these

dates, however, may not be warranted as discussed in Chapter 5, below. Stratigraphic correlation with the Late Archaic component at 14BU25 (see Chapter 3) suggests that the Chelsea phase dates to approximately 4500 B.P. (Table 3). Chelsea phase components were identified at 14BU4 (Fulmer 1976; Artz 1983) and tentatively at 14BU81 (Artz 1981) on the basis of stratigraphic and artifactual considerations. The relationship between 14BU25 and the Chelsea phase is problematic (Root 1981b), due primarily to the small number of diagnostic artifacts from 14BU4, 14BU9 and 14BU81. Sites 14BU25 and 14BU9 may have functioned as base camps (Root 1981b), while 14BU4 and 14BU81 probably represent short term occupations focusing on lithic reduction activities (Artz 1981, 1983).

Summary

The principal occupations of the East Branch Walnut basin apparently occurred from 2000 to at least 4500 B.P., in the Late Archaic period, and from ca. 900 to 1200 B.P. in the late Plains Woodland period. Occupation of the area between 1200 and 2000 B.P. and after 900 B.P. was not intensive. Although evidence from the unidentified pre-Chelsea phase component at 14BU9 suggests the presence of occupations prior to 4500 B.P., these occupations remain essentially unknown.

The interpretations offered above are tentative; they presuppose that the excavated sample of sites from the East Branch Walnut are a representative indicator of the valley's prehistory. In other words, the interpretations assume that the low frequency of early Plains Woodland and Plains Village components does not simply reflect the

fact that such components were not detected by survey and excavation. This assumption requires evaluation. Nevertheless, if the interpretations are tentatively accepted, then the East Branch Walnut prehistoric sequence suggests the outline of a dynamic cultural model in which periods of valley occupancy alternated with periods of relative non-occupancy.

CHAPTER 3

SITE DESCRIPTION AND STRATIGRAPHIC CORRELATION

Although the previous chapter provided a general discussion of soils and geomorphic surfaces in the East Branch Walnut basin, a more detailed account of soil-geomorphic relationships is required to identify the pedogenic and geomorphic processes which shaped the landscape and to order those processes in a chronological sequence. This chapter presents specific information on the soil stratigraphy and archaeological deposits of five archaeological sites. The chapter has two objectives: 1) to describe the topography, soil stratigraphy and archaeological deposits of each site and 2) to define soil-stratigraphic units and soil-forming intervals by correlating soil-stratigraphic and archaeological data among sites.

SITE DESCRIPTIONS

Locations of the five archaeological sites, designated 14BU4, 14BU9, 14BU25, 14BU81 and 14BU87, are shown in Fig. 4, in the previous chapter. These sites suit the purposes of this analysis for the following reasons. First, each possesses at least one archaeological component of known relative or absolute age, thus providing chronological control. Each site was the subject of stratigraphically-controlled archaeological excavation, from which soil profiles and artifact distributions are available. Finally, soil samples from each site were available for analysis, providing data on soil properties.

Radiocarbon dates from archaeological deposits at 14BU9 and 14BU25 provide the data for age assessment in this study. Although

chronometric dates are not available for the remaining sites, relative ages may be estimated from artifacts and stratigraphic evidence. Radiocarbon-dated deposits at 14BU9 and 14VU25 are associated with stratigraphically distinct and morphologically homogeneous groups of projectile points. The occurrence of similar-appearing artifacts in comparable stratigraphic contexts at the remaining sites allows relative age estimates to be made.

Field techniques used by archaeologists in the study area to record cultural and natural stratigraphy varied widely, depending on the scale and objectives of excavations, and on the experience and research interests of individual investigators. Investigations were conducted at 14BU4, 14BU9 and 14BU25 for as many as seven field seasons, involving surface survey, testing and large scale, block excavation. Relatively more subsurface data were available from these sites than from 14BU81 and 14BU87, which were only minimally tested. The initial investigations of 14BU4, 14BU9 and 14BU25, accomplished between 1968 and 1972, focused primarily on establishing a local cultural-historic sequence. Although soil profiles were mapped and described, pedologic and geomorphic interpretations of soil stratigraphy was limited. Excavations and analyses undertaken after 1977 specifically addressed problems of site stratigraphy and geomorphic context and thus provide the most detailed data on site deposits.

14BU4

Site 14BU4 is located approximately 0.5 km northwest of the confluence of the Walnut River and Satchel Creek (Fig. 4). The site,

initially recorded in 1967 (Eoff and Johnson 1968:6-7), was excavated in 1974 (Fulmer 1976:42-59) and 1979 (Artz 1983).

Topography of the site locality is shown in Fig. 6. The site is situated at the edge of the first terrace, which forms a scarp 1.5 m (5 feet) high. In Fig. 6, the 1300 ft contour lies approximately midway on this slope. The terrace scarp which forms the western edge of 14BU4 is separated from the river by a narrow stretch of floodplain (T-0). The Walnut is entrenched approximately 6 m (20 feet) below the T-0 surface.

East of the T-1 edge, the first terrace has little surface relief, with the exception of a series of broad, shallow swales outlined by the 1304 foot contour (Fig. 6). These swales, which are elongate from north to south, are bounded on the east by the abrupt rise of the second terrace. The T-2 scarp rises 4.3 m (14 feet) to a flat surface, the edge of which is delimited by the 1320 foot contour on the locality map (Fig. 6). The terrace surface extends 300 m eastward before grading into an upland surface.

Soils on T-1 and T-0 surfaces are mapped as the Verdigris series, except for the swales at the base of T-2, which are mapped as Brewer series. Soils on the T-2 scarp and surface are mapped as the Norge series (Penner et al. 1975; and Fig. 4).

Prehistoric material on the surface of 14BU4 is concentrated on a slight, convex rise at the edge of the first terrace. This feature is delimited in Fig. 6 by the 1302 foot contour. Archaeological excavations, placed in the densest areas of the surface scatter, include

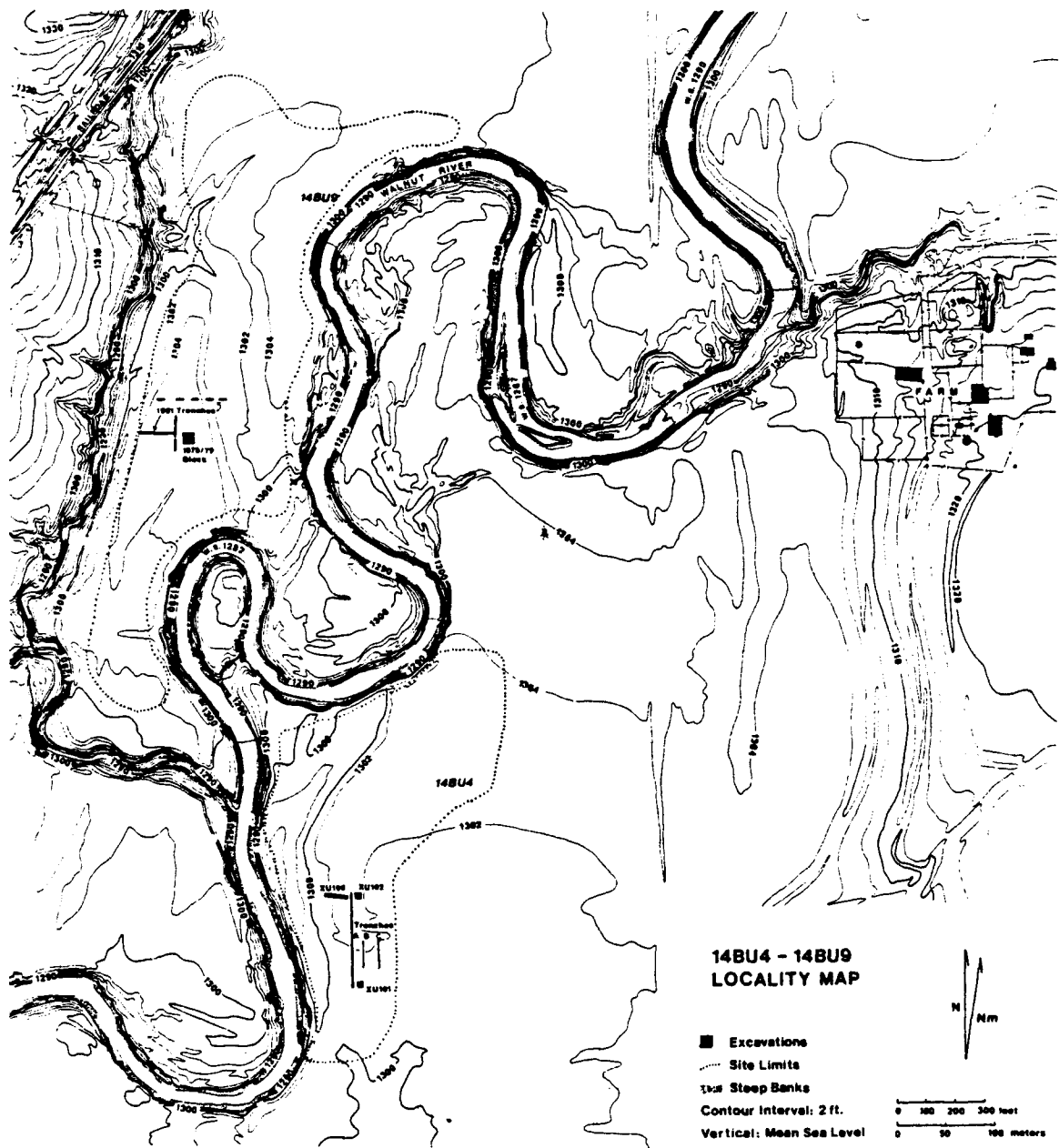


Fig. 6. Topographic map of the 14BU4-14BU9 locality. Source: U.S. Army Corps of Engineers, 1:200 topographic maps.

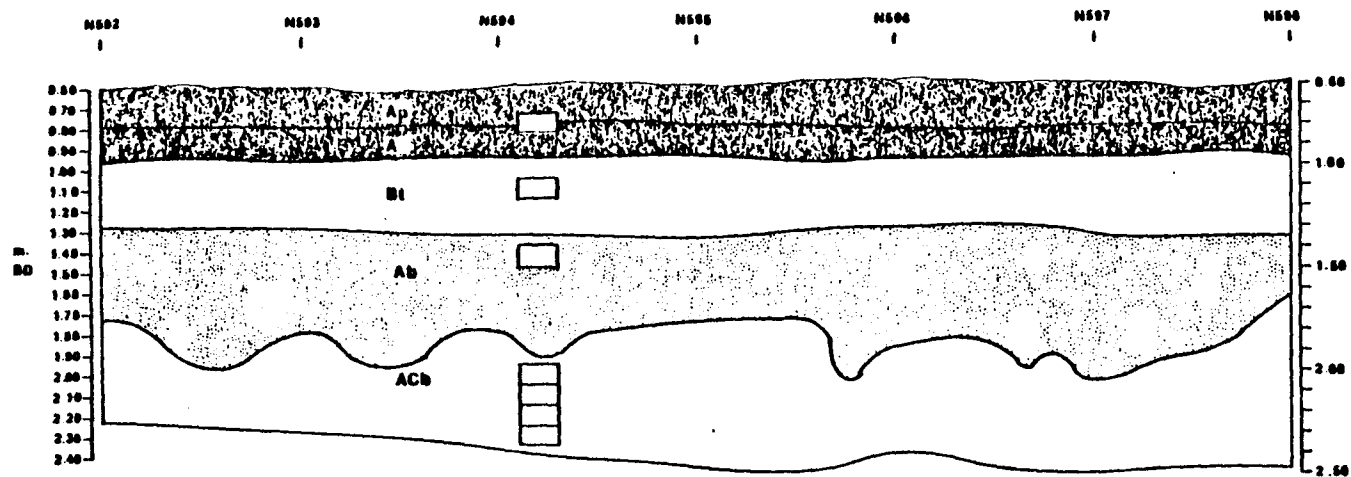
Backhoe Trenches A, B and C; two block excavations (XU101 and XU102); and a hand-excavated test trench (XU100) (Fig. 6). XU100 was excavated in 1974; all other units were dug in 1979.

Soil stratigraphy was described in detail in XU102 and in the north end of Backhoe Trench C. The sequence of soils observed in XU102 (Fig. 7) was as follows:

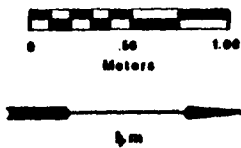
- Ap 0-20 cm. Very dark grayish brown (10YR3/2) silt loam; granular structure; slightly acid; smooth, abrupt boundary.
- A 20-35 cm. Very dark grayish brown (10YR3/2) silt loam; subangular blocky structure; slightly acid; smooth, gradual boundary.
- Bt 35-70 cm. Dark brown (7.5YR4/4) silty clay loam; subangular blocky structure; few, thin, patchy clay films; neutral; smooth, gradual boundary.
- Ab 70-120 cm. Very dark grayish brown (10YR3/2) silt loam; subangular blocky structure; many, coarse, prominent mottles of dark brown (7.5YR4.4); friable; neutral; clear, wavy boundary.
- ACb 120-180 cm. Very dark grayish brown (10YR3/2) silty clay loam; blocky structure; few, thin, patchy clay films; few fine iron and manganese concretions; neutral; smooth, gradual boundary.
- Cb 180-230 cm. Dark brown (10YR4/3) silty clay loam; blocky structure; many fine iron and manganese concretions; medium to slightly acid.

This soil sequence was observed in XU102, XU101, XU100, and Backhoe Trenches A and B. Two soils are present. The surface soil exhibits a dark colored A horizon overlying a clayey, reddish-hued Bt horizon. The second soil is buried at depths ranging from 50 to 70 cm and consists of a mottled Ab horizon overlying ACb and Cb horizons. The surface soil resembles the Vanoss series, as described by Penner et al. (1975:20). The buried soil is similar to the Verdigris series (Penner et al. 1975:21; Appendix A).

This stratigraphic sequence is restricted to soils beneath the low ridge at the edge of the terrace. The Bt horizon is thinner and



14BU4
XU102
Stratigraphic
Profile
West Wall,
Trench A,
along E499



Soil Sample

ARTIFACTS FROM PROFILE

- B Bone
- C Charcoal
- E Burned Earth
- F Flake
- L Limestone

Fig. 7. Stratigraphic profile of XU102 at 14BU4.

its boundaries less distinct in Backhoe Trench B and XU101, situated at the edge of the ridge, than in ridge crest areas of Trench A, XU102 and XU100.

In Trench C, 40 m east of Trench A, the surface soil lacks a B horizon, and no buried soil is apparent (Fig. 8). The pedon observed in the north end of Trench C corresponds closely to that described for the Verdigris series by Penner et al. (1975:22; Appendix A), and is described as follows:

- Ap 0-20 cm. Very dark grayish brown (10YR3/2) silt loam; granular structure; neutral; smooth, abrupt boundary; contains numerous, fine, bioturbated laminae of grayish silt.
- A11 20-52 cm. Very dark grayish brown (10YR3/2) silty clay loam; granular structure; neutral; smooth, gradual boundary.
- A12 52-84 cm. Very dark grayish brown (10YR3/3) silty clay loam; granular structure; neutral; smooth, gradual boundary.
- AC 84-120 cm. Very dark grayish brown (10YR3/2) silty clay loam; subangular blocky structure; many, large mottles of 10YR4/4 and 7.5YR4/4; neutral.

The difference in soil stratigraphy between excavations on and off the ridge crest indicates that the ridge is a distinct soil-geomorphic landscape element. The geomorphic identity of the feature is suggested by its relationship to the edge of the T-1 terrace. The scarp separating the T-0 and T-1 surfaces along the western edge of 14BU4 can be visualized as a relict cutbank of the Walnut River, formed during or following the abandonment of the T-1 surface, but prior to the aggradation of the present-day T-0 surface. Given this interpretation, the low ridge is plausibly interpreted as a natural levee formed at the top of the former cutbank. Natural levees are described as "wedge-shaped ridges of sediment bordering stream channels",

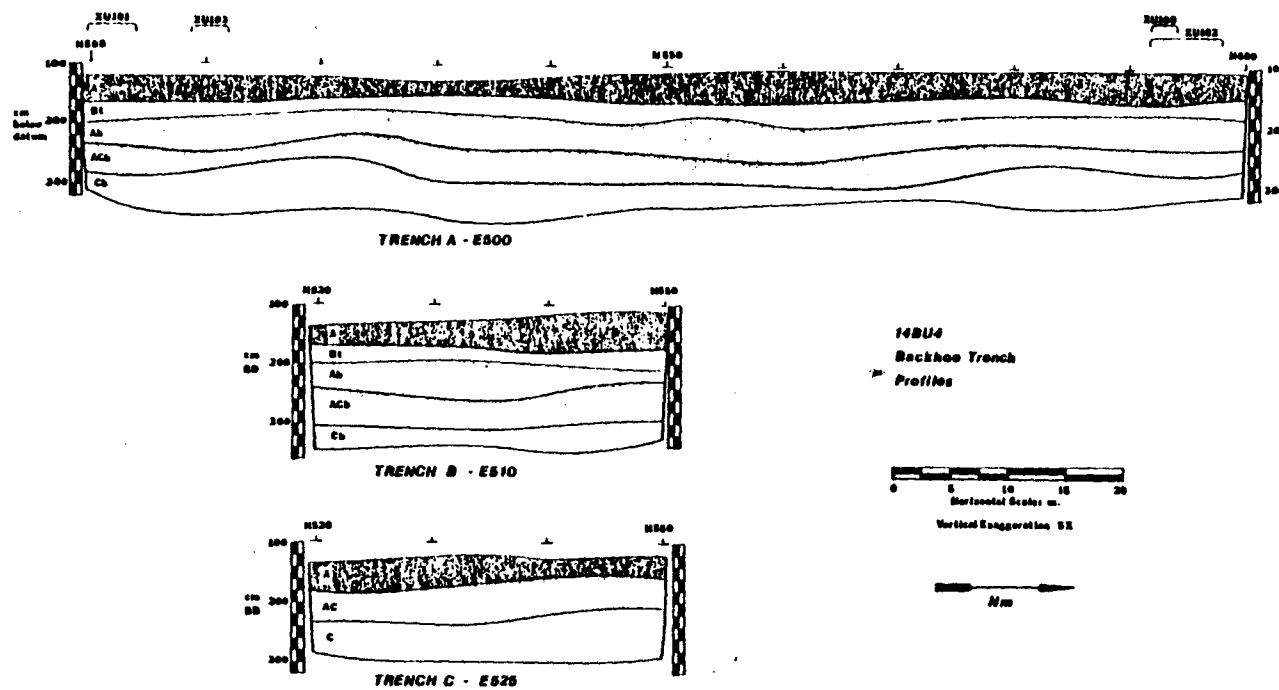


Fig. 8. Stratigraphic profiles of backhoe trenches at 14BU4.

most commonly "best developed on the concave banks of streams" (Allen 1965:121). Levees form as floodwaters overtop the stream bank and spread over the floodplain. In the process, the floodwater experiences a reduction in velocity that results in loss of some suspended sediment (Wolman and Leopold 1957:98). Maximum deposition occurs near the channel, with rates and amounts of deposition decreasing away from the channel. For this reason, levees tend to rise abruptly from stream banks, but slope gradually away toward the valley sides (Allen 1965:121; Reineck and Singh 1975:244-245).

The wedge-shaped profile typical of a modern-day active levee is not observed in the terrace edge ridge at 14BU4, probably as a result of erosion following abandonment of the first terrace. Continual cultivation of the site during the 20th century may have further reduced levee relief.

Two archaeological components were encountered at 14BU4 (Fulmer 1976; Artz 1983). A Plains Woodland component, identified by small, corner-notched and stemmed projectile points and cord-marked ceramics tempered with limestone and indurated clay, was exposed at the site surface, concentrated on the crest and slopes of the abandoned levee. In XU100, XU101 and XU102, Woodland material was primarily restricted to the A horizon of the surface soil. The few specimens encountered in the Bt horizon are presumed to have been introduced by pedoturbation.

A buried component was encountered at depths of 115 to 165 cm below surface in XU100 and XU102, in the ACb of the buried soil. Although the component yielded insufficient organic material for radio-carbon dating, four triangular-bladed, broad-stemmed projectile points

resemble those recovered from the Chelsea phase levels of 14BU9, and suggest a Late Archaic occupation, prior to 4000 B.P. (Fulmer 1976:91; and below).

Distributional data indicate that the Late Archaic cultural stratum is relatively thin. In XU102, 90% of all Late Archaic artifacts (2563 of 2832) were recovered between 135 and 155 cm below surface. In XU100, 78% of all Archaic materials (2452 of 3140) were recovered between 145 and 155 cm below surface.

In XU102, all artifacts larger than 2 cm in any dimension were plotted three-dimensionally in the course of excavation. A regression line fitted to the vertical distribution of artifacts reveals that the cultural deposit conforms to a former land surface which sloped 15 to 20 cm to the south in the 6 m north-to-south length of XU102 (William 1980). The low correlation coefficient ($r = 0.52$) is attributed to a slight westward component to the downward slope, which is evident when vertical distributions from west to east are observed, and to vertical displacement of artifacts by pedoturbation.

The former surface was not associated with visible natural stratigraphy, but was apparent only in three dimensional plots of artifacts. Excavations were not sufficiently extensive to suggest the landform represented by this surface. The lack of plotted data from XU100 precludes tracing the surface in this direction.

To summarize, 14BU4 is associated with an abandoned levee at the edge of the first terrace of the Walnut River. Soil stratigraphy beneath the levee surface consists of an upper soil with A and Bt horizons, which resembles the upper horizons of the Vanoss series.

This soil buries a weakly developed, A-C profile which resembles the Verdigris series. On level surfaces east of the levee, B horizons are absent and an A-C profile like that buried beneath the levee is found.

Plains Woodland artifacts occur on the site surface and in the A horizon of the surface soil, on and off the levee. A Late Archaic component, identified with the Chelsea phase at 14BU9, conforms to a relatively well-defined former land surface, buried 135 to 145 cm below the crest of the levee.

14BU9

The Snyder site, 14BU9, is located 1.6 km north of the confluence of the Walnut River and Satchel Creek, approximately 0.5 km northwest and across the river from 14BU4 (Figs. 4, 6). The site was initially recorded in 1967 (Eoff and Johnson 1968:12-18). Test excavations in 1968 revealed extensive, buried Late Archaic deposits, which were subsequently investigated by large-scale excavation in 1969, 1970 and 1971 (Grosser 1970, 1973, 1977). A final series of large-scale excavations occurred in 1978 and 1979 (Leaf 1981; Haury and Leaf 1983), terminating with backhoe exploration in 1981 (Haury and Leaf 1981).

The site is situated on the first terrace of the Walnut River. On the locality map (Fig. 6), the terrace edge is roughly coincident with the 1300 foot contour. A distinct scarp of 1.5-2.0 m (4-6 feet) is observed near the river, northeast of the 1978-1979 excavation block (Fig. 6). An extensive Plains Woodland surface scatter covers approximately 6.8 ha of terrace surface, bordered by the Walnut River on the east and south, and by an unnamed tributary to the west. A light

surface scatter also occurs on a sloping surface west of the tributary (Leaf 1981).

The site occupies an area mapped as Vanoss series by Penner et al. (1975). The outline of the Vanoss soil area (Fig. 4), is about the same as the prehistoric surface scatter (Fig. 6). The site is dissected by a broad shallow swale, delimited in Fig. 6 by the 1302 and 1304 foot contours, which represent an abandoned channel of the Walnut River. The densest concentration of surface material occurs south of the swale. All major archaeological excavations were located in this portion of the site (Fig. 6).

Detailed observations of soil stratigraphy were made during the 1978-1979 investigations (Leaf 1981; Haury and Leaf 1983). Soil samples were taken from a pedon in the north wall of the 1978-1979 excavation block. This pedon is described as follows:

- Ap 0-20 cm. Dark grayish brown (10YR4/2) silt loam; granular structure; slightly acid; smooth, abrupt boundary.
- A 20-25 cm. Dark grayish brown (10YR4/2) silt loam; granular structure; slightly acid; smooth, gradual boundary.
- Bt 25-50 cm. Dark yellowish brown (10YR4/6) silty clay loam; blocky structure, common fine mottles due to dark coatings on many peds; few, discontinuous clay skins in open pores and root channels; neutral; smooth, gradual boundary.
- Ab 50-75 cm. Very dark grayish brown (10YR3/2) silt loam to silty clay loam; blocky structure; more friable than overlying horizon; common, fine mottles; neutral; smooth, gradual boundary.
- ACb 75-140 cm. Dark brown (10YR3/3) silty clay loam; blocky structure; fine, common mottles; neutral.

The lateral extent of this sequence was observed in seven backhoe trenches (Fig. 6) excavated in 1981 (Haury and Leaf 1981). Trenches 1 and 2, 40 and 30 m long, respectively, form a "T" whose intersection is 10 m west of the northwest corner of the 1978/79 block. Trenches

3 to 7, ranging from 8 to 10 m in length, were excavated along an east-west line 40 m north of the block. A composite "fence" diagram of trench profiles is presented in Fig. 9.

Soil stratigraphy similar to that in the 1978-79 block was observed in Trench 1, the eastern two thirds of Trench 2, Trench 6 and Trench 7. In the backhoe trenches, the buried A horizon becomes thicker in all directions from the intersection of Trenches 1 and 2. At the intersection, the Ab is 20 cm thick. It thickens to 50 cm in the 1978/79 block; 33 cm in the western part of Trench 2; 40 cm in Trench 7 to the north, and 49 cm in Trench 1, to the south (Fig. 9).

In the western end of Trench 2, the buried soil dips sharply to the west. Lying unconformably on the buried A horizon in the last 10 m of Trench 2 is a wedge-shaped deposit of laminated silt which increases in thickness toward the west (Fig. 9). The lamina consist of layers of fine, pale brown sand and darker silts, ranging from 1 to 3 cm thick. The upper boundaries of the laminated deposit are blurred by bioturbation and soil formation associated with the present surface. The wedge probably represents lateral accretion deposits of the intermittent tributary to the west.

Trenches 3 and 6 transect the shallow paleochannel. The trenches, although 8 to 10 m long, "had only the easternmost 2.0 m dug to 2.8 m deep after which the western 6.0 m was ramped" to provide safe access to the deep end (Haury and Leaf 1981:2). Thus, continuous profiles along this transect were not obtained. Fig. 9 suggests a plausible but unconfirmed correlation of separate profiles, as explained below.

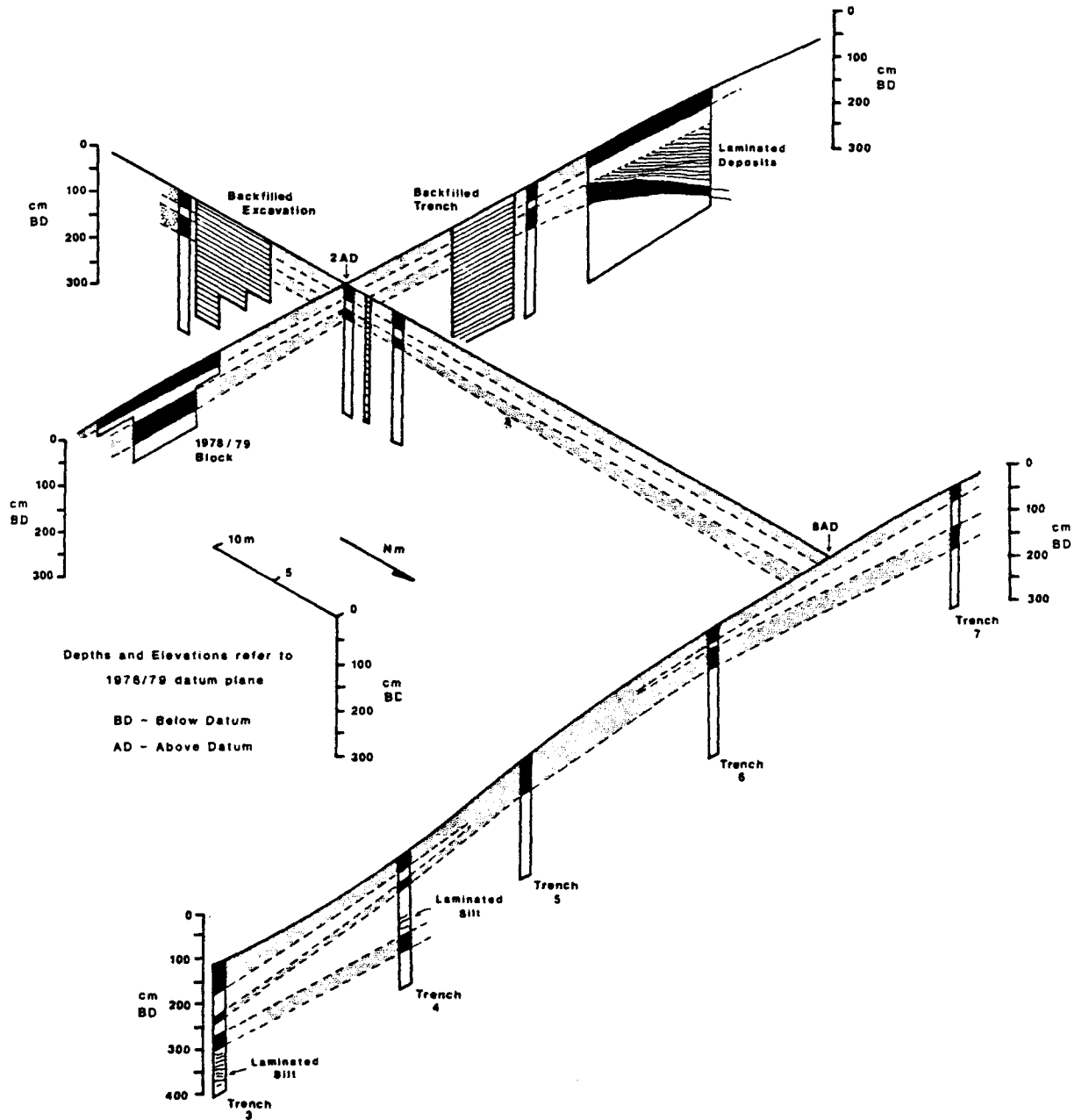


Fig. 9. Composite stratigraphic profiles of backhoe trenches at 14BU9.

Trench 6, at the edge of the channel scar, revealed the two-soil sequence observed in Trenches 1, 2 and 7. In Trench 5, excavated on the slope of the channel scar, only one soil was observed. Archaeological evidence indicates that this soil is an unburied or exhumed exposure of the soil buried beneath the surfaces to the south and west. As discussed below, archaeological excavations recovered projectile points of Late Archaic age only from the buried soil. On the surface of the site, Archaic artifacts occur on the west slope of the paleo-channel (Leaf 1981:176). Given these observations, Late Archaic artifacts deposited in the buried soil apparently crop out on the margins of the channel.

Trenches 3 and 4, excavated in the bottom of the channel scar, reveal three soils (Fig. 9). The buried A horizons are only slightly darker than the overlying and underlying C horizons, suggesting that soil development reflected by them was neither long-lived nor intense. Above the lower soil in Trench 3 and below the lower soil in Trench 4, the C horizons exhibit thin, bioturbated laminations. It is tentatively suggested (Fig. 9) that the upper two A horizons merge into a single horizon between Trench 4 and Trench 5. Whether the lower soil in Trench 4 also slopes upward to the surface at Trench 5 is not known. A third soil was detected in no other unit at the site, although it is possible that excavations did not extend deep enough to encounter it.

Excavations at 14BU9 in 1968-1969 (Grosser 1970) and 1970-1971 (Grosser 1977) encountered soil stratigraphy similar to that observed in the 1979 block excavation and Trenches 1, 2, 5 and 7. The locations

of these excavations were unfortunately not mapped with reference to permanent features such as treelines or fences. Field notes and photographs recording such data are either inadequate or non-existent. However, Trench 1 transected an old excavation unit (Fig. 9) which, based on comparison with published profiles (Grosser 1973), probably represents Grosser's 1969 or 1970 excavation. If this interpretation is correct, then units excavated in 1968, 1969 and 1971 are located west of Trench 1 and south of Trench 2.

Trenches 1 and 2 also encountered a backfilled backhoe trench (Fig. 9). Trench 2 struck the previous trench at an oblique angle, so that it appeared in the trench walls as a 5 m wide scar. Trench 1 cut nearly perpendicularly across the previous trench. The trench may represent one which Grosser (1973, 1977) reports excavating, although, again, the location of the trench was not recorded.

The upper 40 cm of the buried soil at 14BU9 yielded small, corner-notched triangular points and larger, stemmed and corner-notched points. Similar in general to the surficial Plains Woodland material from the site, Grosser (1973:232) considered the buried assemblage to represent a discrete Late Archaic complex based on the absence of ceramics and the component's stratigraphic separation from Woodland levels. Materials from the upper levels of the buried soil were designated the Walnut phase. Two charcoal samples from these levels yielded radiocarbon dates of 2060 ± 80 B.P. and 1970 ± 110 B.P. The former date (N-1276; Yamasaki et al. 1974:340) was recovered at a depth of 45 cm, while the latter (N-769; Yamasaki et al. 1972:229) represents a composite

sample of charcoal collected between 40 and 55 cm BS (below surface).

Beneath the Walnut phase levels of the buried soil, different forms of projectile points occurred. Modified-notching (White 1968: 58-59) is the predominant mode of stem preparation; side-notching and corner-notching is also present. Concave, convex and straight bases are represented. Blade elements range from long and lanceolate to short and triangular. This point group occurred between 80 and 140 cm below surface (40 to 100 cm within the buried soil). Four radiocarbon dates were obtained from charcoal at these levels: 3240 ± 85 (N-1277); 3650 ± 140 (N-770); 3910 ± 160 (N-771); and 3980 ± 100 (N-1278) (Yamasaki et al. 1972, 1974). Charcoal samples were recovered at depths ranging from 84 to 100 cm below surface. No information on sample size or recovery context is available for these dates with the exception of N-770, which was recovered from a basin-shaped refuse pit (Grosser 1973:233).

The third cultural-stratigraphic unit recognized in the buried soil, designated the Chelsea phase, was defined on the basis of projectile points characterized by broad, thick stems formed by modified notching, with straight bases and broad, short blades. Four such points were recovered at depths of 140 to 180 cm beneath the site surface. A charcoal sample from a depth of 178 cm returned a radiocarbon date of 4830 ± 150 B.P. (N-1279), but Grosser (1977:81) questioned the validity of the date, since it conflicts with dates obtained from deeper levels of the site. For present purposes, the temporal placement of the Chelsea phase is referred to as pre-4000 B.P.

The presence of a fourth cultural level at 14BU9 is suggested by artifacts observed in a backhoe trench at a depth of approximately 2.5 m below surface. The 2.5 m deep component is reported to have contained "burned limestone concentrations" (Grosser 1973:228) and a "grinding stone and chipped stone" (Yamasaki et al. 1974). Two charcoal samples from the 2.5 m level yielded dates of 4600 ± 125 B.P. (N-1286, Yamasaki et al. 1974) and 4150 ± 110 B.P. (N-1551, Yamasaki et al. 1977). Grosser (1977:81) considered these dates to be stratigraphically inverted with the date of 4830 ± 105 mentioned above and questioned their validity. These ambiguities are further discussed in Chapter 5.

To summarize, 14BU9 is a multi-component site situated on T-1 of the Walnut River. Soil stratigraphy consists of a moderately developed surface soil overlying a weakly developed buried soil. This soil sequence, as observed in archaeological excavations and backhoe trenches, is co-extensive with an area in which Penner et al. (1975) map Vanoss soils. While the A and B horizons of the surface soil are similar to the Vanoss series, the A-C horizonation of the underlying soil is more similar to the Verdigris series. To the west, the buried soil drops sharply toward an intermittent tributary at the edge of the valley, and is overlain by laminated deposits of the stream. To the east, the buried soil crops out on the edges of a Walnut River paleochannel. In the bottom of the channel, the soil splits into two and possibly three weakly developed A horizons, separated by laminated deposits.

Plains Woodland material was recovered from the site surface and the surface A horizon. The buried soil contained Late Archaic occupational strata, subdivided into 3 cultural stratigraphic units, each

associated with distinct projectile point morphologies. Radiocarbon dates of ca. 2000 B.P. were associated with small, corner-notched points of the Walnut phase, in the upper 40 cm of the buried soil. Dates of 3000-4000 B.P. were associated with stemmed points of the El Dorado phase, recovered from 40 to 110 cm within the buried soil. A pre-4000 B.P. age is assigned to points recovered from the Chelsea phase levels, 140-180 cm within the buried soil. The cultural identity and age of deposits observed 2.5 m below the site surface cannot be determined with present data.

14BU25*

Site 14BU25 is located on Durechen Creek, 4.8 km upstream from the creek's confluence with the Walnut River. The site was recorded during the 1967 survey (Eoff and Johnson 1968:36). Test excavations were conducted in 1968 and 1973, and large scale excavations were carried out in 1969 and 1970. Test excavations were again conducted in 1979. Excavations at the site are reported in detail by Root (1981b).

The site is situated at the southern edge of Durechen Creek valley (Fig. 4). The creek in the site vicinity is entrenched 2.5-3.0 m (8-10 feet), as shown in Fig. 10. Although scarps separating former and present floodplain surfaces are not clearly visible in the site vicinity, the break between the valley floor and the channel trench occurs at approximately 1350-1352 feet in Fig. 10.

A broad swale, interpreted as an abandoned channel of Durechen Creek, occurs about midway between the valley edge and Durechen Creek. Although there is little difference in elevation between the surfaces

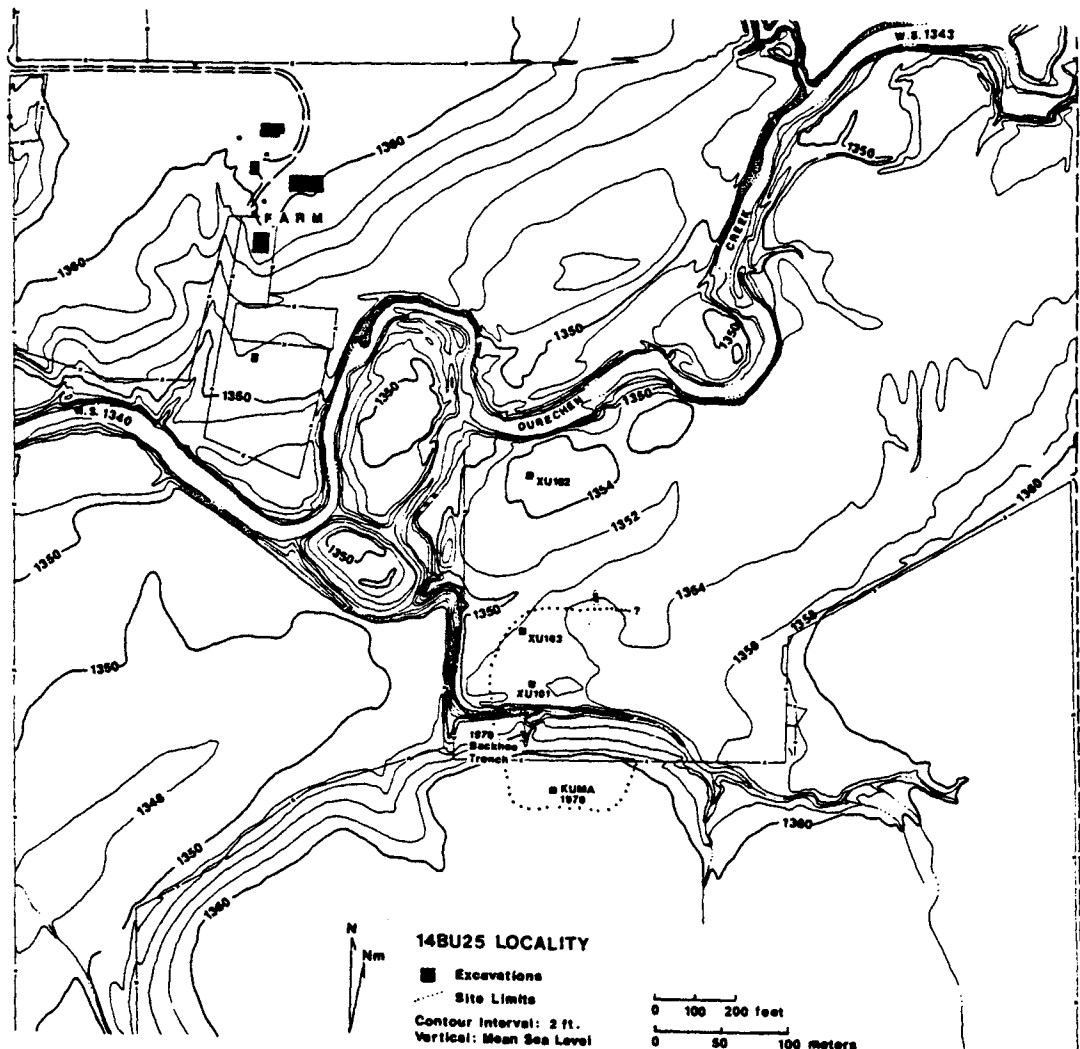


Fig. 10. Topographic map of the 14BU25 locality. Source: U.S. Army Corps of Engineers, 1:200 topographic maps.

north and south of the channel scar, significant differences in soil stratigraphy are noted.

In the study area soils map (Fig. 4), the abandoned channel is mapped as an area of Brewer series. Surfaces between the channel and the present day creek are mapped as Verdigris series, while surfaces between the channel and the valley edge are mapped as Vanoss series. The Vanoss series area appears in Fig. 4 as a long, linear unit which extends for 3.3 km along the south edge of Durechen Creek valley. Since the surface rises gradually toward the valley edge, the location of this soil area corresponds to a valley-edge footslope.

An intermittent tributary,¹ originating in a small sidevalley southeast of the site, crosses 14BU25 from east to west, forming a deep, narrow gully parallel to the valley edge. West of the site, the gully makes an abrupt turn northward to flow toward Durechen Creek. As shown in Fig. 10, the suspiciously right-angle shift in course coincides with a barbed wire fence, suggesting that this portion of the gully's course was established in post-settlement times. The lack of evidence of former stream channels west of the fence suggests that incision of the gully as a whole may be a post-settlement phenomenon.

Site 14BU25 consists of a buried Late Archaic component, which was initially discovered in the banks and floors of the valley-edge gully. Based on data from excavations, the site covers an area at least 100 m in diameter, extending from the southern edge of the channel scar to the upland surfaces beyond the gully.

Soil stratigraphy and cultural stratigraphy were recorded during block excavations of 1969-1970 and in the 1979 test excavations. The

location of the three, 2 x 2 m test units excavated in 1979 are shown in Fig. 10. The 1969-1970 excavation centered on and around the gully. A 110 m² area was excavated. A backhoe trench was also dug from an origin near the southwestern corner of the block, south to the valley edge (Fig. 10).

Soil stratigraphy north of the mid-valley channel scar, as observed in XU102, was significantly different from that observed in excavations south of the scar. XU102 was a 2 x 2 m test unit excavated to determine if archaeological deposits extended north of the channel scar. No such deposits were detected, but the excavation profile was mapped and described (Root 1981b:30-33). No sediment samples were taken, but Munsell color determinations and textural estimates were made. As shown in Fig. 11, a 15 cm thick, dark grayish brown (10YR4/2) plowzone overlies approximately 50 cm of structureless, dark brown (10YR3/3) silty clay. Beneath this unit, to a depth of 117 cm, bands of yellowish brown (10YR5/4) clay loam alternate with slightly thicker layers of dark brown (10YR3/3) silty clay. Between 117 and 126 cm, the alternating layers of silty clays (finer textured) and clay loams (coarser textured) are individually thinner and more closely spaced. Beneath 126 cm, a structureless dark brown (10YR4/3) silty clay extends to at least 165 cm, the base of the excavation.

The XU102 profile represents vertical accretion in an active depositional environment. Soil development is apparently restricted to organic darkening of the upper 15 cm; the absence of soil structure suggests that the underlying sediments constitute alluvium that has

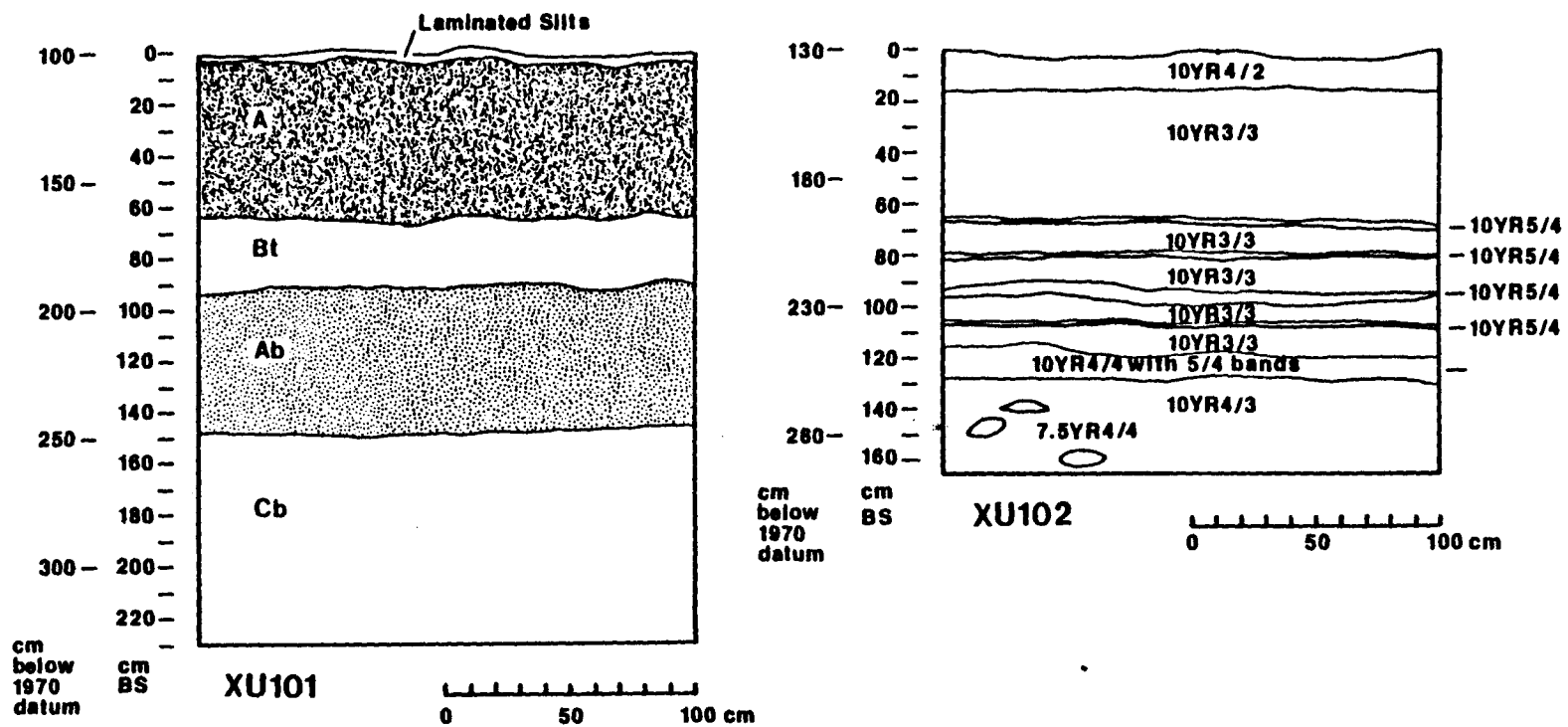


Fig. 11. Stratigraphic profiles of XU101 and XU102 at 14BU25 (after Root 1981b: Figs. 4-5).

been relatively unaltered by pedogenesis. The sedimentary strata consist of alternating layers of clay loam and finer-textured (less sandy) silty clay. Each pair of layers probably was formed during a single flood episode, the coarser clay loam strata deposited in earlier, high energy stages of the episode, and the finer silty clay strata deposited in the waning stages as energy levels dropped.

The absence of soil development and the presence of sedimentary structures in XU102 contrasts markedly with soil stratigraphy observed south of the mid-valley channel scar. Soil stratigraphy recorded in XU101 (Fig. 11) and the 1969-1970 block (Fig. 12), as well as XU103, not illustrated here, consists of two soils: a moderately developed surface soil overlying a weakly developed buried soil.

A detailed description of this sequence was made in XU101, described as follows:

- A 0-30 cm. Black (10YR2.5/1) clay loam; subangular blocky structure; medium to slightly acid; gradual, smooth boundary; upper 5 cm contain fine, wavy, discontinuous laminae of grayish silt.
- B1 30-70 cm. Black (10YR2.5/1) clay loam; blocky structure; slightly acid to neutral; gradual, smooth boundary.
- Bt 70-90 cm. Dark brown (10YR3/3) silty clay; blocky structure; neutral; gradual, smooth boundary.
- Ab 90-145 cm. Very dark grayish brown (10YR3/2) silty clay loam; blocky structure; few, fine mottles of lighter color; neutral; smooth, diffuse boundary.
- Cb 145-230 cm. Dark grayish brown (10YR4/2) grading to grayish brown (10YR5/2) silty clay loam grading to clay loam; structure not recorded; abundant, fine particles of iron and manganese oxides; many, small alluvial cobbles in the lower 30 cm; neutral.

This soil sequence is present from the margins of the channel scar to the valley edge with only minor variations in horizon thickness. The upper boundaries of the surface soil and the buried soil slope

upward toward the valley edge, although the surface slope is slightly greater than that of the buried soil. Between the gully edge and the end of the 1970 backhoe trench (32.5 m), the present day surface rises approximately 1.35 m, while the top of the Ab rises only 0.96 m.

In the south end of the backhoe trench, the buried A horizon lies directly over bedrock. The Ab horizon presumably merges with the surface A horizon farther south to form the upland surface soil. A test pit excavated in the uplands in 1970, approximately 50 m south of the main excavation, encountered a single soil, 80 cm thick, overlying bedrock (Root 1981b:36).

A single cultural stratum was encountered at 14BU25. In the 1969-70 excavations at 14BU25, over 10,000 artifacts were three-dimensionally plotted (Root 1981b:35). The location of only a few artifact categories (complete flakes, chunks and shatter, cores, flake tools, retouched tools and limestone features) in Fig. 12 is sufficient to show the general southward dip of the stratum, from approximately 275 cm below datum north of the gully to 325 cm below datum south of the gully.

In XU101, 20 m north of the gully, cultural material was distributed over a vertical range of 110 cm, although approximately 80% of the material occurred in a 30 cm span between 150 and 180 cm below surface (Root 1981b:30). This corresponds to 250-280 cm below the 1970 datum, placing the cultural stratum in XU101 at an equivalent level with the northernmost materials recovered from the block excavations.

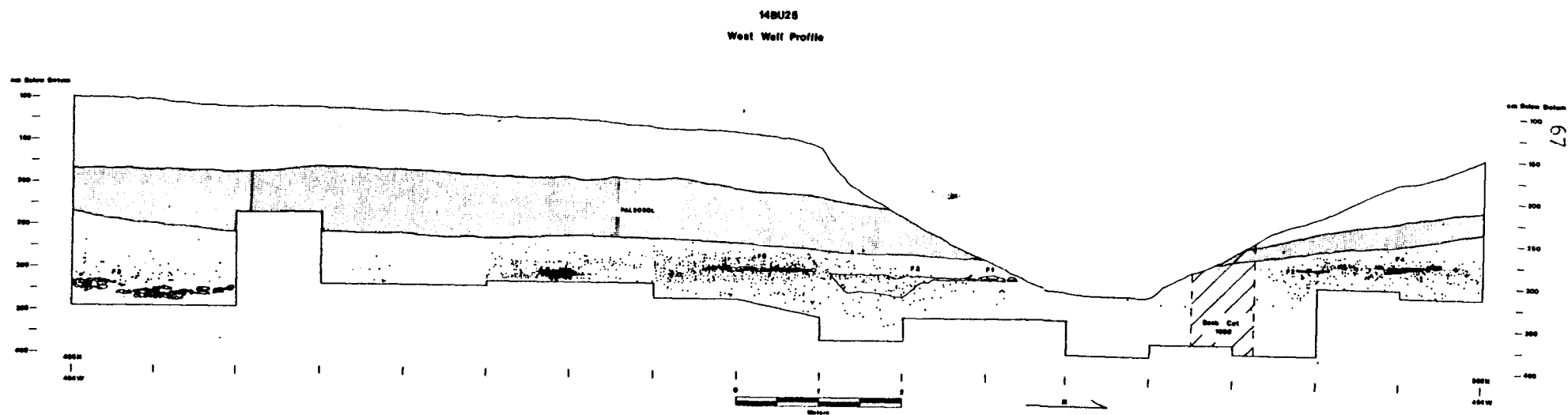


Fig. 12. Stratigraphic profile of the 1969-1970 block excavation at 14BU25
(from Root 1981b:Fig. 6).

XU103, 60 m north of the gully, yielded only 8 artifacts, seven of which were recovered from depths of 140-170 cm below surface (234-304 cm below 1970 datum) (Root 1981b:34). This suggests that the archaeological stratum in XU103 is slightly lower relative to the 1970 datum than in XU101, although the small sample size precludes definitive conclusions.

Artifact distributions suggest that prehistoric occupations occurred on a land surface which, near the valley edge, sloped downward to form a shallow depression. The land surface north of this depression was nearly level, perhaps dipping slightly in the vicinity of XU103. This surface is identified only by artifact distributions; no sedimentary structures or pedological features indicate its presence. The former land surface does not parallel that beneath which either the buried soil or the surface soil formed, since the cultural stratum becomes increasingly deeper, while the A and Ab horizons slope upward at the valley edge.

The archaeological stratum yielded a radiocarbon date of 4435 ± 100 B.P. (UGa-2806), obtained from a 50.6 g sample of charcoal, comprised of small pieces recovered from Feature 5, a limestone cooking facility (Root 1981b:22, 185). The stratigraphic position of the 14BU25 deposit at the base of, and below the Ab horizon of a buried soil is similar to that of the Chelsea phase levels at 14BU4 and 14BU9.

A variety of projectile point forms are represented at 14BU25. A dominant mode of stem preparation was achieved by deep, basal and corner-notching, producing tools with pronounced shoulder barbs (Root

1981b:106). Modified-notch, side-notched and unnotched forms were also recovered. The outline morphology of several modified-notched points are comparable to points from the Chelsea phase levels at 14BU4 and 14BU9, although there is sufficient variation in haft element morphology to reject a hypothesis of direct technological continuity between the sites (Root 1981b:212).

In summary, 14BU25 is a single component, Late Archaic site located at the edge of Durechen Creek valley. Cultural materials occur immediately beneath the Ab horizon of a weakly developed buried soil, which is overlain by a moderately developed surface soil. While the present day surface slopes upward toward the valley edge, the Late Archaic archaeological deposit slopes downward toward the valley edge. The site, dated at 4435 ± 100 B.P., and yields artifacts comparable to those from Chelsea phase deposits at 14BU4 and 14BU9.

The site is associated with a valley-edge footslope, mapped by Penner et al. (1975) as an area of Vanoss soil. The Vanoss series exhibits a moderately developed profile similar to the surface soil observed at 14BU25. The site is separated from the present-day course of Durechen Creek by an abandoned channel. The buried soil observed south of this channel scar is not present north of the scar, where fluvial sediments, relatively unmodified by pedogenesis, are observed.

14BU81

Site 14BU81 is situated on a broad expanse of the Walnut River first terrace. The site is located 730 m northwest of the confluence of the Walnut River and Durechen Creek, and about 400 m east of the

Walnut itself. Initially recorded by a KU survey crew in 1975, little was known of the site until test excavations, conducted in 1978, documented two Late Archaic components (Artz 1981). Backhoe testing was conducted in 1981 (Haury and Leaf 1981).

The locality map (Fig. 13) shows little relief in the immediate vicinity of the site. The 1320 foot contour marks the approximate edge of T-1 due west of 14BU81. Six hundred meters northeast of 14BU81, the T-2 scarp rises 2.5 m (8 feet) above the T-1 surface.

Soils on the relatively flat T-2 surface are mapped as the Norge series (Fig. 4). Those on the T-1 surface are mapped as the Verdigris series, except for a sinuous body of Brewer soil in the immediate vicinity of 14BU81. This soil area appears in Fig. 13 as a depression curving from west to north in the site vicinity, outlined by the 1322 fc contour. North of 14BU81, the Brewer soil body is not defined by surface relief features.

In 1978, a large borrow pit, two to three meters deep, was excavated immediately north of 14BU81. This pit extended from the Walnut River to the T-2 scarp and from the county road bordering 14BU81 to a fence-line 250 m to the north. Where the southwall of the borrow pit intersected the Brewer soil area, two channel fill deposits were exposed. These features were approximately 20 m wide and 2.5 m deep, as determined by pacing and hand level measurement. Based on the sinuous, arcuate outlines of the Brewer soil area, and its association with topographic depressions and channel fill deposits, the Brewer soil body is interpreted as a complex of abandoned channels of the Walnut River.

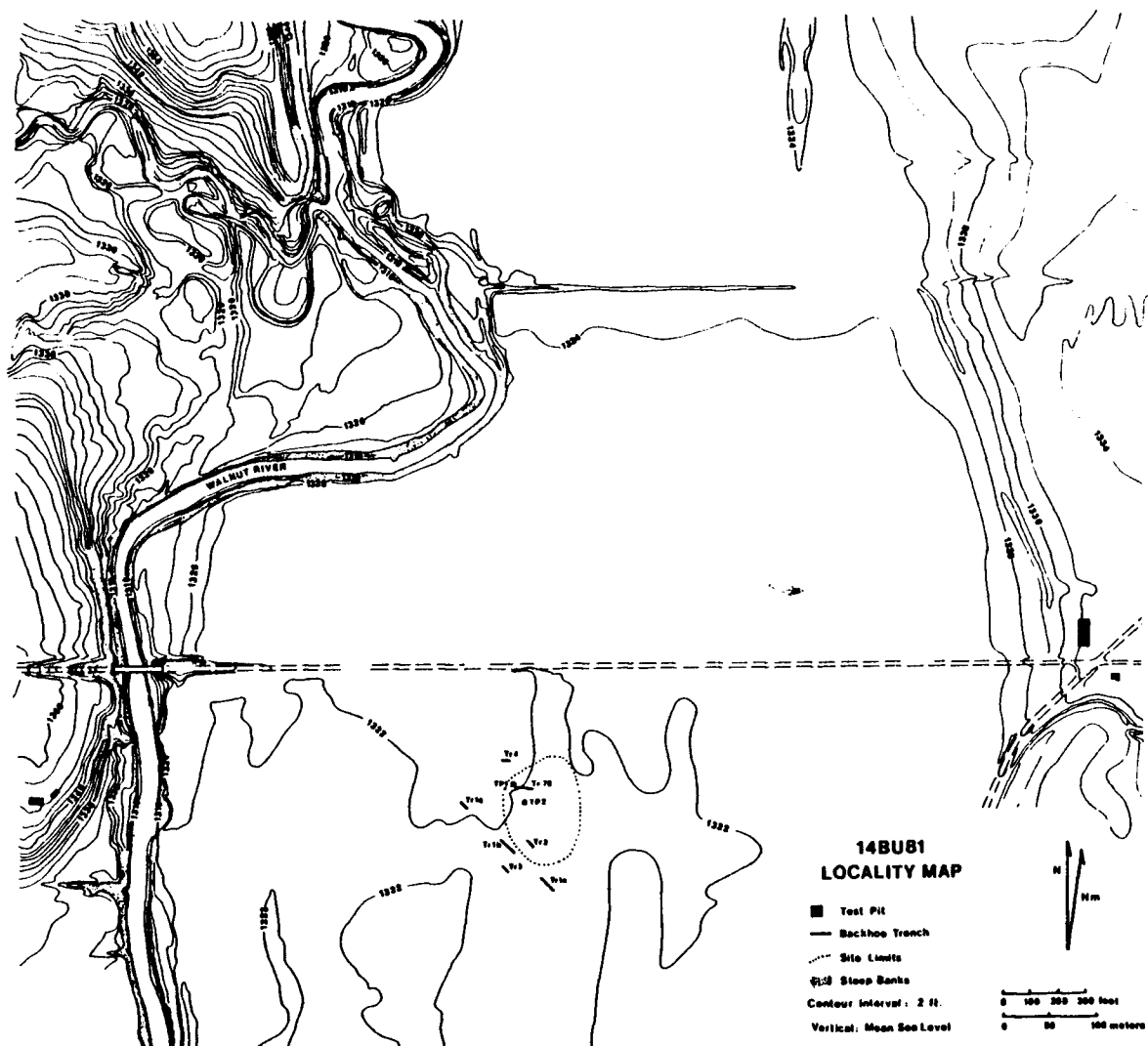


Fig. 13. Topographic map of the 14BU81 locality. Source: U.S. Army Corps of Engineers, 1:200 topographic maps.

Although 14BU81 lies partly within the paleochannel complex, the excavated areas of the site lie largely at the outer margins of the soil body (Fig. 13). The following profile describes the soil observed in TP 2:

- Ap 0-20 cm. Very dark grayish brown (10YR3/2) silty clay loam; granular structure; medium acid; smooth, abrupt boundary.
- A 18-55 cm. Very dark grayish brown (10YR3/2) silty clay loam; subangular blocky structure; slightly acid; smooth, gradual boundary.
- AC 55-155 cm. Dark brown (10YR3/3) grading with depth to dark yellowish brown (10YR4/4) silty clay loam; blocky structure; slightly acid to neutral.

Soil stratigraphy in TP 1, which was excavated to a depth of only 1 m, is identified to TP 2. Similar stratigraphy was observed in the 1978 backhoe trench and 1981 Trench 4. The simple, A-C profile, similar to the Verdigris series was extensively exposed in the walls of the borrow pit north of the site. In the borrow pit exposures, the A-C horizonation is interrupted only by the darker fill of the two channel fill deposits previously mentioned.

Although soil stratigraphy in TP1 and TP2 revealed no evidence of the abandoned channel, cultural stratigraphy indicated its presence. Fig. 14 shows the number of artifacts recovered from the three excavations at 14BU81. To account for the varying thicknesses of excavation units in the two test pits, artifact densities are reported as an average number per 1 m x 1 m x 0.1 m volume of soil. The artifacts recovered from the backhoe trench were recovered from a four-meter long section near the west end of the trench, only a few meters from TP1. Actual artifact counts are shown for the trench in Fig. 14.

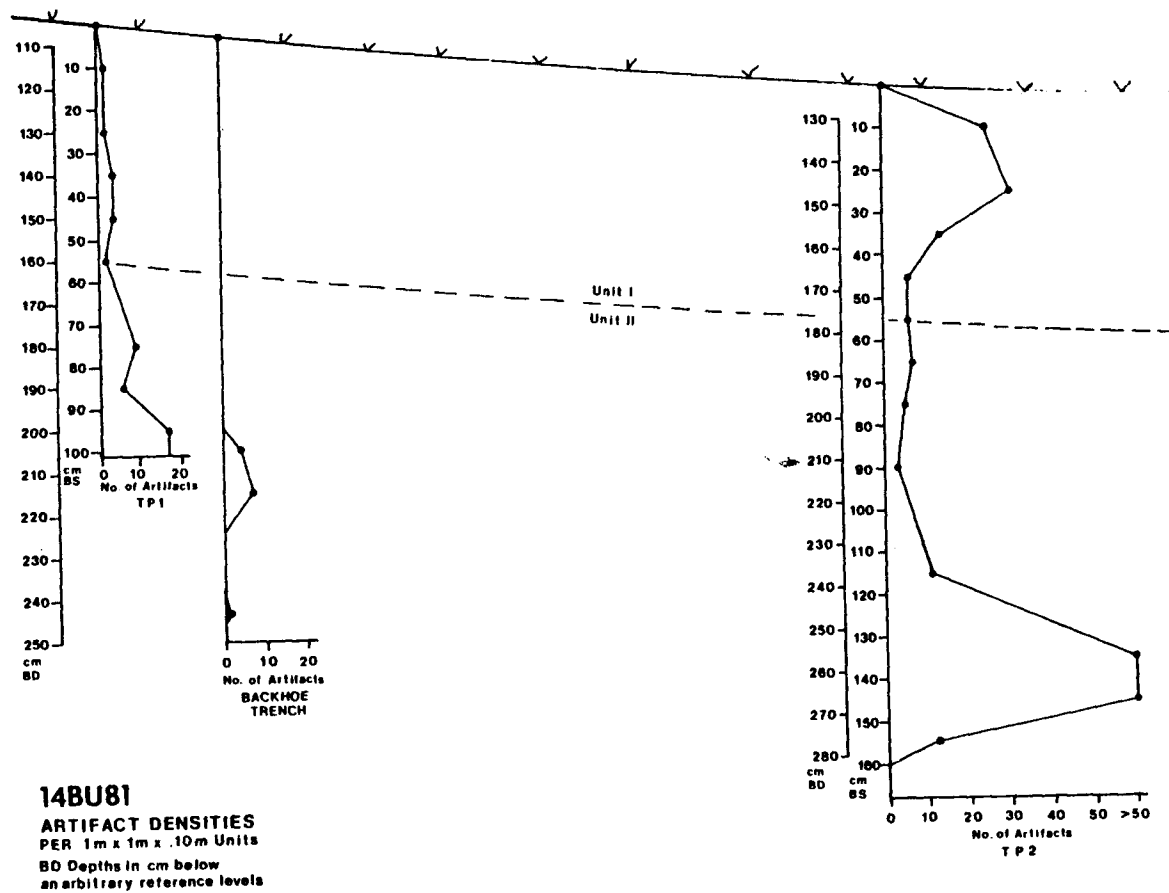


Fig. 14. Depth distribution of artifacts in excavations at 14BU81.

Fig. 14 shows bimodal vertical distributions of artifacts. In TP2, an upper component is marked by a peak in artifact frequency at 20-30 cm BS, while a second, much denser concentration occurs at 130-150 cm. In TP1, although the low density of cultural material makes trends difficult to discern, an increase in artifact frequency is suggested between 30 and 50 cm BS. A stronger mode is present at 90-100 cm and seems to represent the same cultural residue observed in the backhoe trench at equivalent depths.

Artifact comparisons suggest that, despite variation in depth below surface, only 2 components are represented at the site. The upper component, represented by the upper modal densities in Fig. 14 yielded stemmed projectile points comparable to those from the El Dorado phase deposits at 14BU9, dated between 3000 and 4000 B.P. Points assigned to this unit were recovered from 46 cm BS in TP1, 15 cm BS in TP2 and from the ground surface in the vicinity of TP2. The site surface also yielded two small, corner-notched points similar to those recovered from Walnut phase levels at 14BU9 (ca. 2000 B.P.).

Fig. 14 shows a 20 cm disparity in below surface depths of the upper component modal densities between TP1 and TP2. This disparity, however, is cancelled out by a 20 cm loss in surface elevation between the two test pits. The density modes in both TP1 and TP2 occur at depths of 140-150 cm below an arbitrary datum. Thus, the El Dorado phase deposit exposed on the surface near TP2 is buried half a meter in TP1.

Although no diagnostic artifacts were recovered from the basal components in TP1 and TP2, artifact samples from each pit at these

levels exhibit functional homogeneity. Artifacts from the basal levels of both test pits, as well as the backhoe trench consisted almost exclusively of thermally altered chipping debris of Florence A chert. In the lower deposits of TP2 and the trench, most prehistoric materials were concentrated in a 20 cm thick stratum. If these actually represent the same depositional surface, then the surface slopes in the direction of TP2.

Such a slope is consistent with the location of the test pits on the margins of an abandoned stream channel. The earlier, pre-El Dorado phase occupation of the site appears to have occurred on the sloping surface of the abandoned channel margin. The later, El Dorado phase occupations took place after the channel had largely silted in. As suggested by Fig. 14, the El Dorado phase occupation occurred on a nearly level surface. Subsequent erosion re-exposed the deposit in certain areas, leaving it buried in others.

In summary, 14BU81 is associated with a complex of paleochannels situated on the first terrace of the Walnut River. Although the simple A-C soils at the site show no stratigraphic evidence of the former channel's presence, artifact distributions suggest the existence of former land surfaces associated with the channel. The site's oldest component, estimated to be contemporaneous with Chelsea phase levels at 14BU4 and 14BU9, occurs on a surface which slopes down into the channel. The site's more recent El Dorado and Walnut phase occupations occurred on a level surface, representing the nearly silted-in channel. Exposures in a borrow pit north of the site confirm the existence of

the paleochannel, and demonstrate that the weakly developed soil observed at the site is widespread beneath the first terrace surface in the site locality.

14BU87

Site 14BU87 is situated on the first terrace of Durechen Creek, 2.9 km upstream from the creek's confluence with the Walnut River and 2.4 km downstream from 14BU25, previously discussed. The site was recorded by the 1977 survey (Root 1979:46-47) and was tested in 1978 (Artz 1981).

As shown in the locality map (Fig. 15), 14BU87 is situated between Durechen Creek to the north and an intermittent tributary to the east. The two streams converge immediately east of the site. The terrain in the vicinity of the site is dissected by remnants of previous stream channels. Unfortunately, due to the construction of a 3 m high railroad embankment through the locality, the channel patterns are difficult to trace.

Stratigraphy associated with the channel remnant which crosses 14BU87 was revealed in a 2 x 2 m test pit (TP1) and a 70 m long backhoe trench. The test pit profile (Fig. 16) is described as follows:

- Ap 0-8 cm. Very dark gray (10YR3/1) silty clay loam; granular structure; slightly acid; smooth, abrupt boundary.
- Bt 8-40 cm. Dark yellowish brown (10YR4/4) when moist, slightly redder hue when dry; silty clay loam; blocky structure; slightly acid; smooth, gradual boundary.
- ACb 40-100 cm. Dark brown (10YR4/3) silty clay loam; blocky structure; friable; when dry, ped surfaces coated with thin, continuous whitish films which do not effervesce in HCl. Films occur in bands 4-15 cm in thickness, separated by 6 cm thick bands of dark brown not exhibiting whitish films.

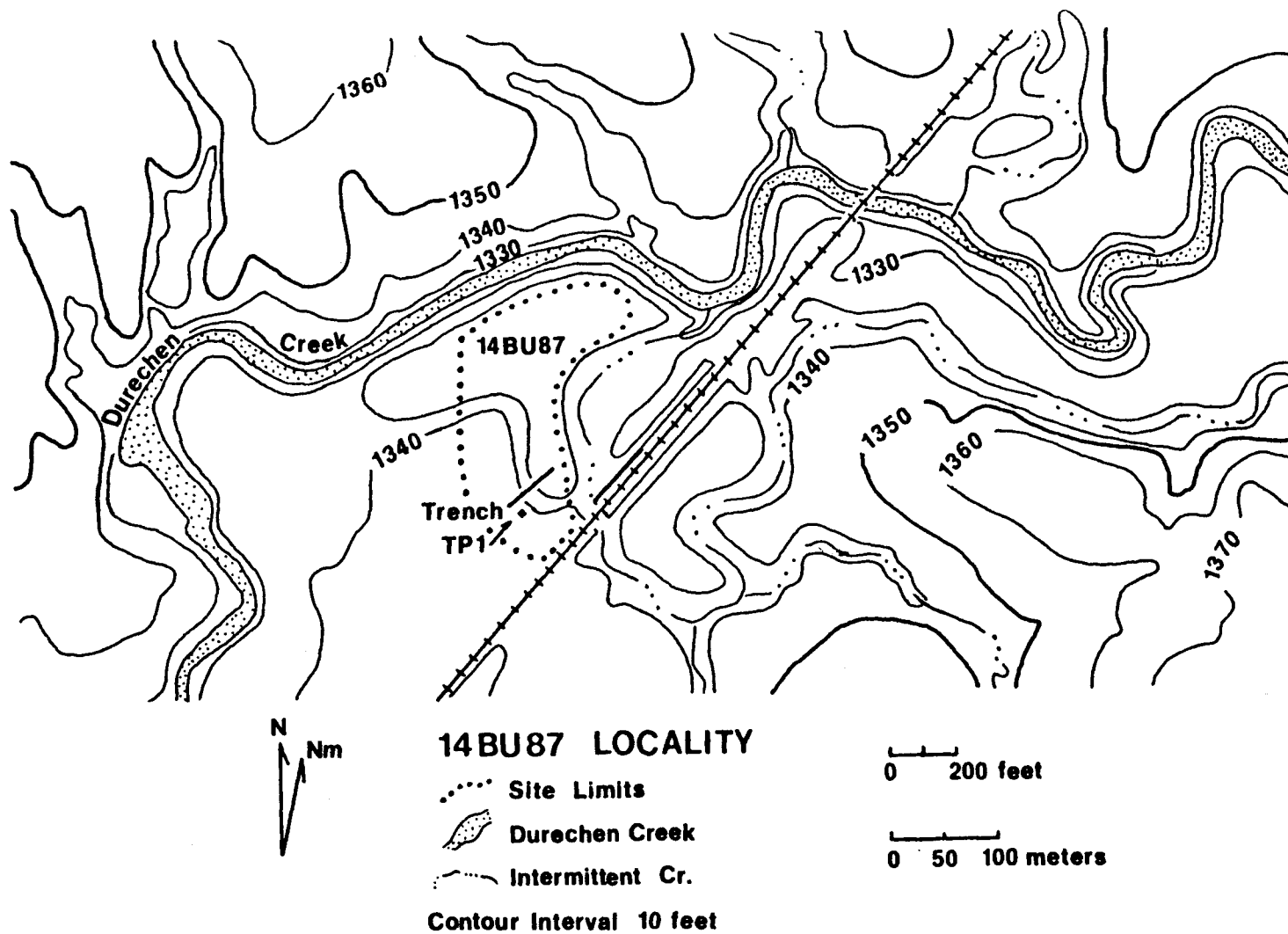


Fig. 15. Map of the 14BU87 locality. Source: U.S. Geological Survey, 1:24,000 topographic map, De Graff quadrangle.

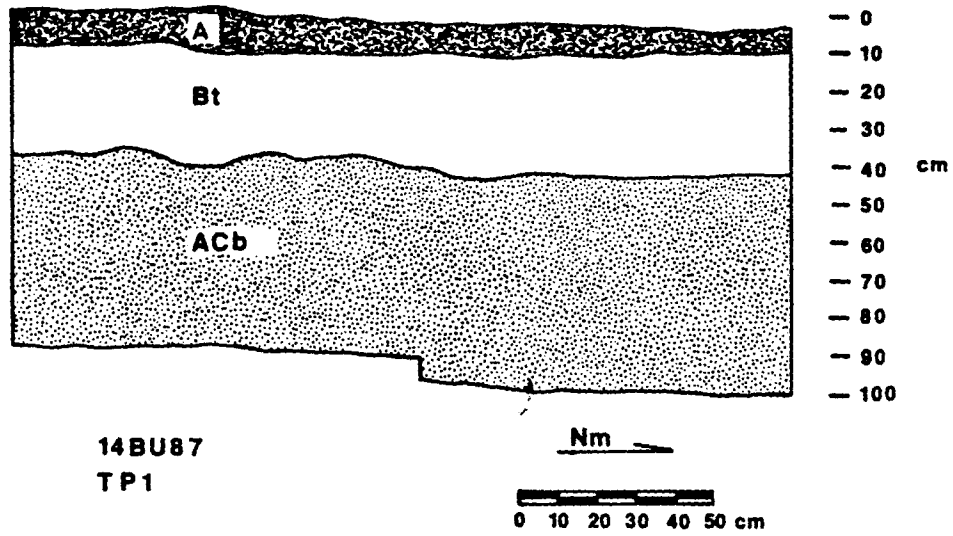


Fig. 16. Stratigraphic profile of TP1 at 14BU87.

TP1 was located on the outer (cutbank) margin of the abandoned channel (Fig. 15). Intensive cultivation of the steep channel slopes has resulted in severe erosion. The only evidence of an A horizon in TP1 is the thin plowzone, which lies unconformably on the clayey subsoil. Horizontal layers in the ACb horizon are probably sedimentary structures, similar to those observed in XU102 near 14BU25 (Fig. 11). The indistinctness of the layers suggests that some pedogenesis has occurred since deposition, blurring the interlayer contacts. Whitish films on ped surfaces not observed in the freshly cut profile, probably represent soluble salts, dissolved in the soil water, which precipitated on the profile surface as the surface dried. Subtle textural differences between the layers, affecting the storage and flow of water through the soil, probably account for the concentration of salts in only a few layers.

Table 4. Distribution of artifacts in TP1, 14BU87, showing chert types and presence/absence of thermal alteration in each level.

Soil Horizon	Depth (cm BS)	Florence A		Florence B		Other		Total		TOTAL
		No Heat	No Heat	No Heat	No Heat	No Heat	No Heat	No Heat	No Heat	
Ap	0-10	17	15	1	1	1	2	19	18	37
	10-20	1	1	-	-	-	-	1	1	2
Bt	20-30	-	-	-	-	-	-	-	-	0
	30-40	-	1	-	1	-	-	-	2	2
ACb	40-50	-	-	-	3	-	1	-	4	4
	50-60	-	-	-	2	-	-	-	2	2
	60-70	-	1	-	3	-	-	-	4	4

Despite the recovery of only 51 artifacts from the test pit, the presence of 2 stratigraphically distinct components is demonstrated. The Ap horizon yielded 37 artifacts (Table 4). Few artifacts were recovered from the Bt horizon; none from the 20-30 cm level. The Ab horizon yielded 10 artifacts (Table 4), as well as small flecks of burned earth and charcoal which were not observed in the Bt horizon.

Although the presence of rodent and root disturbances throughout the excavation suggests the possibility that the ACb materials were introduced from the surface component, this is not supported by the kinds of artifacts recovered (Table 4). Plowzone artifacts are predominantly of Florence A chert, about half of which exhibit the bright red discoloration attributed to thermal alteration. By contrast, the artifacts from the buried soil are predominantly of thermally unaltered, Florence B chert. Despite the small sample sizes, the two groups of artifacts are sufficiently different to support the inference of two components.

The chronological placement of the two components is not clear. No diagnostic artifacts were recovered from the test pit, and the site surface yielded an ambiguous assemblage of projectile points. These included a small, triangular, unnotched projectile point, suggesting a Plains Village affiliation (Root 1979:46-47). Two contracting-stemmed dart points and a corner-notched dart point are like those associated with Plains Woodland as well as Late Archaic Walnut phase occupations in the study area (Fulmer 1976; Grosser 1973:232). The dart points were found on the outer slope of the channel scar, and could

have eroded from a buried soil horizon, or have been deposited subsequently, by later groups. These ambiguities cannot be resolved with present data, although the chronological implications of the stratigraphy in TP1 are discussed in the concluding section of this chapter.

The 70 m backhoe trench originated at a point near TP1, on the cutbank side of the paleochannel, and transected the channel to the opposite, point bar side. As shown in Fig. 17, the first southern 15 m of the trench exposed the same soil stratigraphy observed in TP1. The layered ACb horizon is underlain at approximately 1.5 m BS by alluvium in which no sedimentary structure is visible.

Twenty meters from the south end of the trench, the two-soil sequence is truncated by a massive unit of dark brown silty clay, containing scattered charcoal flecks and chert gravel. Within this unit, underlying the deepest part of the channel remnant is a darker brown, clayier unit, which has a maximum thickness of 1.8 m and a width at the surface of 20 m.

These units are interpreted as channel fill sediments accumulated as the result of the settling of silts and clays suspended in flood-water and rainfall runoff trapped in the channel after its abandonment. The darker colored unit probably reflects greater soil moisture and impeded drainage in the bottom of the channel scar. Soil development in the channel fill unit was restricted to a weak A horizon, 20 cm in thickness at the top of the unit.

Beneath the channel fill deposits, in the northern half of the trench, point bar deposits of the abandoned channel were exposed. These

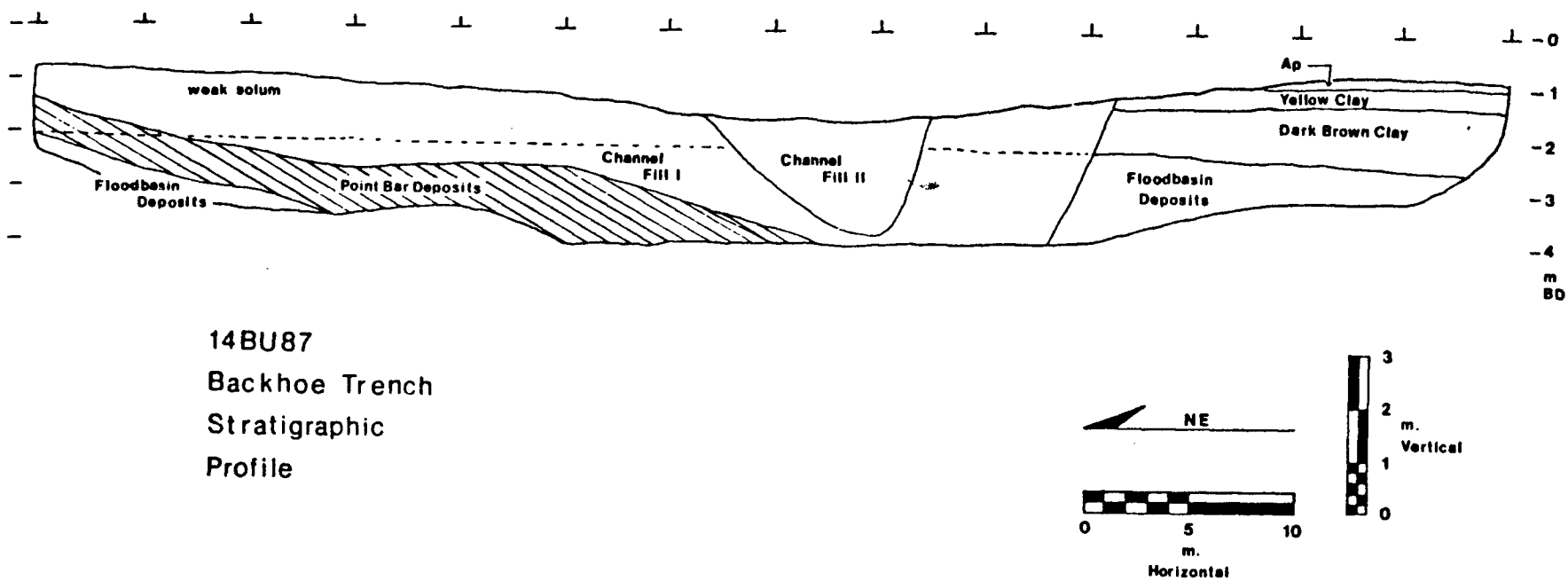


Fig. 17. Stratigraphic profile of backhoe trench at 14BU87.

sediments were not mapped in detail, but consisted of alternating sets of fine and coarse grained bedding, which probably represent the building of the point bar by lateral accretion. In general, the thickest and coarsest sets were observed in the lower portions of the point bar. The layers became thinner with distance north along the trench. In the upper parts of the point bar deposits, within 5 m of the north end of the trench, the laminations within a meter of the surface were indistinct due to pedogenic disturbance. In the north end of the trench, the point bar deposits were underlain by dark brown clays, similar to the alluvium observed in the southern end of the trench.

No dateable organics were recovered from the paleochannel sediments. However, since the channel truncates the two-soil sequence exposed in the test pit, the channel probably formed after the development of the upper soil.

To summarize, a test pit excavated at 14BU87 on the outer margin of an abandoned channel of Durechen Creek reveals two soils: a moderately developed surface soil overlying a weakly developed buried soil. Two archaeological components were detected, one at the surface, and the other in the upper 30 cm of the buried soil. The chronological and cultural placement of the components can not be determined from archaeological evidence alone, although Plains Village and Plains Woodland or Late Archaic Walnut phase occupations are probably represented. Woodland and/or Late Archaic projectile points occur on the south slope of the paleochannel and could have eroded from either the upper or lower soil. Formation of the paleochannel post-dates

both episodes of soil formation evidenced in the test pit, since the channel truncates the two soil sequence.

STRATIGRAPHIC CORRELATION

Correlation is the procedure by which the time-equivalence of spatially separated stratigraphic units is determined (Morrison 1967: 38). The five sites discussed in the preceding sections exhibit similarities in soil stratigraphy and archaeological associations which permit the recognition and correlation of two time-parallel soil-stratigraphic units. Two concepts, the soil-stratigraphic unit and the time-stratigraphic unit, are¹ useful in this regard.

A soil-stratigraphic unit is defined as "a soil with physical features and stratigraphic relations that permits its consistent recognition and mapping as a stratigraphic unit" (American Commission on Stratigraphic Nomenclature 1970:10). Two such units may be defined on the basis of the preceding site descriptions. The first unit consists of the weakly developed, A-C soils which are buried at sites 14BU4, 14BU9, 14BU25 and 14BU87, and exposed at the surface at and around 14BU81. This unit is informally designated the Snyder Soil, since as discussed below, the identification of the unit relies on the radiocarbon sequence from 14BU9, the Snyder site. A second soil-stratigraphic unit consists of a moderately-developed soil which overlies the Snyder Soil at all five sites except 14BU81. This unit is informally designated the Nuttall Soil, after 14BU4, which Fulmer (1976) designated the Nuttall site.

Although soil-stratigraphic units are theoretically defined in terms of pedologic features alone, in practice their definition generally includes time-stratigraphic considerations as well. As defined by Morrison (1967:10), a soil-stratigraphic unit "maintains a consistent stratigraphic relationship to older and younger stratigraphic units with which it is associated, and it is defined and used on the basis of this relationship." Thus, a soil-stratigraphic unit generally coincides with a time-stratigraphic unit that refers to the specific interval of time in which the soil formed (American Commission on Stratigraphic Nomenclature 1970:18).^{*} In the case of soil-stratigraphic units, the corresponding time-stratigraphic unit may be termed the soil-forming interval.

Soil-forming intervals for the two soil-stratigraphic units proposed above are established by archaeological associations. This chronological evidence is summarized in Table 5. The soil forming interval represented by the weakly developed A-C profiles of the Snyder Soil is best dated at 14BU9, which yielded multiple, radiocarbon dated archaeological components. The upper boundary of the buried soil at 14BU9 is associated with dates of ca. 2000 B.P. from the site's Walnut phase deposits. Dates of ca. 3000-4000 B.P. were obtained from El Dorado phase levels, 50-80 cm beneath the top of the buried soil. The Chelsea phase levels, in the Cb horizon at 14BU9, are not radiocarbon dated, but like other levels, yielded a relatively distinct assemblage of projectile points.

Projectile points morphology permits correlation of archaeological deposits and associated soil stratigraphy at other sites with the 14BU9

radiocarbon sequence. The Late Archaic stratum at the base of the ACb horizon at 14BU4 is correlated by projectile point morphology with the Chelsea phase (pre-4000 B.P.) levels at 14BU9. A Late Archaic stratum located in the ACb and Cb horizons at 14BU25 yielded a date of 4435 ± 100 B.P. and was associated with projectile points similar to Chelsea phase artifacts from 14BU4 and 14BU9. The surface and A horizon at 14BU81 yielded artifacts of the Walnut and El Dorado phases, suggesting correlation with the buried A horizon at 14BU9. Although no diagnostic artifacts were recovered from the pre-El Dorado phase component in the C horizon at 14BU81, a Chelsea phase affiliation is posited from stratigraphic and technological comparisons with the Chelsea levels at 14BU4.

Table 5. Correlation of cultural- and soil-stratigraphic units at five archaeological sites in the East Branch Walnut valley.

UNITS	14BU4	14BU9	14BU25	14BU81	14BU87
Nuttal Soil					
Plains Woodland (ca. 1000-1500 B.P.)	X	X	-	-	X
Snyder Soil					
Walnut phase (ca. 2000 B.P.)	-	X	-	X	X(?)
El Dorado phase (3000-4000 B.P.)	-	X	-	X	-
Chelsea phase (pre-4000 B.P.)	X	X	X	X(?)	-
Note: "X" denotes presence of unit at site.					

Similarities in archaeological and natural stratigraphy of the weakly developed soils discussed above suggests the presence of a single soil-stratigraphic unit, the Snyder Soil, which formed synchronously

throughout the East Branch Walnut basin between 2000 and 4000 B.P. Tentatively, the buried soil at 14BU87 is correlated with the Snyder Soil. The archaeological component detected in the upper 30 cm of the buried soil at this site is assumed to be contemporaneous with the Walnut phase, based on stratigraphic comparison with 14BU9. Since there is no independent archaeological evidence to support this assignment for 14BU87, the assignment is speculative.

The Nuttall Soil second soil-stratigraphic unit, is recognized at 14BU4, 14BU9, 14BU25 and 14BU87. At these sites, the weakly developed Snyder Soil was buried by renewed aggradation sometime after 2000 B.P. A moderately developed soil formed in the post-2000 B.P. sediment. Plains Woodland components occur at the surface of this soil at 14BU4, 14BU9 and 14BU87. Since, as discussed in Chapter 2, most Plains Woodland occupations of the East Branch Walnut basin are thought to have occurred between 1000 and 1500 B.P., it follows that post-2000 B.P. aggradation ended and soil formation was initiated between ca. 1000 and 1500 B.P. Pedogenesis, continuous to the present, has resulted in the formation of a moderately developed soil (the Nuttall Soil) at 14BU4, 14BU9, 14BU25 and 14BU87. The assignment of the surface soil at 14BU25 to the Nuttall Soil is tentative, since the site did not yield independent evidence of time equivalence with the Nuttall Soil at the other sites.

Both the Nuttall and Snyder Soils formed in sediment deposited on the T-1 surface of the East Branch Walnut valley. As previously discussed in Chapter 2, the soils of this surface are mapped by Penner et al. (1975) as the Vanoss, Verdigris and Brewer series (see Chapter 2, Fig. 4). As demonstrated in Tables 6 and 7 (see below), the pedons of

the Nuttall and Snyder Soils described in this chapter exhibit soil properties similar to those of the Vanoss and Verdigris series, respectively. Before examining these similarities, the difference between a soil-stratigraphic unit and a soil series must be discussed.

For purposes of this study, the principal difference between a soil-stratigraphic unit (e.g., the Nuttall or Snyder Soil) and a soil series (e.g., Vanoss or Verdigris) is in the relative importance of stratigraphic relationships in the definition of each. A soil series is defined solely in terms of soil properties. Stratigraphic relationships are not considered in the definition of soil series (Morrison 1967:12; American Commission on Stratigraphic Nomenclature 1970:10; Birkeland 1974:248). By contrast, an essential element in the definition of a soil-stratigraphic unit is the specification of the unit's stratigraphic position relative to underlying and overlying soils and sediments. Simply stated, a soil-stratigraphic unit describes a soil formed in a given body (or bodies) of sediment, beneath a given surface of the landscape. A soil series describes a soil exhibiting a given set of soil properties, regardless of the stratigraphic or geomorphic context of the soil. "Soil-stratigraphic unit" and "soil series" are simply conceptual devices for describing and comparing soils. There is no a priori reason that a soil described as representing a given soil-stratigraphic unit cannot also be described as representing a given soil series. However, one must keep in mind that use of the former concept emphasizes the stratigraphic relationships of the soil, while use of the latter emphasizes only the properties of the soil.

The Nuttal Soil (a soil-stratigraphic unit) and the Vanoss series both possess minimally developed argillic (Bt) horizons. Table 6 compares characteristics of Nuttal Soil B horizons described in this chapter with the range of properties described for Vanoss series B horizons by Penner et al. (1975). The B horizon of the Nuttal Soil resembles the Vanoss series B in its blocky structure, color, thin, patchy clay skins and neutral to medium acid soil reaction (pH). The Nuttal Soil B horizons described in this chapter are less thick (range 20-35 cm) than is typical of the Vanoss series B (range 33-63 cm).

Pedons of the Snyder Soil described in this chapter resemble the Verdigris series (Table 7). In both buried and unburied pedons, the thickness, color, texture and reaction of Snyder Soil A horizons are largely within the ranges described for the Verdigris series (Appendix A). Exceptions include reaction at 14BU81 (neutral rather than slightly to medium acid) and A horizon thickness at 14BU9 (25 cm, as opposed to 50-88 cm). Also the Snyder Soil at 14BU87 lacks an A horizon and exhibits sedimentary structure not mentioned in the Verdigris series description.

The Nuttal and Snyder Soils sufficiently resemble the Vanoss and Verdigris series, respectively, that the mapped distributions of the two series may be used to describe the lateral extent of the two soil-stratigraphic units. Further evidence supporting this correlation is presented in Chapter 5, which discusses the different geomorphic contexts in which the Nuttal Soil/Vanoss series and the Snyder Soil/Verdigris series are found.

Table 6. Comparison of Bt horizon properties of the Nuttall Soil with those described for the Vanoss series by Penner et al. (1975).

Soil	Thickness (cm)	Color	Texture	Structure	Clay Skins	Reaction
VANOSS	33-63	10YR4/4 to 7.5YR5/4	silty clay loam	blocky	thin, patchy	neutral to medium acid
NUTTAL	20-35	10YR3/3 to 7.5YR4/4	silty clay to silty clay loam	subangular blocky to blocky	present	neutral to slightly acid
14BU4	35	7.5YR4/4	silty clay loam	subangular blocky	present	neutral
14BU9	25	10YR4/6	silty clay loam	blocky	present	neutral
14BU25	20	10YR3/3	silty clay	blocky	present	neutral
14BU87	32 ^a	10YR4/4	silty clay loam	blocky	present	slightly acid
^a eroded						

Table 7. Comparison of A horizon properties of the Snyder Soil with those described for the Verdigris series by Penner et al. (1975).

Soil	Thickness	Color	Texture	Reaction
VERDIGRIS	50-88 cm	10YR4/1 to 10YR3/2	silt loam or silty clay loam	slightly acid to medium acid
SNYDER	25-84 cm	10YR3/2 to 10YR3/3	silt loam or silty clay loam	neutral to medium acid
14BU4 (buried)	50	10YR3/2	silt loam	neutral
(unburied)	84	10YR3/2 to 10YR3/3	silt loam to silty clay loam	neutral
14BU9 (buried)	25	10YR3/2	silt loam to silty clay loam	neutral
14BU25 (buried)	55	10YR3/2	silty clay loam	neutral
14BU81 (unburied)	55	10YR3/2	silty clay loam	slightly acid to medium acid
14BU87 (buried) ^a	-	-	-	-
^a Snyder Soil lacks A horizon.				

CHAPTER 4

SOIL PROPERTIES AND PEDOGENIC PROCESSES

This chapter discusses the morphological, physical and chemical properties of the Nuttal and Snyder Soils. Following the methodological approach outlined in Chapter 1, soil properties are viewed as the manifestation of pedogenic processes, which in turn are responsive to environmental factors. The objective of this chapter is to examine the properties of each soil and infer from them the dominant trends in pedogenesis during each soil-forming interval.

The discussion is presented in three sections. The first describes the analytical techniques employed to measure soil properties. The second section discusses pedogenesis of the Nuttal Soil. The third section is primarily concerned with pedogenesis of the Snyder Soil during the interval 4000-2000 B.P., but also considers processes which operated after 2000 B.P., modifying the primary properties of the Snyder soil.

ANALYTICAL METHODS

Laboratory analysis was conducted on 84 samples collected from the pedons which were described in detail in Chapter 3. All samples were taken from the walls of archaeological excavations in the course of archaeological field work. At all sites but 14BU4, samples were collected at 10 cm intervals. At 14BU4, in XU102, one sample each was taken from the A, Bt and Ab horizons, at depths of 15-25 cm, 45-55 cm and 75-85 cm, respectively. Samples were taken at 10 cm intervals in the cultural stratum between 130 and 190 cm. Below 190 cm, samples were taken every

20 cm to a depth of 230 cm. At 14BU4, in Trench C, four samples were taken from within soil horizons identified in the field, at depths of 24-40, 45-55, 70-80 and 100-110 cm. In hindsight, continuous-interval sampling would have been preferred, since the interpretations offered in this chapter rely on vertical patterns in soil properties. Despite this sampling irregularity, the data from 14BU4 are believed to adequately reflect vertical trends in soil properties at the site.

Samples were air-dried following collection, then stored. Samples from 14BU81 and 14BU87 were stored in sealed plastic bags for approximately 15 months before analyses were undertaken. Samples from 14BU4, 14BU9 and 14BU25 were stored in sealed paper bags for approximately 3 months prior to analysis.

All analyses except loss-on-ignition were carried out by personnel of the Kansas State Soil Testing Laboratory, Kansas State University, Manhattan, Kansas. Analyses were directed by Dr. David Whitney, Extension Specialist in Soil Testing, who kindly provided information and bibliographic citations concerning analytical procedures used in the state laboratory. Loss-on-ignition was determined by the author in the Soils Laboratory, Department of Geography/Meteorology, University of Kansas. Prior to all analyses, soil subsamples were crushed to destroy structural aggregates, then passed through a 2 mm (U.S. Standard 10-mesh) sieve (Eik et al. 1980).

Soil texture, reported as the percentage by weight of sand (0.05-2.00 mm), silt (0.05-0.002 mm) and clay (less than 0.002 mm), was determined by hydrometer analysis (Day 1965:562-564). Fifty grams

of oven-dry soil were dispersed in sodium hexametaphosphate solution, mechanically mixed to break up soil aggregates, then transferred to a 1000 ml settling cylinder and thoroughly shaken to obtain a uniform suspension of soil in liquid. A Bouyoucos hydrometer was inserted and read after 40 seconds had elapsed. A second reading was taken two hours after insertion. These intervals represent the time required for sand- and silt-sized particles, respectively, to settle from suspension, according to Stoke's law relating particle size to settling velocity (Day 1965:549-550). Hydrometer readings, representing the weight in grams of sediment remaining in suspension after each interval, were corrected for temperature effects and converted to percentages by weight.

Soil pH was determined electropotentially by inserting the electrodes of a pH meter into a 1:1 suspension of 5 g soil and 5 ml distilled water (McLean 1980). In the Kansas State Soil Testing Lab, the 1:1 suspension is allowed to equilibrate for 20 minutes prior to reading pH (Whitney, personal communication).

Available phosphorus was determined by the Bray-1 procedure (Bray and Kurtz 1945) as recommended by Knudsen (1980). Phosphorus was extracted from 1 g of soil with a solution of ammonium fluoride and hydrochloric acid. Using the Fiske-Subbarow method (Knudsen 1980:15), the extract was treated with hydrochloric acid and ammonium molybdate in the presence of an aminonaphthol-sulfonic acid reducing agent to produce a color reaction. Concentration of phosphorus in parts per million was determined colorimetrically by comparison with a series of standard phosphorus solutions of known concentration, using a spectrophotometer (Knudsen 1980:15).

Easily oxidizable organic matter was measured by the modified Walkley-Black method of Graham (1948; see also Schulte 1980). One g samples of soil were treated with an excess of potassium dichromate to oxidize the organic matter. Oxidation was facilitated by adding sulfuric acid to generate heat. The mixture was allowed to stand for an hour, then diluted with water and filtered. The filtrate was read colorimetrically with a spectrophotometer to determine the amount of potassium dichromate remaining after oxidation. By assuming that the potassium dichromate treatment oxidizes 77% of the soil's organic carbon, and that soil organic matter is approximately 58% carbon (Schulte 1980: 28-29), the percentage by weight of soil organic matter was calculated from the amount of potassium dichromate remaining after oxidation.

To determine loss on ignition (LOI), 4.0-4.5 g of soil were oven-dried at 105° C for at least 6 hours to remove hygroscopic moisture, then heated in a muffle furnace (Wilde et al 1964:19-20). Weights of the sample before and after ignition were measured to 0.0001 g on an analytical balance. The loss of weight due to ignition was expressed as a percentage of the pre-ignition, oven-dry weight. Following the recommendation of Smith and Atkinson (1975:174) that LOI be determined at several temperatures to check for variations in weight loss due to losses from inorganic soil constituents, LOI was determined after 3 hours at 600° C (Wilde et al. 1964:19-20) and after 8 hours at 400° C (Jackson 1958; Smith and Atkinson 1975).

PEDOGENESIS OF THE NUTTAL SOIL: 1500-1000 B.P. to Present

The Nuttal Soil formed in sediment deposited after 2000 B.P. The presence of Plains Woodland occupations at the soil's present day surface suggests that deposition ceased and pedogenesis began some time during the interval 1500-1000 B.P. Data on the physical and chemical properties of the Nuttal Soil are available from pedons at 14BU4, 14BU9, 14BU25 and 14BU87. The Nuttal Soil is not present at 14BU81. Where identified, the Nuttal Soil consists of a dark, organic rich A horizon overlying a clayey Bt horizon.

Clay Translocation

The Nuttal Soil can be considered to have a minimally developed argillic B horizon. Argillic horizons are characterized by the accumulation of illuvial clay, or clay which has been translocated downward from overlying horizons. The presence of translocated clay is symbolized by a subscript "t" following the B horizon designation.

The presence of illuvial clay in the Nuttal Soil is suggested by higher clay contents in the B horizon, relative to overlying horizons. As shown in the textural distributions of Figs. 18-24, the four Nuttal Soil pedons exhibit slight "bulges" of clay in the B horizon.

According to Soil Survey Staff (1975:27) criteria, an argillic horizon must have at least 1.2 times as much clay as an overlying horizon, when the overlying horizon has 15% to 40% clay. As shown in Table 8, which presents ratios of B to A horizon clay contents for the four pedons, the critical value of 1.2 is met for all pedons by comparing maximum B to minimum A clay content. Even by maximizing the ratio in this way,

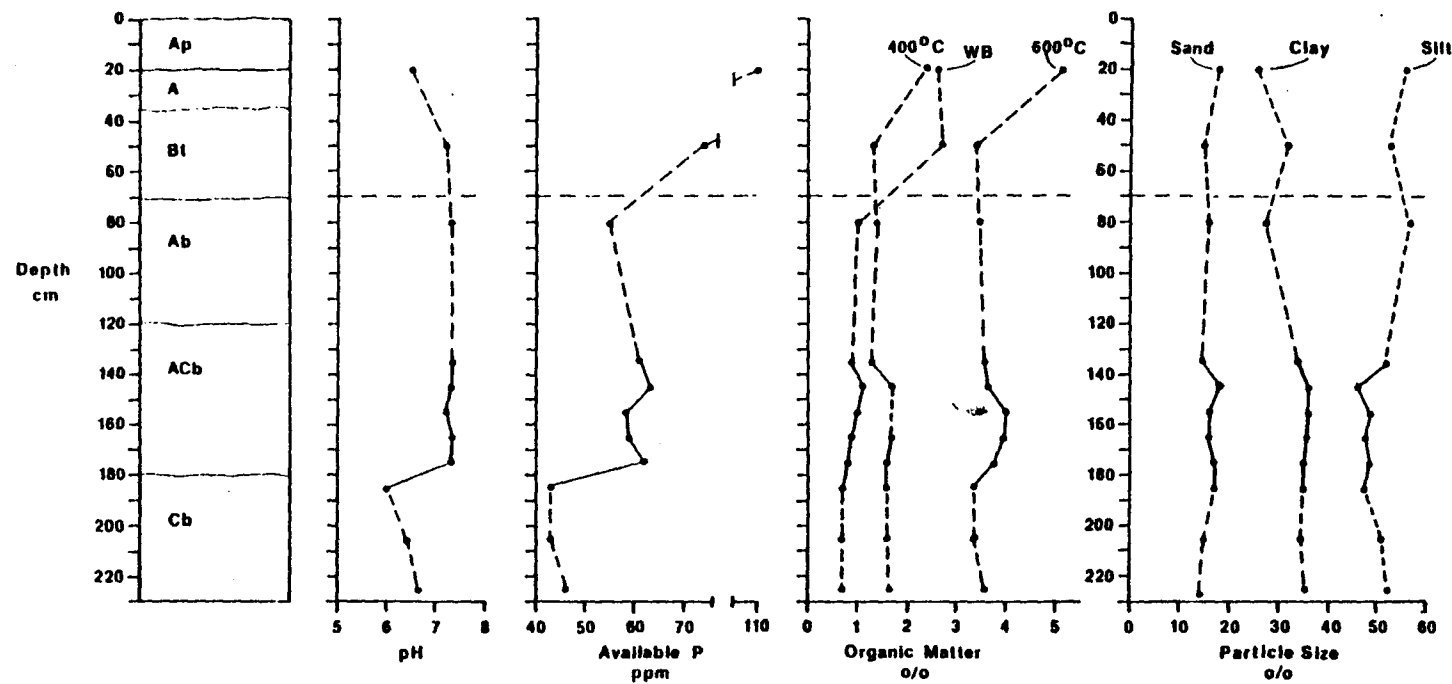


Fig. 18. Soil properties at 14BU4 (XU102).

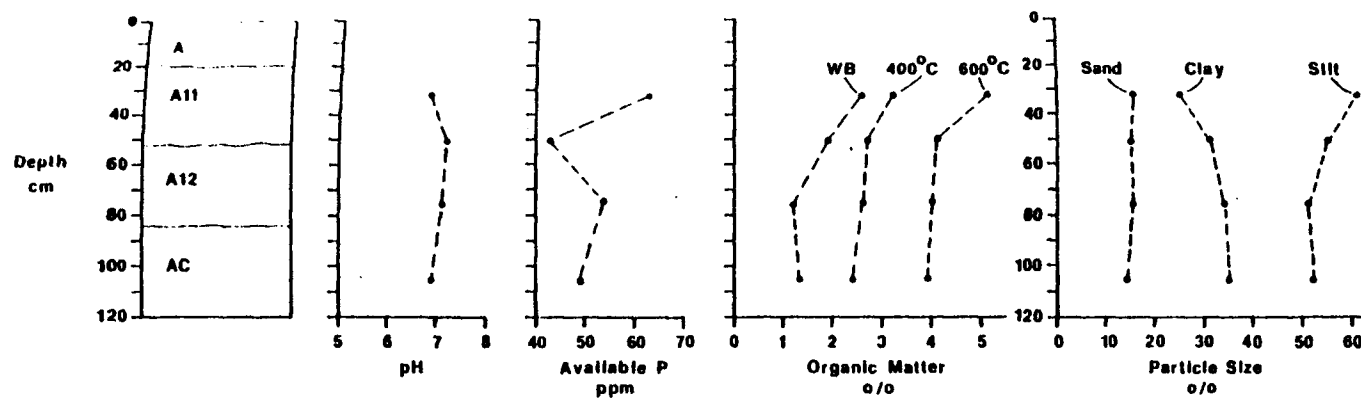


Fig. 19. Soil properties at 14BU4 (Trench D).

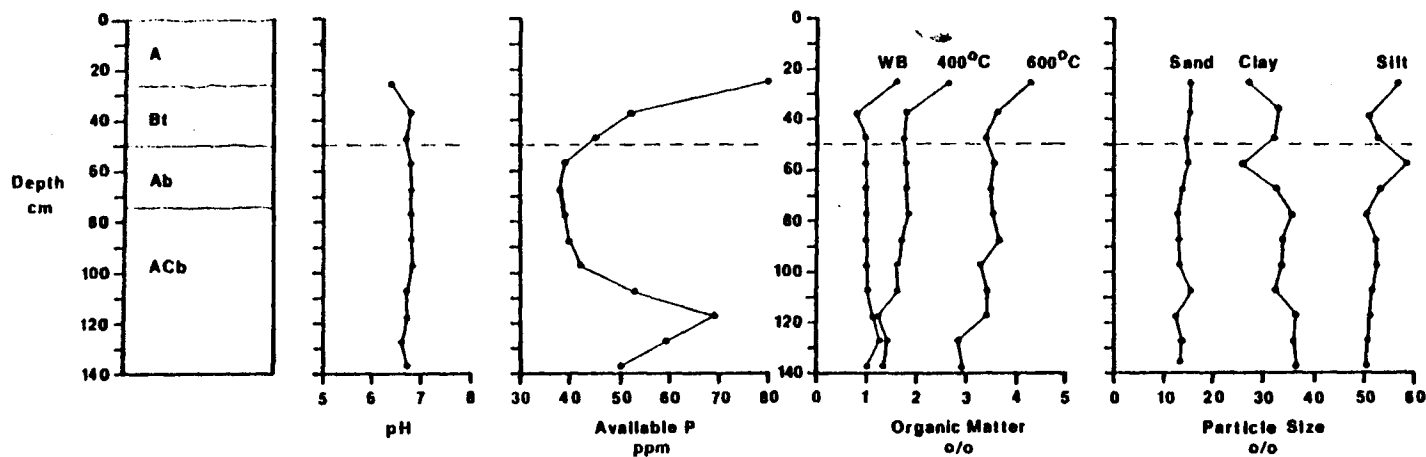


Fig. 20. Soil properties at 14BU9 (1978-1979 block).

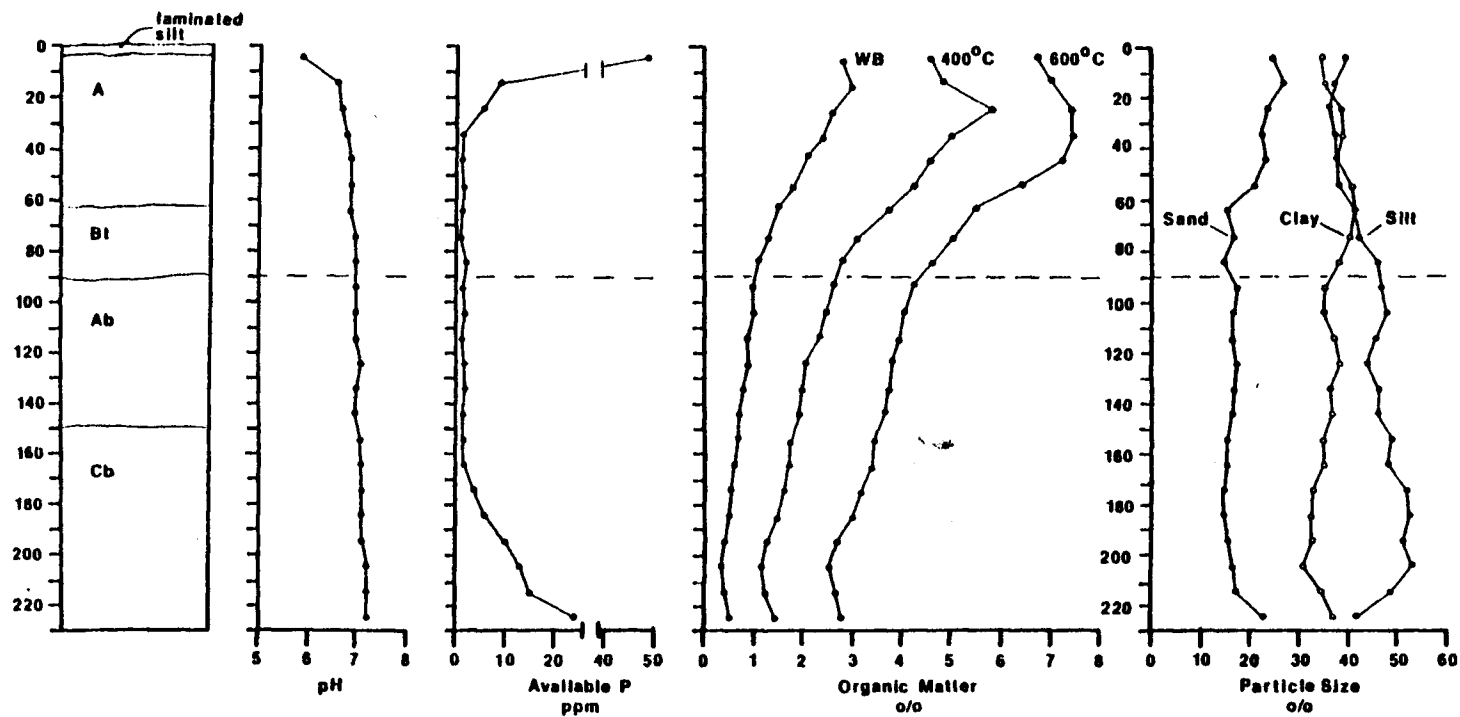


Fig. 21. Soil properties at 14BU25 (XU101).

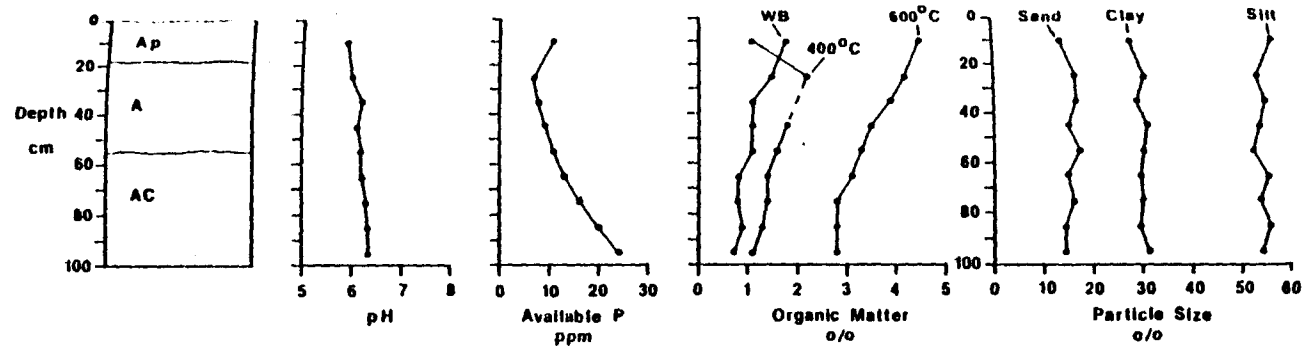


Fig. 22. Soil properties at 14BU81 (TP1).

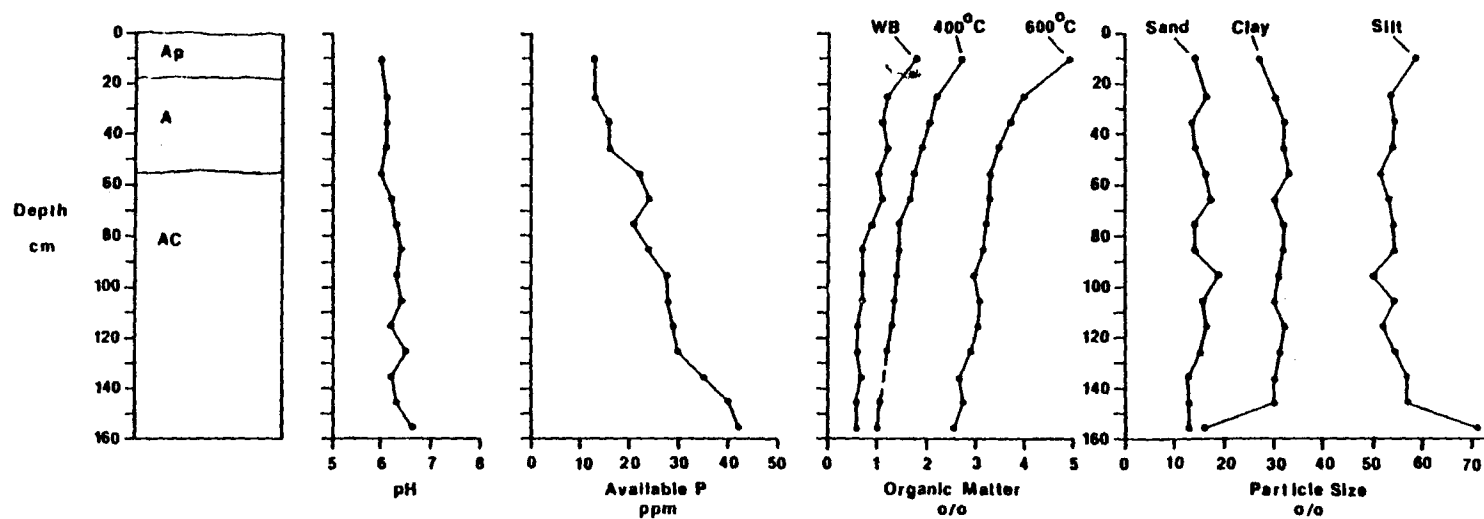


Fig. 23. Soil properties at 14BU81 (TP2).

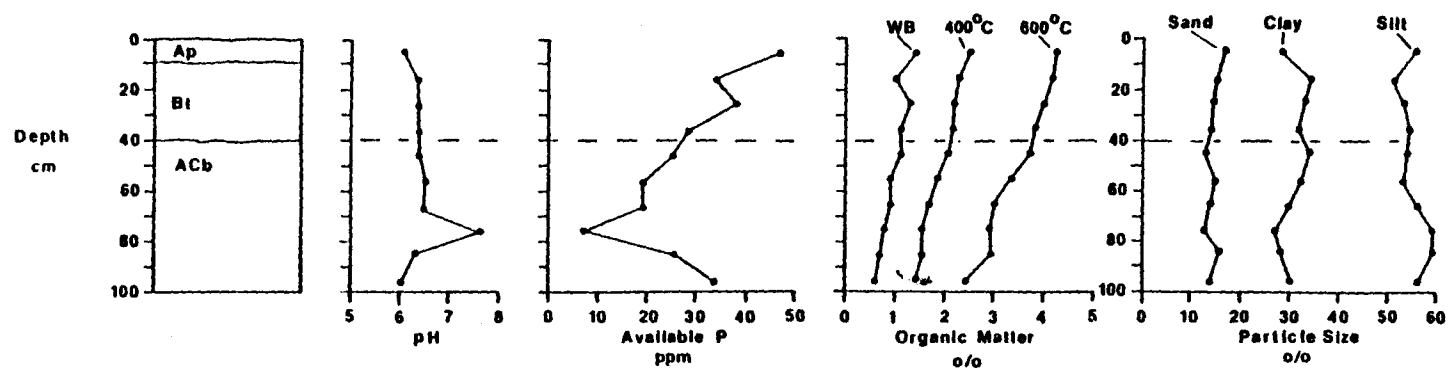


Fig. 24. Soil properties at 14BU87 (TP1).

Table 8. Ratios of B to A horizon clay contents in the Nuttal Soil.

Site	A horizon		B horizon		$\frac{B_{max}}{A_{min}}$	$\frac{B_{avg}}{A_{avg}}$
	Minimum	Average ^a	Maximum	Average ^a		
14BU4	25.6	-	32.1	-	1.25	-
14BU9	27.6	-	33.6	33.1	1.22	1.20
14BU25	35.2	38.0	42.6	40.6	1.21	1.07
14BU87	28.0	-	34.0	32.9	1.21	-

^a Average computed only for those pedons having more than one sample per horizon.

the critical value is exceeded by only a few hundredths of a percent. When average clay contents are compared, 14BU25 does not meet the critical figure of 1.2. Thus, in terms of clay content, the Nuttal Soil pedons exhibit only minimally developed argillic horizons.

Clay content alone is insufficient to identify an argillic horizon, since processes other than illuviation can account for this property. Chittleborough and Oades (1979) list 3 processes which may contribute to clay accumulation in B horizons, as follows:

"1. Sedimentary layering. The differences in particle size distributions are the result of initially non-uniform parent material.

"2. Formation of clay in position by differential weathering. Under this mechanism, there has been no clay movement.

"3. Downward translocation of clay within an initially homogenous parent material."

These alternatives can be evaluated using various lines of evidence. Parsons and Balster (1967) and Parsons et al. (1968) use geomorphic

evidence to demonstrate that the thick, heavy clay B horizons of planosols in the Willamette Valley, Oregon, represent ancient lacustrine sediments, thus representing alternative 1, above. Alternative 2, in situ formation of B horizon clay, is detectable by mineralogical analysis, comparing ratios of minerals which are highly weatherable versus those which are resistant to weathering (Chittleborough and Oades 1979). Translocated clay can also be distinguished from in-place weathering products by comparing A and B horizons in terms of fine clay/total clay ratios (Walker and Hutka 1979), and in terms of relative proportions of fine clay minerals such as montmorillonite (McKeague and St. Arnaud 1969). High proportions of fine clays in B horizons indicates translocation, since finer-sized particles will translocate more readily through the small pores of the soil.

The best evidence for clay translocation is the presence of clay skins, or films of clay formed on ped surfaces and in open pores as a result of the downward movement and precipitation of clay (McKeague and St. Arnaud 1969:429). Clay skins, if correctly identified (McKeague and St. Arnaud 1969:430), provide conclusive evidence of illuviation (St. Arnaud and Whiteside 1964:97). However, the absence or weak expression of clay skins is not sufficient to rule out illuviation. Clay skins either do not form or do not persist in soils with clay content of 40% or more, since such soils are susceptible to shrinking and swelling (Nettleton et al. 1969). Birkeland (1974:112) cites studies which suggest that in soils with greater than 20-40% clay, pore spaces may be too small to permit clay movement, except along structural discontinuities (e.g., gaps formed between peds by dessication). Since soil

structural aggregates become increasingly less stable with increasingly finer textures (White 1963), clay skins in clayey soils will tend to be poorly developed or impersistent.

An examination of peds from 14BU4 and 14BU9 using a binocular microscope reveals thin, discontinuous clay skins in the B horizons of the Nuttal Soil. At 14BU9, clay skins occur on the walls of open pores and root channels (see pedon description, Chapter 3). The clay content of the B horizon (30-40%) is sufficiently high to suggest that the shrinking and swelling of clays is a factor in the weak expression of clay skins. The presence of clay skins, however, is sufficient evidence to suggest that some illuviation of clay has occurred in the Nuttal Soil.

The process of clay translocation in soil requires the wetting of the upper horizon, thereby suspending clay particles, and the transport of these particles via percolating water into lower horizons (Sehgal et al. 1976; Birkeland 1974:106-113; McKeague and St. Arnaud 1969:431). Experimental studies suggest that clay movement stops where the downward movement of soil water is arrested (McKeague and St. Arnaud 1969:430). Clay is then "filtered out" onto ped surfaces as water soaks into unsaturated peds, or is precipitated on ped surfaces as the water evaporates or is transpired by organisms (McKeague and St. Arnaud 1969:431; Buol and Hole 1961:379).

Shallowly-buried paleosols may promote the formation of illuvial horizons in the soils above them by creating a barrier to percolating water. For example, soils developed on prehistoric effigy mounds in

Iowa exhibit clay skins, indicative of illuviation at the contact of the mound fill and an underlying, buried B horizon (Parsons et al, 1962: 493). The B horizon represented the original soil, which was truncated during mound construction. Parsons et al. (1962:493) suggest that downward water movement is impeded at the contact, resulting in the precipitation of suspended clay as the matrix dries. Although the authors do not specifically draw the conclusion, the buried B horizon's higher clay content and greater bulk density, relative to the overlying fill, probably rendered the buried B horizon comparatively impermeable, thus accounting for impeded drainage at its upper boundary.

A similar process of B horizon initiation was observed by Hallberg et al. (1978) in soils developed on 100-year-old spoils derived from the excavation of railroad cuts in western Iowa. They suggested that the translocation and accumulation of clay within this short period of time was

"...enhanced by low bulk density of the spoils, allowing relatively rapid percolation of soil water. Clay accumulation was further enhanced by impeded internal drainage caused by the buried soil and surface at a shallow depth (64-66 cm)" (Hallberg et al. 1978:343).

In addition, the Ab horizon of the underlying Tama series soil had greater bulk density than a nearby, unburied pedon of the same series. This increase was partly attributed to compaction of the soil at the time of railroad construction, but primarily to post-burial loss of fibrous organic matter from the profile. Hallberg et al. (1978:343) postulate that the increased bulk density and decreased permeability of the buried soil contributed to impeded drainage and clay accumulation in the overlying horizon.

Although bulk density was not determined for the Nuttall Soil, evidence is presented later in this chapter for the post-burial deterioration of organic matter in the Snyder Soil A horizons. This deterioration may have increased the bulk density of the buried soil sufficiently to arrest downward movement and initiate clay accumulation. The similarities between the Nuttall Soil B horizon and those described by Hallberg et al. (1978) and Parsons et al. (1962) suggest that similar pedogenic processes are involved, and that clay accumulation in the Nuttall Soil is reasonably attributed to translocation.

Clay translocation is necessarily a lengthy process. Studies of the time required for Bt horizon genesis suggest that 1000 B.P., the minimum date for stabilization of the Nuttall Soil surface, is a minimum age for the formation of incipient argillic horizons. In the Willamette Valley, Oregon, cambic B horizons (i.e., "color" B horizons, differentiated from overlying and underlying horizons by stronger chromas and redder hues, but lacking evidence of clay illuviation) form within 555 years, while well developed argillic horizons form within 5250 years (Parsons et al. 1970). In soils developed on prehistoric mounds in Iowa (Parsons et al. 1962; McComb and Loomis 1944), clay skins formed within 1000 to 2500 years. Within 2000 to 2500 years, soils on Middle Woodland mounds developed blocky ped structure comparable to that of nearby, undisturbed soils, although the latter soils exhibit larger peds and more pronounced clay skins (Parsons et al. 1962:496). Alluvial soils in Pennsylvania show evidence of clay translocation within 470 years, although a B horizon with a charcoal date of 1955 B.P. was considered to be intermediate between cambic and argillic stages, based

on fine clay/total clay ratios (Bilzi and Coilkosz 1977).

The above studies, representing a variety of geomorphic and pedologic situations from humid and sub-humid North America indicate that at least 1000-2000 years are required for the formation of minimally developed argillic horizons like those present in the Nuttal Soil. The estimate of 1000-1500 B.P. for the initiation of Nuttal Soil pedogenesis is within this range.

Color Development

Soil colors of the Nuttal Soil B horizons (Table 9) have higher chromas (the final digit in the Munsell notation) and, at 14BU4, a higher hue (the first expression in the notation), than the overlying horizons. The development of reddish hues and bright chromas, leading to red and yellow-brown soil color, results from the chemical weathering of minerals, primarily by oxidation and hydrolysis (Birkeland 1974:12; Buol et al. 1980:22,80-83). According to Buntley and Westin (1965:579), variation in hue and chroma reflects "alterations in the parent material associated with the intensity of weathering, stage of soil development and drainage." These authors found that the product of hue and chroma provides a useful index of relative profile development, which they termed "color developmental equivalent" or "CDE". CDE values for Nuttal Soil pedons are shown in Table 9. As suggested by Buntley and Westin (1965), the two hues (10YR and 7.5YR) are assigned values of 1 and 2, respectively.

CDE's of Nuttal Soil Bt horizons range from 3 at 14BU25 to 8 at 14BU4. The sites may be ranked in terms of increasing Bt horizon CDE

Table 9. Color developmental equivalents (Buntley and Westin 1965) of the Nuttall Soil.

Site	Soil Horizon	Munsell Color	Developmental Color ^a
14BU4	A	10YR3/2	1 x 2 = 2
	B2t	7.5YR4/4	2 x 4 = 8
14BU9	A	10YR4/2	1 x 2 = 2
	B2t	10YR4/6	1 x 6 = 6
14BU25	A	10YR2.5/1	1 x 1 = 1
	B1	10YR2.5/1	1 x 1 = 1
	B2t	10YR3/3	1 x 3 = 3
14BU87	A	10YR3/1	1 x 1 = 1
	B2t	10YR4/4	1 x 4 = 4

^a Hue times chroma, where 10YR = 1 and 7.5YR = 2.

as follows: 14BU25, 14BU87, 14BU9 and 14BU4. This ranking is similar to that determined by Bt/A horizon clay ratios (Table 8) and suggests that the contrast between soil horizons, reflecting the relative degree of profile development, is weakest at 14BU25 and 14BU87, and strongest at 14BU4 and 14BU9. At 14BU25, a color contrast with the surface horizon is identified only in the B2t; the B1, although distinguished from the A horizon by structural development, is identical to the A in soil color.

Organic Matter Accumulation

The surface horizons of the Nuttal Soil at 14BU4, 14BU9 and 14BU25 meet Soil Survey Staff (1975:14-16) requirements for classification as mollic epipedons. The most important of these requirements are that the epipedon

"...contains 1% or more organic matter, with color values darker than 5.5 dry and 3.5 moist. Structure cannot be massive and hard. Base saturation is over 50%." (Buol et al. 1980:386)

The Nuttal Soil surface horizon is within the established criteria (see Appendix B and Chapter 3, pedon descriptions), although base saturation cannot be evaluated for lack of data. The severely eroded Nuttal Soil epipedon at 14BU87 is excluded from the following discussion.

The dominant process in the genesis of mollic epipedons is melanization, "the process of darkening of the soil by addition of organic matter" (Buol et al. 1980:259). The process occurs fairly rapidly; A horizons developed on 100 year old spoils in Iowa met most criteria for classification as a mollic epipedon except for slightly lighter

color (Hallberg et al. 1978:34).

Melanization, reflected by high organic matter content and dark soil color, is recognized in the Nuttal Soil A horizon at 14BU4 and 14BU9 and in the Nuttal Soil A and B1 horizons at 14BU25. The high value of Walkley-Black organic matter obtained from the Bt horizon at 14BU4 is anomalous, since the strong color development of the horizon indicates that organic matter originally present has been oxidized. Loss on ignition, a second technique used to determine organic matter, failed to detect high values in the B horizon. Thus, the high Walkley-Black value is probably attributed to an error in sampling or analysis.

At 14BU25 (Fig. 21), both Walkley-Black and LOI data show increases in organic matter at depths of 20 to 40 cm in the Nuttal Soil (Appendix B, Table B.4). The low organic matter content of the surface samples probably reflects the fact that the surface of the sampled pedon at 14BU25 was underlain by 5 cm of recently deposited laminated silt (see Chapter 3, Fig. 11). Plowing may have mixed organic-poor sediment with the underlying A horizon and lowered organic matter contents throughout the 20 cm plow layer.

Anthropogenic Enrichment of Phosphorus

The high values of available phosphorus in the surface samples of all four pedons reflect accumulation of phosphorus-rich organic residues. Enhancement by agricultural fertilizers is probably also reflected, since the sites have been intensively cultivated for many years. Phosphorus concentrations are highest at 14BU4 and 14BU9, in both the A and Bt horizons of the Nuttal Soil. At 14BU25, the surface sample

yielded 49 parts per million, compared to 110 ppm and 80 ppm at 14BU4 and 14BU9, respectively (Appendix B). In addition, at 14BU25, phosphorus declined abruptly below 10 cm in depth, reaching concentrations of only 1-2 ppm in the Bt horizon. By contrast, B horizons at 14BU4 and 14BU9 yielded 74 and 50 ppm available phosphorus, respectively.

The high concentration of available phosphorus at 14BU4 and 14BU9, relative to 14BU25, may reflect the presence at the former two sites of surficial Woodland components. Prehistoric human occupation contributes phosphorus to the soil in the form of "...domestic refuse, food wastes, plant and animal remains, excreta (and) bodies...." (Proudfoot 1976:93). At 14BU25, the absence of a surface archaeological component (Root 1981a:26) may account for the lower concentration of available phosphorus in the Nuttall Soil. Alternatively it should be noted that the pH of the uppermost sample at 14BU25 is 5.9, lower than at either 14BU4 (pH = 6.5) and 14BU9 (pH = 6.4). At pH's lower than 6.0 and higher than 7.0, increasingly greater proportions of soil phosphorus are fixed in insoluble forms, and thus are not "available" for use by plants (Brady 1974:467). The low phosphorus values at 14BU25 thus may reflect the form of phosphorus present rather than, or as well as, the amounts present. Phosphorus distribution in the East Branch Walnut pedons is discussed extensively with reference to the Snyder Soil, below.

Pedogenic Processes and Trends

To summarize, the major pedogenic processes evidenced by properties of the Nuttall Soil are melanization and clay translocation. The dominant

trend of both processes is toward increased horizonation, manifested in the Nuttall Soil by the formation of minimal argillic B horizons which are distinct in color and texture from the overlying, organic rich mollic epipedons. The extent of horizonation varies slightly among the pedons. Textural and color-developmental contrasts are weakest at 14BU25 and strongest at 14BU4 and 14BU9. However, organic matter and phosphorus characteristics vary less between horizons at 14BU4 and 14BU9, than at 14BU25. This characteristic may reflect anthropogenic enhancement of organic matter and phosphorus in the 14BU4 and 14BU9 surface horizons as a result of Woodland occupations.

PEDOGENESIS OF THE SNYDER SOIL: 4000-2000 B.P.

The Snyder Soil formed in sediment deposited between 4000 and 2000 B.P. At sites 14BU4, 14BU9, 14BU25 and (tentatively) 14BU87, the Snyder Soil A horizon is buried beneath the Nuttall Soil. Archaeological evidence presented in Chapter 3 indicates that the Snyder Soil was buried soon after 2000 B.P., since the formation of the Nuttall Soil was in progress by 1000 to 1500 B.P. The Snyder Soil must therefore have formed prior to 2000 B.P.

Pedogenic processes operative during the 4000-2000 B.P. soil-forming interval are difficult to infer from soil properties alone, since these properties were undoubtedly altered following the burial of the Snyder Soil after 2000 B.P. Burial imposed significantly different pedogenic conditions on the Snyder Soil. Simply stated, the surface horizon of the Snyder Soil became a subsoil horizon of the overlying Nuttall Soil. As a consequence, the primary properties of

the soil, developed prior to 2000 B.P., were modified as the soil acquired secondary characteristics as a result of burial. Unburied Snyder Soil pedons may also have experienced post-burial alteration, if pedogenesis before and after 2000 B.P. proceeded under fundamentally different environmental conditions. In addition, the presence of archaeological components in the Snyder Soil at all five sites may also influence soil properties, such as organic matter and phosphorus.

Properties of the Snyder Soil thus reflect three sets of pedogenic processes:

1. Primary processes, by which the Snyder Soil formed during the interval 4000-2000 B.P.;
2. Secondary processes, operative after 2000 B.P., which modified primary properties of the Snyder Soil;
3. Processes of anthropogenic enhancement, influencing both primary and secondary processes through the accumulation and decomposition of organic refuse deposited by prehistoric occupants of Snyder Soil surfaces.

In that processes of organic decomposition begun during the Snyder Soil Forming Interval have undoubtedly continued to the present, the third set of processes constitutes an additional source of post-burial alteration.

Although paleopedological studies have successfully discriminated primary and secondary characteristics of polygenetic soils using pedological evidence alone (e.g., Bryan and Albritton 1943; Ruhe 1956; Bos and Sevink 1975; Valentine et al. 1980), the present study uses ancillary radiocarbon and archaeological evidence to suggest the dominant pedogenic processes operative prior to 2000 B.P. As the following paragraphs demonstrate, these processes were 1) cumulation, or the continual addition of sediment to the Snyder Soil surface; and 2)

homogenization of the Snyder Soil by pedoturbation. Once these dominant processes are identified, properties of the Snyder Soil are examined to discern further details of pre-2000 B.P. pedogenesis and to ascertain the nature and extent of post-burial alteration and anthropogenic modification of soil properties.

Cumulization

Buol et al. (1980:97) define cumulization as "the accumulation of mineral material onto the surface of the soil by either air or water." The extent to which this accumulation inhibits soil horizonation is a function of the rate of sediment accumulation. If sediment aggrades rapidly, pedogenesis may not occur at all. However, if the rate of deposition is sufficiently slow, new sediment is incorporated into the surface horizon as it is deposited (Nikiforoff 1949:228). A soil formed by the incorporation of steady increments of fresh sediment is termed a "cumulative soil" (Nikiforoff 1949:228; Riecken and Poetsch 1960:275).

The cumulative nature of the Snyder Soil is indicated by radiocarbon dates from 14BU9, which suggest that sedimentation during the 4000-2000 B.P. interval was sufficiently slow to allow pedogenesis to occur. Estimated rates of aggradation for the upper one meter of the Snyder Soil at 14BU9 (Table 10) are calculated from paired groupings of three dates (2060 ± 80 , 3240 ± 85 and 3980 ± 100 B.P.). Three dates (1970 ± 110 , 3650 ± 140 and 3910 ± 160) were not used in the calculations. The only recorded vertical provenience of these samples are intervals of depths (Chapter 2, Table 3). These dates probably represent composite samples of small pieces of charcoal collected from

Table 10. Estimated rates of aggradation for the upper 1 m of the Snyder Soil, 14BU9.

Radiocarbon Age (yrs B.P.)	Sample Depth (cm)	Depth Difference (cm)	Minimum Age Difference (yrs)	Maximum Age Difference (yrs)	Estimated Aggradation Rate (cm / yr)
Upper 1 m of Snyder Soil					
2060+80	45	83	1740	2100	.040-.048
3980+100	128				
Upper half of interval					
2060+80	45	39	1015	1345	.029-.038
3240+85	84				
Lower half of interval					
3240+85	84	44	555	925	.048-.079
3980+100	128				

throughout a vertical interval. This probably accounts for their relatively large sigmas, and limits their utility in estimating aggradation rates. However, it should be noted that the disregarded dates do not contradict the linear trend of the three discrete-depth samples, as shown in Fig. 25.

The average aggradation rate for the upper one meter of the Snyder Soil, estimated from dates of 2060 $\pm$ 80 and 3980 $\pm$ 100 B.P., is .05 cm/year (Table 10). The lower half of the 83 cm interval may have accumulated more rapidly than the upper half. The lower, 44 cm thick increment, separating dates of 3240 $\pm$ 85 and 3980 $\pm$ 100 B.P., accumulated at an estimated average rate of 0.05 to 0.08 cm/year. The upper, 39 cm thick interval, separating dates of 2060 $\pm$ 80 and 3240 $\pm$ 85 B.P.,

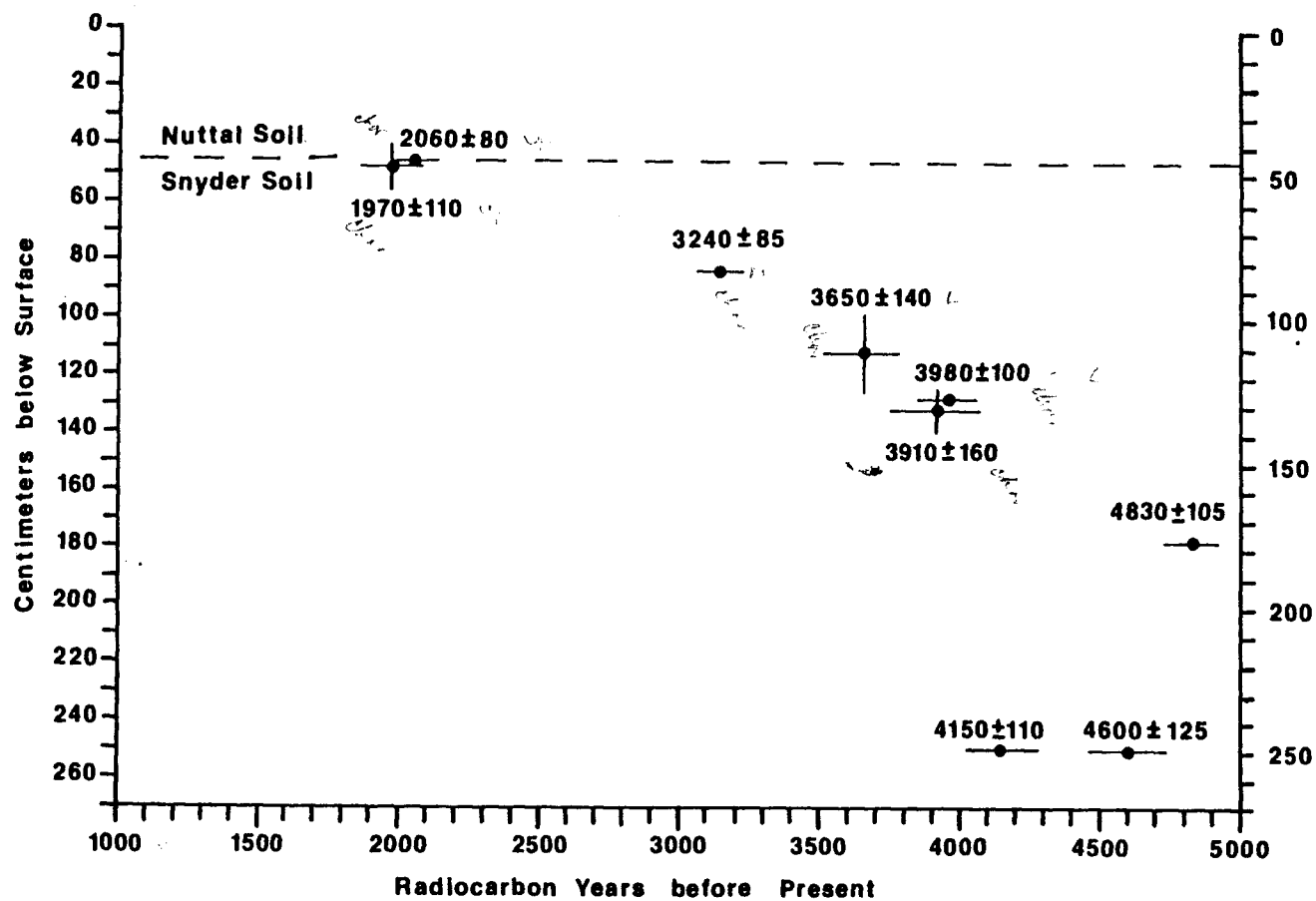


Fig. 25. Age-depth relationships of radiocarbon dates from 14BU9.

accumulated more slowly, at 0.030.04 cm/year (Table 10).

The rates shown in Table 10 never exceed 1 mm of sediment per year. Undoubtedly, there was variation in the actual amount of sediment deposited each year on the Snyder Soil surface. For instance, there may have been years in which no accumulation occurred and others in which several centimeters were deposited. Table 10 does not indicate this sort of variation. Nevertheless, the average rate of aggradation is sufficiently slow to allow the conclusion that pedogenesis probably accompanied deposition, and that the Snyder Soil is a cumulative soil, formed between 4000 and 2000 B.P.

Homogenization

Hoeksama (1953:30) applied the term "homogenization" to the processes by which the sedimentary structure of alluvial deposits is destroyed, creating a homogeneous soil material. Pedogenetic mixing of soil, referred to as pedoturbation (Hole 1961), is accomplished by a variety of agents and processes including freeze-thaw, shrink/swell of clays, the growth and decay of plant roots, and the activities of burrowing animals. Hole (1961) and Wood and Johnson (1977) provide comprehensive discussions of pedoturbational vectors. Hoeksama (1953:30) attributes the homogenization of alluvial soils largely to the actions of soil flora and fauna.

Archaeological evidence from the Snyder Soil at 14BU9 suggests that homogenization accompanied cumulation. Vertical distributions of artifacts recovered from excavations at 14BU9, presented in Figs. 26-27 and Table 11, show that although Late Archaic artifacts are

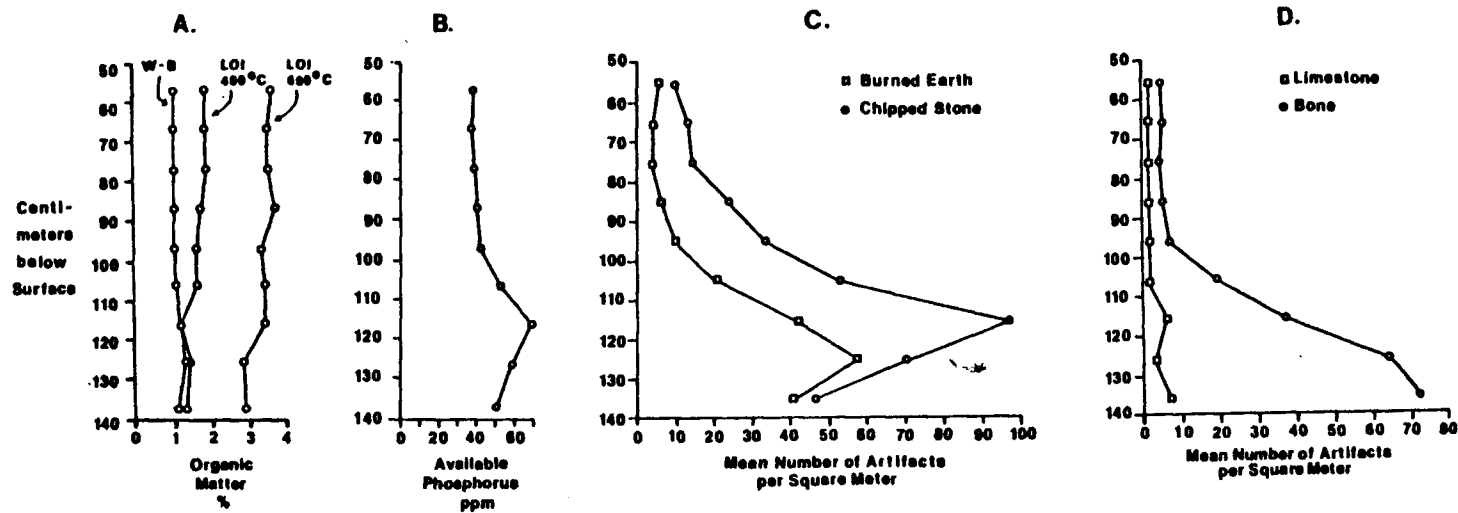


Fig. 26. Soil properties and artifact distributions from 14BU9. Artifact distributions after Haury and Leaf 1983: Figs. 4.6, 4.7, 4.9, 4.10.

distributed throughout the Snyder Soil, the distribution is not uniform. Rather, the distributions show readily apparent modes, defined by high artifact frequencies, above and below which frequencies decline.

In the 1979 block excavation (Haury and Leaf 1983), maximum frequencies of chipped stone, limestone and burned earth were encountered at depths of 100-120 cm in the Snyder Soil (Fig. 26c-d), and are paralleled by increases in organic matter and available phosphorus, presumed to represent anthropogenic enhancement of soil properties (Fig. 26a-b). Unlike other artifact classes, animal bone continues to increase in frequency below 120 cm (Fig. 26d). This distributional pattern, attributed to differential preservation, is explained under "Soil Phosphorus", below.

Data from the 1968 excavations also reveal a bimodal vertical distribution of artifacts. Grosser (1970) presents data on the absolute and relative frequencies of thermally altered chert flakes per 25 cm level in the 1968 excavation. Although he does not present the total number of flakes, these can be calculated from his data and are presented here in Table 11. The resultant bimodal distribution has peak frequencies at 50-75 cm and 100-125 cm.

Fig. 27 shows the vertical distribution of chipped stone artifacts in the eight, 2 x 2 m squares of the 1969 block excavation. Because the thickness of vertical provenience units in the 1969 excavation varied markedly both within and among squares (Grosser 1970:34), the vertical distribution of artifacts for the excavation as a whole cannot be determined from Grosser's data. However, given the thickness of

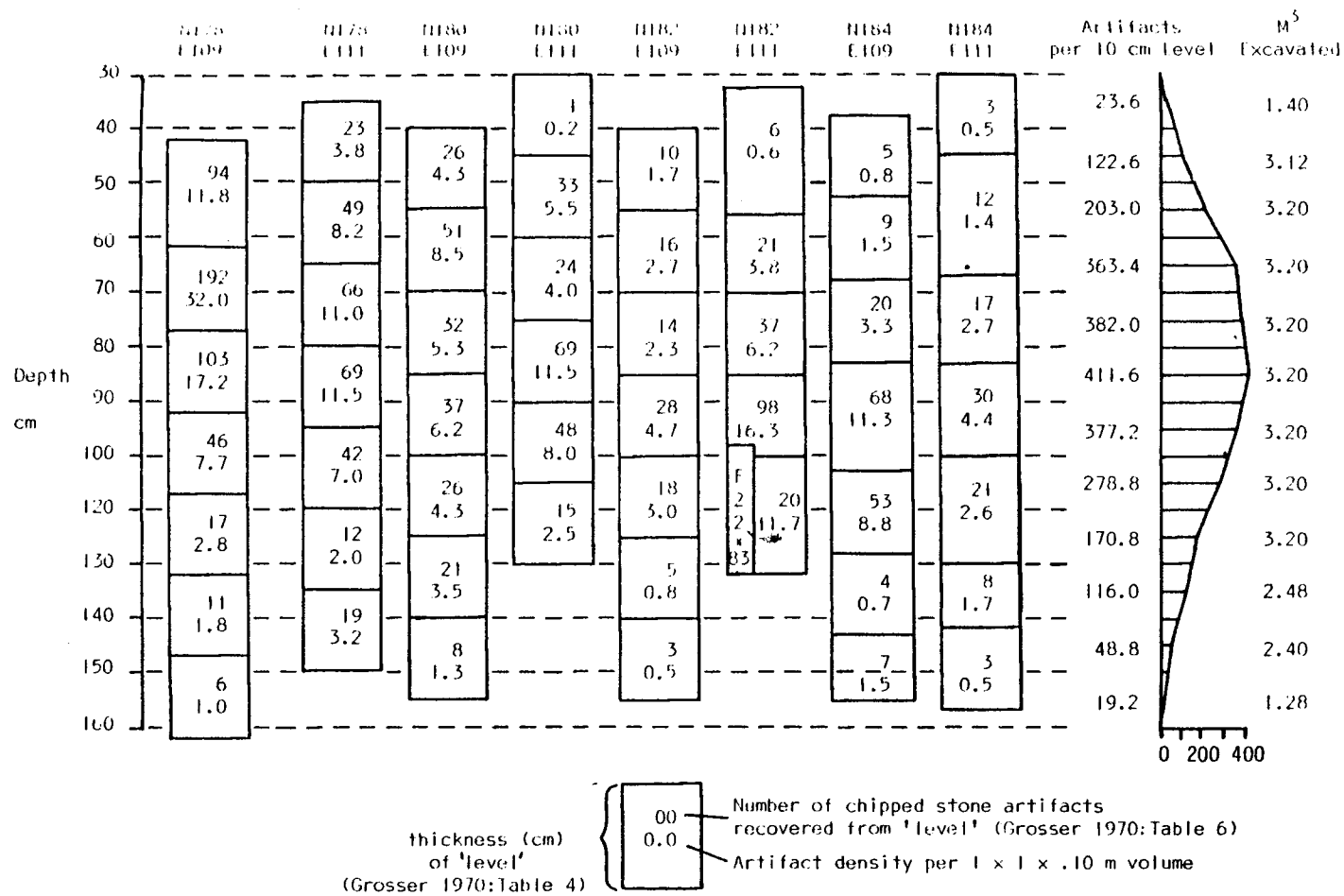


Fig. 27. Depth distribution of chipped stone artifacts from 1969 excavations at 14BU9.

individual vertical provenience units for each square (Grosser 1970: Table 4) and the number of artifacts recovered from it (Grosser 1970: Table 6), the average density of artifacts per $2.0 \times 2.0 \times 0.1$ m excavated volume is determined for each cell in Fig. 27. The number of artifacts per 0.1 m vertical interval is then estimated by summing the density figures horizontally across Fig. 27, determining weighted averages when a vertical interval intersects two cells of the same square. The resulting distribution, shown in the second to the last column of Fig. 27, reveals a single mode in artifact frequency at 80-90 cm. It is assumed that the 1970 excavation at 14BU9 encountered similar artifact distributions. Unfortunately, Grosser's report of these excavations (Grosser 1977) provides no data on artifact frequencies.

Modes in the vertical distribution of artifacts discussed above indicate levels within the slowly aggrading Snyder Soil at which particularly intensive occupations, reflected by the amounts of discarded material, occurred. Although levels of lower artifact frequency might indicate a less intensive occupation, this explanation does not adequately account for the gradual declines in frequency above and below the modes. In other words, it is difficult to envision prehistoric occupants of the site gradually increasing, then decreasing the rate of artifact discard to produce the distributions observed. The distributions are more parsimoniously viewed as representing levels of relatively intensive occupation, from which artifacts were subsequently dispersed, upward and downward, by post-depositional processes.

The vertical distributions shown in Fig. 26-27 and Table 11 mask the presence of less intensive, but nonetheless important, site

Table 11. Vertical distribution of chert flakes in 1968 block excavation at 14BU9.

Level (cm BS)	Frequency of Heat-Treated Flakes		Total Flakes ^b	Excavated Volume Yielding Flakes ^c
	Absolute ^a	Relative ^a		
35-50	22	.308	71	4
50-75	338	.246	1374	9
75-100	197	.364	541	8
100-125	256	.187	1369	7
125-150	97	.135	719	4
150-175	27	.071	380	4

^a Data from Grosser (1970: Table 3).

^b Relative frequency of heat-treated chert divided by absolute frequency.

^c In m³; data from Grosser (1970: Table 5).

occupations. Although all three excavations yielded diagnostic artifacts of the site's Walnut phase component, only the 1968 excavation reveals sufficient variation in frequency to form a Walnut phase mode in the upper portions of the Snyder Soil. However, in the 1969 excavation (Fig. 27), individual squares yielded high densities of artifacts in the Walnut phase levels. For instance, 94 artifacts were recovered between 42 and 62 cm in square N178 E109, and 51 artifacts were present between 55 and 70 cm in square N180 E109. Although these small, vertically and horizontally restricted clusters are probably associated with the Walnut phase occupation, their presence is masked by the higher densities of the El Dorado phase component, which forms the most prominent mode at 80-90 cm.

The gradual decline in artifact frequencies above and below modal levels is reasonably attributed to pedoturbational processes. Observations made during the 1978-1979 excavations suggest the operation of at least three pedoturbational agencies: burrowing animals, tree roots and dessication cracks (Leaf 1981; Haury and Leaf 1983).

Excavators identified numerous rodent burrows in the Snyder Soil at 14BU9 (Haury and Leaf 1983). Burrowing fauna recovered as bone from 14BU9 include gophers (Geomys bursarius), voles (Microtus ochrogaster), moles (Scalopus aquaticus) and ground squirrels (Spermophilus spp.), all of which were probably intrusive into the prehistoric deposit. Archaeologically-recovered bones of these taxa have relatively well-preserved surfaces, indicating less extensive attrition than remains of rabbit, bison, antelope and deer, which were butchered and used for food by the prehistoric inhabitants of the site (Haury 1981:412, 426, Table 8.6). This inference is supported by Grosser (1977:100) who states that mole remains were recovered from filled rodent burrows during the 1970 excavations.

Burrowing rodents and other soil fauna such as earthworms and ants are effective pedoturbational vectors (Wood and Johnson 1977:318-328; Buol et al. 1980:260-261; Thorp 1949). The collapse of abandoned burrows and tunnels causes soil (and artifacts contained within the soil) to subside. Mammalian burrowers bring artifacts and soil to the surface, while earth brought to the surface by ants and earthworms can bury surface materials.

Tree root traces were recognized in the Nuttall and Snyder Soils at 14BU9 by iron oxide staining surrounding filled-in root channels

(Leaf 1981:179; Limbrey 1975:82). The growth and decay of roots results in vertical and horizontal displacement of soil and artifacts (Wood and Johnson 1977:328); more extensive disturbance results from tree fall (Limbrey 1975:286-290), but no evidence of this latter vector is known from the Snyder Soil.

Frost-heaving and shrinking/swelling of clay are other pedoturbational vectors which have probably been effective in redistributing artifacts. Although soil properties reported in this study do not directly reflect these vectors, archaeological evidence may provide evidence of their operation. For example, the recovery of brick fragments from a depth of 110-120 cm in the Snyder Soil at 14BU9 (Haury and Leaf 1983) is most likely attributed to transport of artifacts via dessication cracks. Johnson et al. (1977) and Wood and Johnson (1977) discuss artifact movements caused by shrink-swell and frost heaving.

The radiocarbon and archaeological evidence discussed above demonstrates that the Snyder Soil at 14BU9 formed under conditions of cumulation and homogenization between 4000 and 2000 B.P. During this interval, slow aggradation and pedogenesis proceeded simultaneously, as increments of sediment were incorporated into the soil. The slowly aggrading soil provided surfaces on which relatively dense concentrations of prehistoric debris accumulated. These deposits were subjected to pedoturbation as aggradation continued, and as a result, materials deposited on discrete surfaces were dispersed upward and downward, resulting in distributions of artifacts which gradually decline above and below levels of modal artifact frequency.

It is assumed that the Snyder Soil formed under the same conditions of cumulation and homogenization throughout the East Branch Walnut valley. Although no excavated archaeological site in the study area provides long-term radiocarbon and cultural stratigraphic records comparable to those from 14BU9, all deposits excavated in the upper horizons of the Snyder Soil provide evidence of homogenization. At 14BU81 (Fig. 17) and 14BU25 (Root 1981b:30), artifacts in the Snyder Soil are diffusely distributed around modal levels. At 14BU25, Root (1981b:30) recorded the presence of rodent burrows and tree root stains. At 14BU4, although no prehistoric component occurred in the Ab horizon of the Snyder Soil, rodent burrows were abundant. Homogenization was not as apparent in the buried pedon at 14BU87. Faint horizontal stratification reflecting sedimentary structure is preserved in this pedon, indicating that complete homogenization of the deposit was not achieved.

Soil Texture

With the exception of 14BU4 and 14BU9, textural properties of the Snyder Soil vary little with depth, although at 14BU81 (TP 2) and 14BU25, there are significant changes toward the base of the profile. Only at 14BU4 and 14BU9 do pronounced textural changes accompany a transition between horizons of the Snyder Soil. At 14BU4 (XU102), clay content increases from 27.6% in the upper Ab horizon to 34.1% in the ACb horizon (Appendix B, Table B.1). Similarly, at 14BU9, clay increases from 29.4% in the Ab to an ACb average of 35.3% (Appendix B, Table B.3). Increases in clay content in both pedons are accompanied by decreases in silt content, while the percentage of sand remains constant.

As discussed previously with reference to the Nuttal Soil, increases in clay content with depth can be attributed to the accumulation of illuvial clay, the in situ weathering of clay minerals or variations in the grain size of parent material (Chittleborough and Oades 1979). As discussed previously, the development of argillic horizons requires at least 1000 years of surficial stability. Accepting the arguments of preceding sections concerning the dominance of cumulation and homogenization in the genesis of the Snyder Soil, it is not likely that the surface was ever sufficiently stable to permit appreciable clay translocation.

Mineralogical data required to investigate the extent to which ACb horizon clay may have formed in situ were not acquired in this study. It is conceivable that in situ clay formation accompanied aggradation of the Snyder Soil, although the low hue and chroma of the horizons suggest, that mineral weathering has not been intensive (see below).

The textural change may also reflect a shift in deposition, with coarser, siltier sediments deposited in the upper portions of the Snyder Soil. However, mineralogical data which might be employed to test this hypothesis (Chittleborough and Oades 1979) are not available.

It is tentatively suggested that variations in clay content observed in the Snyder Soil at 14BU4 and 14BU9 are not attributable to clay translocation or extensive in situ clay formation. This conclusion rests largely on previously discussed evidence for slow aggradation of the Snyder Soil surface, a condition which would have inhibited the textural differentiation of soil horizons. It is also supported by soil color data, discussed below.

Depositional changes are most likely reflected at the base of the profiles at 14BU25 (Fig. 21) and 14BU81 (TP 2). Increases in clay and sand between 200 and 230 cm at 14BU25 (Fig. 21) may reflect transition to a lower stratigraphic unit, termed Zone D by previous workers, which was encountered in the 1968 excavations at depths greater than 230 cm (Root 1981b:35). At 14BU81 (TP 2), the increase in silt and decrease in clay at the base of the profile (Fig. 23) probably represent the transition to a lower, coarser-textured sedimentary unit. This layer was not visually identified in the 1978 excavations (Artz 1981).

Soil Color

A horizons of the Snyder Soil, in both buried and unburied pedons, exhibit soil colors of 10YR3/2. Hues of 10YR are observed to considerable depths in the AC and C horizons of all pedons. Chroma remains constant throughout the profile at 14BU25, increases to /3 at 14BU4 and 14BU9, and increases to /3 and /4 at 14BU81. Values increase to 4/ and 5/ with increasing depth at 14BU4, 14BU25 and 14BU81. Hue and chroma, as discussed previously, provide an index of the intensity of mineral weathering in soil (Buntley and Westin 1965). The lack of color contrast between horizons of the Snyder Soil thus indicates relatively weak weathering and profile development. In particular, the subsoil horizons of the Snyder Soil exhibit less well-developed soil colors than the Nuttal Soil, in which subsoil chromas ranged from /3 to /6 (Table 9). This lack of color development suggests that the Snyder Soil is less well developed than the Nuttal Soil.

Table 12. Thickness in centimeters of Nuttal and Snyder Soil A horizons.

	14BU4 XU102	Trench C	14BU9	14BU25	14BU81	14BU87
Nuttal Soil	35	-	25	30	-	8
Snyder Soil	50	84	25	55	55	-

Melanization, or the darkening of the upper horizons of soil by organic matter, is evident in the dark coloration of the Snyder Soil in both buried and unburied pedons. As the following section suggests, the Snyder Soil has suffered post-burial losses of organic matter, especially from buried pedons. Organic deterioration has undoubtedly also weakened contrasts in soil colors between the upper and lower horizons of the Snyder Soil.

Dark soil color was the basic criterion for identifying A horizons in both buried and unburied pedons of the Snyder Soil. Snyder Soil A horizons, with the exception of the 25 cm thick Ab at 14BU9, are thicker than Nuttal Soil A horizons (Table 12). The greater A horizon thickness probably reflects cumulation processes (Soil Survey Staff 1975:297).

Organic Matter

Low amounts of organic matter, as determined by the Walkley-Black method, are present in the buried A horizons of the Snyder Soil, compared to unburied Snyder and Nuttal Soil A horizons. Walkley-Black organic matter in the Ab horizons at 14BU4 (XU102), 14BU9, 14BU25 and 14BU87 does not exceed 1.0%. By contrast, in surface A horizons, organic

matter ranges from 2.6-3.0% in the Nuttall Soil at 14BU25 to 1.1-1.8% in the Snyder Soil at 14BU81.

The low concentration of Walkley-Black organic matter in the Snyder Soil Ab horizons suggests that the buried soil has suffered post-burial depletion of organic matter. In an active soil, fresh organic matter is constantly introduced, largely in the form of waste products and dead tissues of plants and animals. Once deposited in the soil, organic residues swiftly decompose. As Jenkinson (1981:550) notes, "the amount of organic matter in a soil at a given time is a function of both the rate of decay of organic matter and its rate of addition." Ideally, the rates of addition and decay are equivalent, and maintain the level of soil organic matter in dynamic equilibrium.

In a buried soil, which is effectively removed from the principle sources of renewed organic matter (Nikiforoff 1949:227), the rate of decay overbalances the rate of addition and organic matter is lost. Appreciable losses due to burial are recorded in soils for time spans ranging from 100 to 10,000 years. Comparing an unburied pedon in Iowa with a pedon of the same soil series buried for 100 years, Hallberg et al. (1978:342) found an organic carbon content of 1.4-1.7% in the upper 20 cm of the buried A horizon, compared to 2.6% for the same depth interval in the unburied A horizon. Mattingly and Williams (1962) compared A horizon characteristics of an unburied soil in Wiltshire, England, with those of a nearby soil buried for 1500 years beneath a Roman amphitheater. They found that organic carbon decreased from 2.42% in the unburied soil to 0.77-0.80% in the buried soil. Similarly, the upper six inches of a soil buried 4000 years beneath a Neolithic burial

mound in Ireland showed a 78% loss in organic carbon (4.9 to 1.1%) relative to a nearby unburied soil (Gardiner and Walsh 1966). Stevenson (1969:474) illustrated losses of similar magnitude by comparing the organic carbon contents of buried Pleistocene-age paleosols from the midwestern U.S. with those of similar, but currently active soils.

Despite demonstrable losses of organic matter, buried soils commonly retain dark coloration (Stevenson 1969; Yaalon 1971:33), as do buried pedons of the Snyder Soil. Stevenson (1969:475) suggests that some organic matter resists decomposition because it is tightly held by clay minerals. According to Greenland and Hayes (1981:11), 50-80% of the organic matter in soil is adsorbed (i.e., held by physical or chemical forces to the surfaces of colloidal particles) by clay minerals. Stevenson et al. (1957:75) provide evidence that some organic compounds are held between the layers of expandible lattice clays. Although in an active soil only a small portion of the organic matter actually penetrates clay particles in this manner (Greenland and Hayes 1981:12), those compounds which succeed become relatively inaccessible to microbial decomposers (Jenkinson 1981:551). Thus, as organic matter decomposes following the burial of a soil, it is reasonable to expect that the most tightly held organic compounds will resist decomposition for the longest period, and therefore will comprise increasingly greater portions of the organic component.

The Walkley-Black method measures only the most readily oxidized forms of organic matter, such as undecomposed tissue and humus. The method does not detect relatively inert organic constituents such as charcoal, which is nearly elemental carbon (Allison 1965:1367; Jackson

1958:205). To determine if residual organic matter held in the Snyder Soil Ab horizons could be detected by a technique capable of more complete oxidation, loss on ignition (LOI) was employed.

Although much of the weight lost during ignition is due to the oxidation of organic matter, other soil constituents are inevitably lost. The amount and source of the resultant error varies with the duration and temperature of ignition. Times and temperatures of ignition recommended by various authors (Table 13) range from 850° C for 0.5 hour to 375° C for 16 hours. Ball (1964:85) criticizes ignition at 850° C, since, at these temperatures, clay minerals begin to break down, losing structural water. Even at temperatures ranging from 450° C to 600° C, Ball (1964:86) notes that some clays lose structural water. Other weight losses result from the loss of hydroxyl (OH⁻) ions, from the decomposition of salts and from the evolution of CO₂ from carbonates (Jackson 1958:207). Jackson (1958:225), Ball (1964:86) and Smith and Atkinson (1975:174) state that inorganic weight losses from these sources are minimized by ignitions at 350°-400° C for 7, 8 or 16 hours (Table 13). Even at these temperatures, some loss of hydroxyl ions and structural water occurs (Smith and Atkinson 1975:174) and organic matter may be incompletely oxidized (Jackson 1958:207). Given the many factors which affect LOI, Smith and Atkinson (1975:174) recommend that determinations be made at several temperatures to check for variations in weight loss:

In the present study, LOI was determined at 600° C for 3 hours and 400° C for 8 hours. At both high and low temperatures, LOI reveals depth distributions similar to those obtained by the Walkley-Black method (Appendix B; Figs. 18-24). In all pedons, LOI is at or near a maximum

Table 13. Combinations of time and temperature recommended by various authors for determining loss on ignition.

Temperature	Duration	Source
850° C	9.5 hr	de Leenheer et al. 1957
600° C	3.0 hr	Wilde et al. 1964:19
375° - 425° C	8.0 hr	Smith and Atkinson 1975
350° - 400° C	7.0-8.0 hr	Jackson 1958:225
375° C	16.0 hr	Ball 1964:86

in the surface horizon, and tends to decrease with depth. Buried A horizons of the Snyder Soil show no obvious increases in LOI, relative to overlying Nuttall Soil B horizons. LOI, therefore, does not detect appreciable amounts of residual organic matter in the buried A horizons, even though the dark coloration of these horizons indicates its presence.

In the present study, neither Walkley-Black, which achieves maximum oxidation temperatures of 120° C (Allison 1965:1373), nor ignition at 400° C completely oxidized all organic matter. Following ignition at 400° C for 8 hours, samples from the Nuttall Soil A and Snyder Ab horizons at 14BU4 (XU102) exhibited dark brown (10YR3/3) and brown (10YR5/3) colors, respectively, suggesting incomplete oxidation of organic matter. Even when ignition at 400° C was extended to 16 hours for a group of samples from 14BU4, little additional oxidation was achieved. Indeed, the losses incurred by ignition for 8 and 16 hours are essentially identical (Fig. 28).

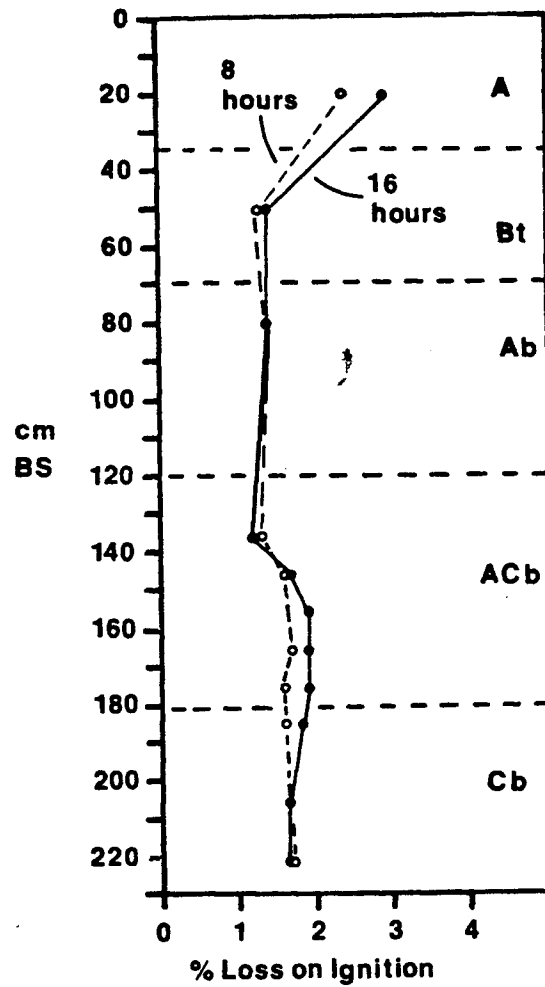


Fig. 28. Comparison of loss-on ignition results for samples from 14BU4 (XU102).

Ignition at 600° C is assumed to have completely oxidized organic matter, since the samples, when removed from the muffle furnace, exhibited reddish yellow colors (e.g., 7.5YR6/6), observed in a surface horizon sample from 14BU81). However, as discussed above, weight losses in addition to organic matter undoubtedly occurred at 600° C, and could potentially mask subtle variations in the depth distributions of organic matter.

The difficulty encountered in achieving complete oxidation of organic matter, not just within the Snyder Soil Ab, but throughout the profiles of both the Snyder and Nuttal Soils, suggests that a small portion of the organic matter is held in relatively inaccessible forms. As previously noted, organic matter and clay form strong physical and chemical associations in soil (Greenland and Hayes 1981:550-552; Brady 1974:18, 147). High clay content (30-40%) is probably an important factor contributing to the inaccessibility of organic matter in the East Branch Walnut pedons.

For purposes of the present study, it is concluded that the dark coloration of the Snyder Soil Ab horizons indicates the presence of organic matter, even though present data do not provide accurate measurement of the amount. The low organic matter content of the buried A horizons, as determined by Walkley-Black and LOI, is attributed to post-burial deterioration of organic matter. The dark colors of the Snyder Soil pedons reflect melanization which occurred during the 4000-2000 B.P. soil forming interval. Low organic matter contents of the buried A horizons probably reflect depletion as a result of decomposition and lack of new additions following burial. The dark coloration of the

buried A horizons is attributed to the presence of residual organic compounds held tightly to, and possibly within, clay particles.

In addition to melanization and post-burial degradation, organic matter data also reflect anthropogenic enhancement of the Snyder Soil. At 14BU4 (Fig. 18) and 14BU9 (Fig. 20), Walkley-Black and loss on ignition reveal increases in organic matter associated with prehistoric occupations. At 14BU4, within a single cultural stratum between 130 and 160 cm, Walkley-Black shows a small peak at 140-150 cm, while more prominent peaks in LOI occur between 140 and 180 cm.

At 14BU9, where modal artifact frequencies increase between 100 and 140 cm (Fig. 26c,d), Walkley-Black shows a small peak at 125-230 cm (Fig. 26a). LOI at 400° C shows no definite mode (Fig. 26a). A decrease at 115-120 cm is anomalous since it occurs within a cultural layer where an increase in organic matter is expected. At 600° C, no prominent modes were noted in the Snyder Soil, although decreases from 3.36 to 3.31% at 90 cm, and from 3.41 to 2.85% at 120 cm may reflect decreases in anthropogenic carbon. These decreases occur immediately below levels of maximum artifact frequency. The lower decrease, at 120 cm, corresponds to the El Dorado phase levels identified by artifact distributions (Fig. 26). The upper shift is not associated with discernible changes in artifact distribution in the 1979 block (Fig. 26), but occurs near the base of the zone between 40 and 85 cm which Grosser (1970, 1973) identified as the Walnut phase component of the site. No shifts in organic matter corresponding to buried occupations are noted at 14BU25, 14BU81 or 14BU87 (Figs. 21-24). This suggests that occupation at these sites, at least in the vicinity of the sampled pedons,

was less intensive than at 14BU4 or 14BU9, resulting in lesser quantities of anthropogenically derived organics.

Figs. 18 and 20 suggest that LOI reveals more pronounced patterns of anthropogenic organic matter than the Walkley-Black method, probably due to the inability of Walkley-Black to detect "inert" or elemental forms of organic carbon such as charcoal (Allison 1965:1373). The increases in LOI within occupational levels may indicate increases in the concentration of finely divided charcoal particles, such as were macroscopically observed at both 14BU4 and 14BU9 during excavation. These results suggest the potential value of LOI as an indicator of the presence and intensity of buried prehistoric occupations. The procedure is simple, requiring no great laboratory expertise and only a few items of specialized equipment. Aside from its simplicity, the principle advantage of LOI in archaeological applications is its ability to detect inert forms of carbon such as charcoal. These forms, which more widely used techniques such as Walkley-Black do not detect, are perhaps of the greatest significance to archaeologists, since their presence in a site often directly reflects prehistoric activity.

Available Phosphorus and pH

Available phosphorus consists of those readily soluble forms of soil phosphorus which are most easily utilized by, and thus "available" to, plants. Only a small portion of the total soil phosphorus occurs in available form. The remainder is combined in sparingly soluble organic and inorganic compounds.

Fig. 29 illustrates the reactions which occur among phosphorus forms of differing solubility. Soil phosphorus occurs in solid and liquid

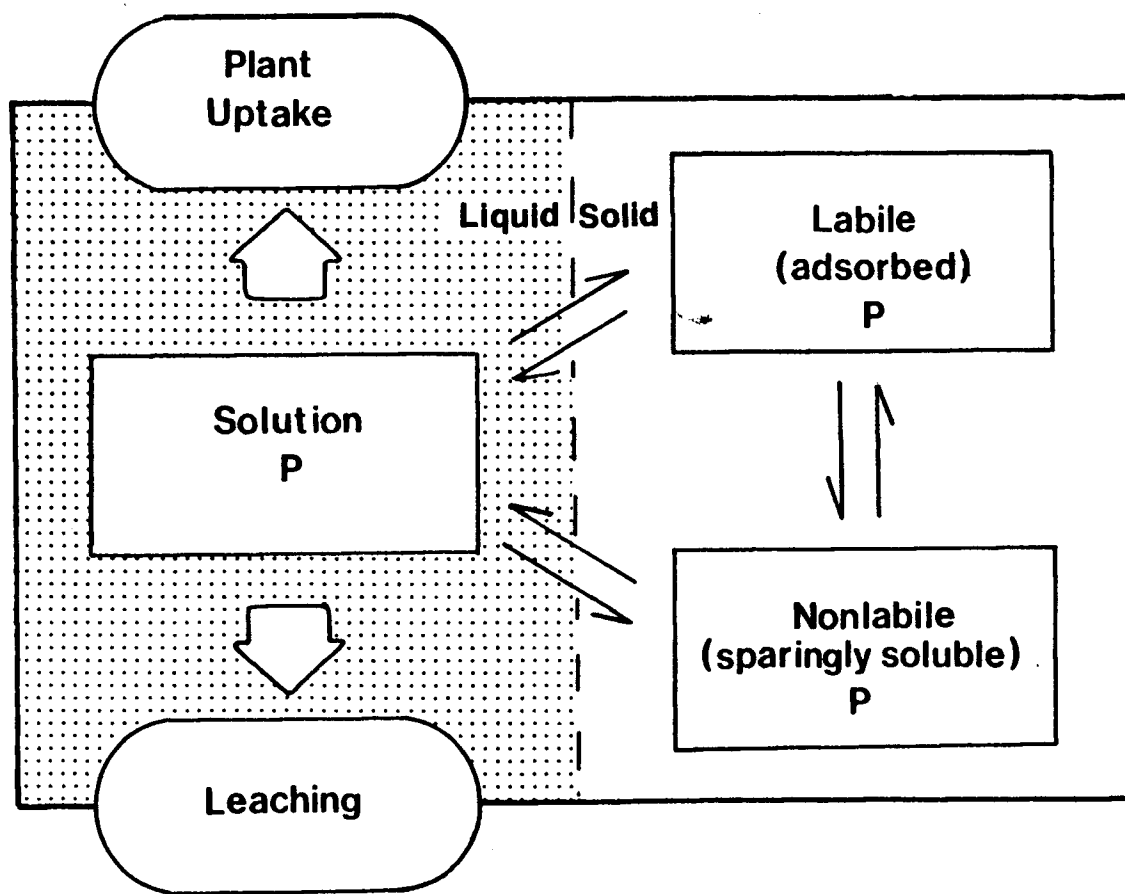


Fig. 29. Reactions among phosphorus forms of differing solubility.

phases (Larsen 1967:156-157). The liquid phase is comprised of phosphate ions and phosphorus salts dissolved in the soil solution (Brady 1974:386-361; Larsen 1967:156-157). Solid phase phosphorus occurs in labile and nonlabile forms. Labile forms are adsorbed at the surfaces of colloidal particles of clay and humus (Larsen 1967:160-161). The term labile refers to the susceptibility of adsorbed phosphorus to enter the soil solution (Larsen 1967:183-184). Phosphorus in solution and adsorbed phosphorus are available to plants and are the forms measured by the Bray-1 procedure used in this study to determine phosphorus (Smeck and Runge 1971:956-957). Nonlabile phosphorus includes phosphate minerals (predominantly apatite) inherited from soil parent material (Brady 1974:458), organic phosphorus contained in plant and animal residues (Brady 1974:460) and sparingly soluble phosphate compounds of iron, aluminum, manganese and calcium (Larsen 1967:185). The latter form precipitates from solutional and adsorbed forms and is often referred to as secondary inorganic phosphorus (Williams and Walker 1969).

By far the largest portion of soil phosphorus is in nonlabile form. Organic forms alone can account for 20-80% of the total soil phosphorus (Larsen 1967:154). Phosphate inherited from parent material weathers through time to negligible quantities, and is rapidly fixed in organic and secondary inorganic forms (Williams and Walker 1969:217-218; Walker and Syers 1976:5-6).

Although not abundant at any single time, the available (solutional and labile) forms are nonetheless important, since it is through gains and losses of these forms by translocation that anisotropic depth distributions of phosphorus develop. As shown in Fig. 29, phosphorus

dissolved in soil solution is redistributed by plants, whose roots absorb phosphorus and transport it upward, and by soil water, which leaches phosphorus both downward and laterally (Runge and Riecken 1966; Smeck 1973). Because large portions of the total soil phosphorus occur in sparingly soluble forms, phosphorus translocation is a slow process. Little movement occurs in time spans of less than 1000 years. However, over periods of 1000 to 1,000 years, large quantities of phosphorus are translocated (Smeck and Runge 1971:958).

The soil phosphorus system (Fig. 29) maintains an equilibrium between the solid and liquid phases (Larsen 1967:158). Losses from solution are replaced by releasing adsorbed phosphorus and dissolving nonlabile forms. Additions to soil solution are adsorbed or precipitated in solid-phase forms (Larsen 1967:166).

Two factors influence the proportions of phosphorus maintained in the above forms. The amount of labile phosphorus is controlled by the capacity of colloidal particles to adsorb phosphorus. When the adsorption sites on the surfaces of these particles are filled, phosphorus must either precipitate in secondary inorganic form, or remain in solution where it will be subject to plant uptake or leaching.

Reactions within the soil phosphorus system are also highly dependent on soil pH, which determines the form and solubility of nonlabile phosphorus present in soil. Phosphorus is most soluble at pH's of 6.0-7.0; the amount of readily soluble (available) phosphorus is greatest in soils of this pH range (Brady 1974:390). As pH decreases below 6.0, increasing amounts of available phosphorus are precipitated as sparingly soluble compounds of iron, aluminum and manganese. As pH increases

above 7.0, phosphorus is precipitated as sparingly soluble calcium phosphates (Brady 1974:390; Smeck 1973:199).

Before discussing evidence in the Snyder Soil for phosphorus translocation, pH-related transformations must be considered. Decreases in available phosphorus at depth in the Snyder Soil at 14BU4 (XU102) and 14BU87 are explained by changes in pH. At 14BU4 (XU102), available phosphorus decreases from 62 ppm at 170-180 cm to 43 ppm at 180-190 cm, as pH decreases from 7.3 to 6.0 (Fig. 18; Appendix B, Table B.1). At the lower pH, iron, aluminum and manganese are more active in the soil solution, combining more readily with soluble phosphorus to form insoluble compounds. The presence of abundant small concretions of iron and manganese in the Snyder Soil C horizon at 14BU4 (see pedon description, Chapter 3) confirms increased activity of these elements at depth at the site. Such concretions form by alternate wetting and drying of soil, generally in the presence of fluctuating water tables (Runge and Riecken 1966:630; Drosdoff and Nikiforoff 1940). The form of phosphorus in the Cb horizon at 14BU4 is thus influenced by ground water fluctuation.

The influence of pH on phosphorus availability is also observed in the Snyder Soil at 14BU87. Between the 60-70 and 70-80 cm levels, available phosphorus decreases from 19 to 7 ppm, as pH increases from 6.5 to 7.6. Available phosphorus and pH return to 25 ppm and 6.3, respectively, at 80-90 cm (Fig. 24; Appendix B, Table B.7). The decrease in available phosphorus is possibly due to the precipitation of calcium phosphates at the higher pH. It is likely that the 70-80 cm sample was taken from one of the whitish bands which precipitated from the Snyder

Soil at 14BU87 upon exposure to air (see Chapter 3). The color of the bands is attributed to soluble salts which were precipitated as the excavated profile dried. These salts are a probable source of phosphorus-fixing calcium ions.

With the two exceptions noted above, pH of the Snyder Soil varies little with depth and remains largely within the range of maximum phosphorus availability (pH 6.0-7.0). Soil pH decreases slightly with depth at 14BU4 (Trench C) and 14BU9 (Fig. 19-20), while at 14BU25 and 14BU81 (Figs. 21-23), pH increases slightly with depth. In all these pedons, available phosphorus tends to increase with depth. Since no marked changes in pH accompany the latter increases, processes other than pH-related transformations influence available phosphorus distributions in the Snyder Soil.

As mentioned previously, soil phosphorus is redistributed by plants and soil water. Phosphorus transported upward by plant roots, is incorporated into living tissue and is returned to the surface horizon as organisms die. The constant removal of phosphorus by plants and leaching results in the depletion of phosphorus between 30 and 60 cm below surface. In this zone, uptake from plants is at a maximum. Above this zone, in the surface horizon, phosphorus values are high, due to organic decomposition and, in cultivated soils, additions of agricultural fertilizer. Below the zone of depletion, phosphorus gradually increases, reaching maximum values at depths of 1 to 2 meters. This zone is beneath the depth penetrated by most roots and also received phosphorus leached downward by soil water (Runge and Riecken 1966:629). The above

observations are summarized in Table 14.

Depth distributions of phosphorus similar to those observed by Runge and Riecken (1966) have been documented in soils from Iowa, Illinois and Nebraska (studies cited by Smeck 1973:201), New Zealand (Runge et al. 1974; Walker and Syers 1976) and Scotland (Wells and Saunders 1956). The same translocation processes inferred by Runge and Riecken (1966) plausibly account for the distributions observed in all these soils (Smeck 1973:201).

Phosphorus redistribution by the processes summarized in Table 14 is reflected to some extent in all the East Branch Walnut pedons, although the patterns are obscured at some sites by other processes which influence phosphorus content, such as pH-related transformations and anthropogenic enhancement. In addition, the shallow burial of the Snyder Soil by the Nuttal Soil at five of the seven pedons imposes further difficulties, since an attempt must be made to distinguish to what extent phosphorus in the Snyder Soil has been modified by processes operating after 2000 B.P.

The pedon which most closely matches the model of phosphorus distribution in Table 14 is 14BU25 (Fig. 21). Available phosphorus is high in the A horizon of the Nuttal Soil, due to the phosphorus present in plant residues and agricultural fertilizers. Minimum values in the Nuttal B and the Snyder Ab horizons reflect maximum losses by plant uptake and leaching. No accumulation of available phosphorus is noted in the Snyder Ab, although according to Table 14, an increase might be expected. Reasons for this deviation are discussed later in this section.

Table 14. Characteristics of the depth distribution of soil phosphorus (after Runge and Riecken 1966; Smeck 1973).

Depth below Surface (cm)	Horizon	Relative Phosphorus Concentration	Pedogenic Processes	Net Effect of Processes
0-30	Upper A	High	Organic decomposition and fertilizer additions contribute P	P accumulates
30-60	Lower A, Upper B	Minimum	Plant uptake and leaching losses of P are maximized	P is depleted
60-100	Lower B	Increase	Losses due to plant uptake decrease as rooting depths of most plants are exceeded	Less intensive P depletion
100-200	Lower B, Upper C	Maximum	Plant uptake is minimal, leached P accumulates	P is maintained or enhanced
200+	C	Variable ^b	-	-

^b Controlled by content and weathering stage of P in parent material (Smeck 1973).

Available phosphorus steadily increases with depth at 14BU25, reflecting the lessening intensity of plant removal.

The concentration of available phosphorus at depth is probably a product of pedogenesis during the 4000-2000 B.P. soil forming interval. If the increase in available phosphorus with depth was a product of pedogenesis during the Nuttal soil forming interval, then the increase should begin higher in the profile, at depths below surface of 60 to 100 cm (Table 14). However, the increase begins at 170 cm, which is 80 cm below the top of the Snyder Soil. Thus, the lower maximum most likely developed below the Snyder, rather than the Nuttal, Soil surface.

The available phosphorus distribution at 14BU81 also conforms to the expected pattern. In TP 1 (Fig. 22), available phosphorus is relatively high in the surface horizon, decreases to a minimum at 20-30 cm, then increases to a maximum at 100 cm, the base of the excavation. In the TP 2 profile (Fig. 23), no surface maximum is noted. From minimum values at 20-30 cm, available phosphorus increases steadily to the base of the profile at 160 cm. A slight peak in the distribution at 50-70 cm may indicate a lower maximum associated with decreased intensity of plant use. Although the peak value in TP 2 occurs at a shallower depth than was observed in TP 1, it is recalled from Chapter 3 that the upper 25-30 cm of the Snyder Soil at TP 2 has been eroded in modern times. Thus, the peak presently observed between 50 and 70 cm probably occurred between 75-80 and 95-100 cm beneath the original ground surface. This depth agrees more closely with the model's expectations (Table 14). Erosion also accounts for the lack of surface accumulation of available phosphorus in TP 2.

In the Ab horizons of buried Snyder Soil pedons at 14BU4 (XU102), 14BU9, 14BU25 and 14BU87, available phosphorus reaches or approaches minimum values. This contrasts with Nuttal Soil A horizons and unburied Snyder Soil A horizons (with the exception of 14BU81, TP 2), in which high values of available phosphorus are noted. A horizons generally show high concentrations of phosphorus, due to enrichment from organic residues (Table 14). Such accumulations often persist following burial. For example, the upper and lower limits of paleosols buried in Pleistocene loess in New Zealand were identified on the basis of total phosphorus distribution (Runge et al. 1974). Total phosphorus was relatively high in the upper Ab of each paleosol, then declined to a minimum before increasing to a maximum near the base of the paleosol. The top of the next underlying soil was indicated by an abrupt decrease in total phosphorus between the Cb horizon of the upper soil and the Ab horizon of the lower soil (Runge et al. 1974:Figs. 1-2).

Such patterns are not noted in transitions from the Nuttal to the Snyder Soil at 14BU4 (Fig. 18) and 14BU25 (Fig. 21). At these sites, the Snyder Ab possesses near-minimum values of available phosphorus. Slight increases in available phosphorus in the upper samples of the Snyder Ab at 14BU9 (Fig. 20) and 14BU87 (Fig. 24), suggest the presence of small accumulations in the Ab horizons. The low concentration of available phosphorus in Snyder Soil Ab horizons has 3 alternative explanations. Two of these involve post-burial processes which removed phosphorus after 2000 B.P., while the third involves processes which inhibited phosphorus accumulation prior to 2000 B.P.

The first alternative states that the upper A horizon of the Snyder Soil was eroded prior to deposition of the overlying sediment. As suggested for the unburied Snyder Soil pedon at 14BU81 (TP2), erosion of 25-30 cm of soil is sufficient to remove the upper A horizon maximum. Although no evidence of an erosional contact is noted at the upper boundary of the Snyder Soil, development of the Nuttall Soil Bt horizon may have obscured such a contact.

The second alternative states that plant utilization and leaching during the Nuttall Soil forming interval removed phosphorus from the Snyder Soil. The maximum depletion of phosphorus by the Nuttall Soil should occur between 30 and 60 cm beneath the Nuttall Soil surface (Table 14). The upper portions of the Snyder Ab at 14BU9 and 14BU87 are within this zone of depletion and thus are likely to have been influenced by depletion following burial. However, the upper boundaries of the Snyder Soil at 14BU4 (XU102) and 14BU25 are at 70 and 90 cm, respectively, well below the expected zone of depletion. Phosphorus at the latter two sites, therefore, should be relatively removed from extensive post-burial depletion.

A third explanation for low available phosphorus in the Snyder Soil Ab is that pedogenesis prior to 2000 B.P. inhibited phosphorus accumulation in the surface layers. Like other soil properties, phosphorus develops anisotropy only if sufficient time and suitable conditions are provided. Studies of the time required to develop available phosphorus anisotropy in New Zealand (Walker and Syers 1976) suggest that only weak anisotropy could develop during the 2000 years in which the Snyder Soil formed. At the Cherokee site in northwestern Iowa, 4000 to 6000

years are required to develop anisotropic distributions of available phosphorus (Hoyer 1980:48). In addition, processes of cumulation and homogenization further inhibit anisotropism. Runge and Riecken (1966:628) note that zones of phosphorus accumulation and depletion do not occur in Vertisols, which are dominated by homogenization processes. Most importantly, Runge and Riecken (1966:629) found that the surface horizons of poorly drained soils with high water tables in Iowa are lower in phosphorus than better drained soils. They suggest that high water tables restrict the rooting depth of plants, leading to "a concomitant increase in the recycling of phosphorus" as plants are forced to rely on a restricted volume of soil for their phosphorus requirements (Runge and Riecken 1966:629). As discussed in Chapter 6, the Snyder Soil may have formed in an environment characterized by unentrenched streams with higher water tables than the present.

The three alternative explanations considered above are not mutually exclusive, which increases the difficulty of determining which processes, or combination of processes, have influenced the Snyder Soil. For example, if available phosphorus accumulation was inhibited prior to 2000 B.P., then further post-burial depletion of phosphorus would be difficult to detect. Further analysis of the forms of phosphorus present in the pedons are required to evaluate the three alternatives postulated above.

Anthropogenic enhancement of phosphorus has undoubtedly occurred in all the pedons studied here, since all contain prehistoric components. Coinciding peaks in the distributions of artifacts and available phosphorus are noted at 14BU4 (XU102) and 14BU9. At 14BU81, increases in

available phosphorus at 100 cm in TP1 and between 140 and 160 cm in TP2 may reflect anthropogenic enhancement, since buried occupations occur at these depths (Chapter 3, Fig. 14). However, translocational processes may also have contributed to the increases. The occupations buried between 150 and 180 cm at 14BU25 and between 40 and 70 cm at 14BU87 do not correlate with phosphorus increases. Differences in amounts of phosphorus associated with occupational strata probably reflect differing intensities of site occupation; 14BU4 and 14BU9, which show more prominent peaks of phosphorus, were perhaps more intensively occupied than 14BU25, 14BU81 and 14BU87. This inference is supported by organic matter data discussed in the previous section, which suggested an identical ranking of occupational intensity.

Prehistoric occupations contribute phosphorus to soil largely in organic forms. Animal bone, which is the major anthropogenic phosphorus source preserved in Snyder Soil archaeological deposits, was recovered in significant quantities from 14BU9 and 14BU25. Only a few fragments of bone were obtained from 14BU4 (Artz 1983) and 14BU81 (Artz 1981:11). No bone was recovered from the Snyder Soil at 14BU87 (Artz 1981).

Comparisons of the distribution of bone and available phosphorus at 14BU9 and 14BU25 suggest that bone preservation is influenced by phosphorus redistribution processes. At 14BU25, the prehistoric cultural stratum is defined by a peak in chipped stone artifacts between 160 and 180 cm (Fig. 30; Root 1981b:30). As noted by Root (1981b:30), bone does not conform to this distribution. The number and total weight of bone per 10 cm level tends to increase with depth between 200 and

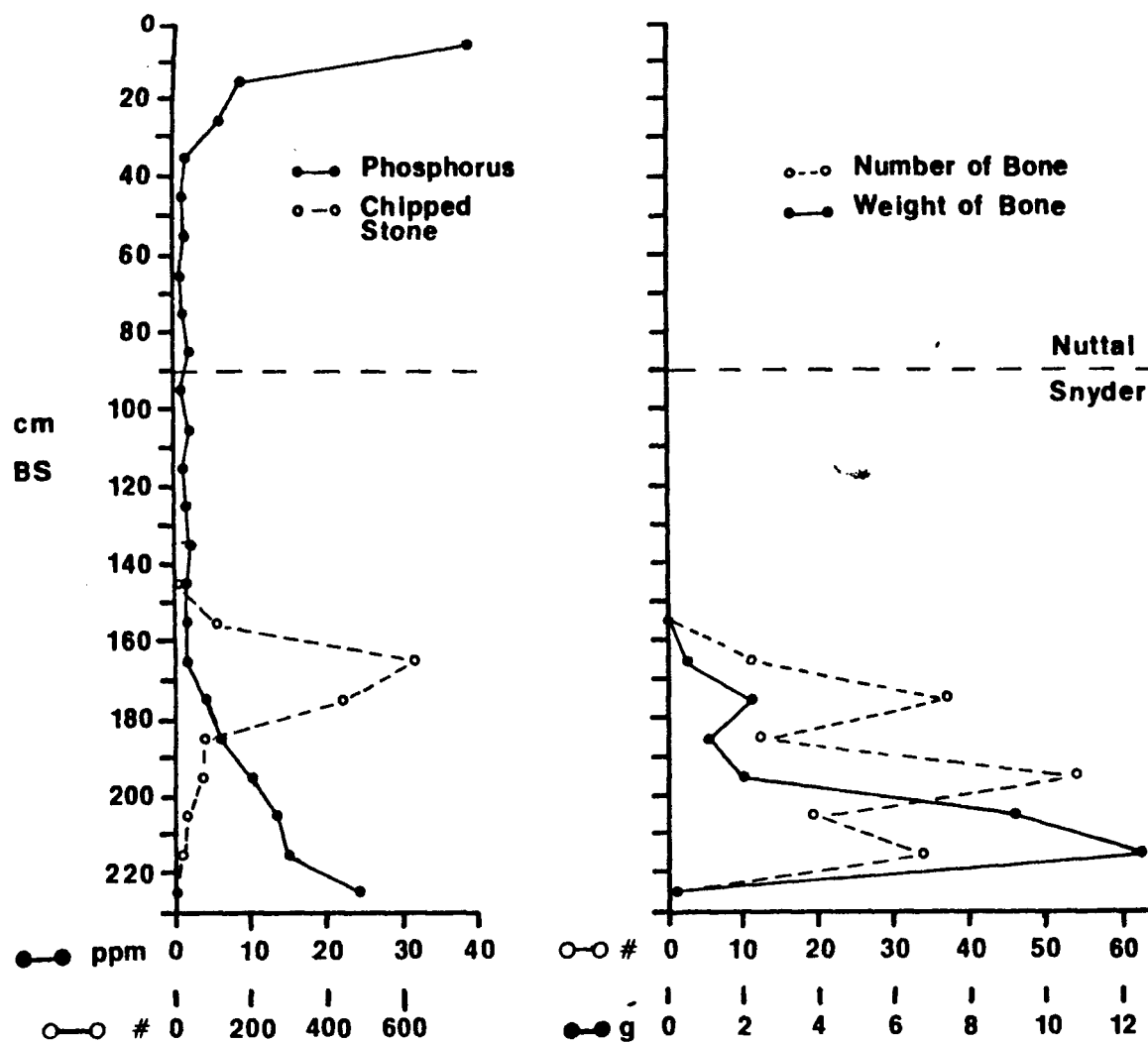


Fig. 30. Depth distribution of available phosphorus, chipped stone and bone in XU101 at 14BU25 (artifact data from Root 1981b: Table 1)

220 cm (Fig. 30). At 220 cm, the amount of both bone and chipped stone decrease to negligible quantities. The depth distribution of bone by number is a more irregular than the distribution by weight (Fig. 30), probably due to the fragility of the bone and its tendency to break into small pieces during excavation (Root 1981b:30). Weight is thus the more representative measure of bone quantity at 14BU25.

Root (1981b:30) found no archaeological evidence that prehistoric activity accounted for differences in bone and chipped stone distribution. He suggested that the anomalous distribution represented either differential preservation or non-cultural origins of the bone. Soil phosphorus data support the former alternative.

As shown in Fig. 30, increases in weight of bone with depth are matched by increases in available phosphorus. The principle inorganic constituent of bone is hydroxyapatite, a calcium phosphate mineral (Chaplin 1971:13). In bone, crystals of this mineral are arranged along bundles of collagen, a fibrous protein (Chaplin 1971:13; Hare 1980:208). As collagen decomposes, a slow process which may require thousands of years (Garlick 1969), hydroxyapatite is released into the soil. Like other calcium phosphate minerals present naturally in soil parent material, hydroxyapatite freed from bones is a sparingly soluble (nonlabile) phosphate, and given proper conditions, will dissolve and enter solutional or labile forms (Fig. 29).

The low values of available phosphorus in the upper horizon of the Snyder Soil at 14BU25 represent plant uptake and leaching losses, while increases in phosphorus below 180 cm were attributed to decreased

intensity of plant use and accumulation of leached phosphorus. In the zone of depletion, phosphorus is taken from the soil solution at a faster rate than it can be replaced from labile and nonlabile forms (Fig. 29), resulting in a constant drain on solid-phase forms. During the 4500 years following the burial of the prehistoric component at 14BU25, considerable amounts of phosphorus could be released by the decomposition of bone, dissolved in soil solution and translocated.

Given the small quantity and small size (averaging less than 1 g) of bone fragments at 14BU25, it is likely that bone contained in the cultural deposit above 180 cm may have been completely removed by weathering.

By contrast, plant utilization was less intensive below 180 cm and phosphorus levels were maintained and perhaps even enhanced by illuvial accumulations. With sufficient phosphorus in adsorbed form to maintain an equilibrium with solutional forms, the rate of dissolution of nonlabile forms, including that contained in bone, was lower. Bone was thus preserved at depth.

Data from 14BU9 provide further evidence of the influence of pedogenesis on bone preservation. Distributions of chipped stone, burned earth and limestone from the 1979 excavation (Fig. 26c-d) exhibit definite peaks between 110 and 130 cm, indicating the modal level of the relatively dense El Dorado phase component. Both bone frequency (Fig. 26d) and available phosphorus (Fig. 26b) are at minimum values in the upper 50 cm of the Snyder Soil (50-100 cm BS). At 100 cm, both increase abruptly. At 115-120 cm, available phosphorus reaches a peak value which coincides with modes in chipped stone and limestone distributions,

then declines. Bone, by contrast, does not peak, but continues to increase steadily to 140 cm, the base of the excavation.

Depletion of phosphorus from the upper 50 cm of the Snyder Soil reflects plant uptake and leaching losses, from either the Nuttall or Snyder Soil surfaces. As discussed with reference to 14BU25, phosphorus removal probably prompted release of solid-phase phosphorus into the soil solution, thus accelerating bone decomposition. The increase in available phosphorus below 100 cm thus reflects the weathering of bone from the El Dorado phase component. The available phosphorus peak, 60-70 cm below the top of the Snyder Soil, is not sufficiently deep to represent a maximum formed by the Snyder Soil, unless erosion removed the upper layers of the Snyder Soil prior to burial. The available phosphorus peak is, however, at an appropriate depth beneath the Nuttall Soil surface to represent a lower maximum of the Nuttall Soil.

Despite the fact that available phosphorus decreases with depth beneath from its peak value at 110-120 cm, bone frequencies rise to the base of the profile. Whether this is due to increases in available phosphorus with depth is not certain. At 14BU25, the increases in available phosphorus associated with a lower maximum of the Snyder Soil began 80 cm below the top of the Snyder Soil. Since the 1979 excavations at 14BU9 penetrated only 90 cm into the Snyder Soil, it is not known if a similar increase is present at 14BU9. The presence of such an increase would, however, confirm the association of the maximum at 100 cm with the Nuttall Soil.

The preceding observations on available phosphorus and bone distributions indicate that anthropogenic sources of phosphorus, such as animal

bone, function within the phosphorus system in soil. Pedogenic processes which effect transformations and translocations of naturally occurring phosphorus have an equal influence on the transformation of organic remains deposited by humans. Thus, in the case of 14BU25 and 14BU9, the preservation potential of bone varies in response to factors which determine the relative intensity of phosphorus translocation. In the upper portions of the Snyder Soil, available phosphorus was depleted due to leaching losses and plant utilization. In these levels, the dissolution of sparingly soluble forms of phosphorus (including calcium phosphates present in bone) accelerated to keep pace with the constant drain on solutional phosphorus, and thus bone was rapidly decomposed. Below the zones of maximum plant use, an equilibrium was maintained between phosphorus in solution and in solid form. This enhanced bone preservation by inhibiting the dissolution of nonlabile phosphorus.

Although archaeologists are aware of the potential problems introduced by differential preservation of bone (Binford and Bertram 1977), the relationships between bone preservation and soil chemistry are poorly understood. Poor bone preservation is often accounted for by simplistic statements referring to soil acidity (e.g., Gordon and Buikstra 1981). Even studies concerned specifically with archaeological sources of phosphorus (e.g., Proudfoot 1976; Eidt 1973; Woods 1975; Zabel 1976) do not discuss the vertical or lateral redistribution of soil phosphorus. Some authors (e.g., Zabel 1976; Proudfoot 1976) assume that soil phosphorus is not susceptible to extensive translocation. However, as shown in this study, significant translocations can occur within 4000 years

or less, with marked consequences for the preservation of anthropogenic phosphorus sources.

Serious error may result if behavioral inferences are derived from bone distributions without first considering natural transformations. For example, Grosser (1977) noted that faunal remains from the upper portions of the Snyder Soil (Walnut phase) at 14BU9 consisted solely of large bones of deer and bison, while El Dorado phase levels in the lower portions of the soil contained a diversity of small and large species. From this evidence, he inferred that a fundamental change in subsistence behavior occurred between the two phases (Grosser 1977:106-107). The present study suggests that the absence of small-boned species from the Walnut phase fauna may be a factor of poor preservation, with only the larger, denser bones surviving to the present. The soil data were, of course, not available to Grosser. However, it is suggested that the faunal assemblage requires reanalysis from this new perspective.

To summarize, the distribution of available phosphorus in the Snyder Soil reflects the operation of translocational processes which, through the agencies of plants and soil water, removed phosphorus from the upper 70-100 cm of the Snyder Soil, and allowed its accumulation in the lower profile. Elements of this pattern are obscured by pH changes and/or anthropogenic enhancement at all sites but 14BU25 and 14BU81. The accumulations at depth probably formed during the Snyder Soil forming interval, largely as a result of the reduced intensity of plant removal at depth. The low phosphorus levels in the Snyder Ab horizons may reflect the intensive loss of phosphorus due to high water tables during the Snyder Soil forming interval, or alternatively, may

reflect post 2000 B.P. erosion or post-burial depletion beneath the Nuttall Soil. Phosphorus content of the soil drastically affected bone preservation at 14BU4 and 14BU25. Bone decomposed relatively rapidly in phosphorus-depleted zones of the Snyder Soil, and is more abundant at depth, where decreases in the intensity of leaching and plant uptake increase bone preservation potential.

Pedogenesis Processes and Trends

The major pedogenic processes which contributed to the formation of the Snyder Soil during the interval 4000-2000 B.P. were 1) cumulation, or the continual addition of sediment to the soil surface; and 2) homogenization, or the mixing of soil by pedoturbation. Since both processes inhibit the differentiation of soil properties to form horizons, the dominant trend in pedogenesis during the interval 4000-2000 B.P. was toward haploidization, the maintenance of profile isotropism.

Although radiocarbon and archaeological evidence were the basis for inferring the operation of cumulation and homogenization, other soil properties support this inference. Soil colors of the Snyder Soil show only slight contrast between horizons, indicating that mineral weathering is not extensive. Textural differences between horizons of the Snyder Soil, noted only at 14BU4 and 14BU9, suggest that neither clay translocation nor in situ clay formation were prominent processes. The Snyder Soil exhibits overthickened A horizons, a characteristic of cumulative soils.

Of the Snyder Soil properties examined in this study, only available phosphorus shows anisotropic distributions. Depletion of available

phosphorus in the upper horizons of the Snyder Soil represents, in part, losses of phosphorus to plant uptake and leaching. Post-burial erosion, or plant uptake from the Nuttal Soil surface, may also account for depletion of available phosphorus in the upper portions of the Snyder Soil. Increases of available phosphorus with depth at 14BU25 and 14BU81 indicates a low intensity of plant uptake from the lower horizons of the Snyder Soil during the 4000-2000 B.P. soil forming interval.

Buried Late Archaic cultural strata at 14BU4 (XU102) and 14BU9 contributed organic matter and phosphorus to the Snyder Soil. Although buried components were present at the other three sites, no appreciable anthropogenic enhancement of soil properties was noted.

SUMMARY AND CONCLUSIONS

This chapter examined the morphological, physical and chemical properties of the Nuttal and Snyder Soils. The analysis identified pedogenic and anthropogenic processes which contributed to the development of the observed soil properties. Pedogenic processes were discussed under two headings: processes contributing to the development of the Nuttal Soil during the interval 1500-1000 B.P. to the present, and those contributing to the development of the Snyder Soil during the interval 4000-2000 B.P. In addition, post-burial modifications of Snyder Soil properties were recognized, as were modifications of both soils by organic residues from prehistoric occupations.

The processes identified by the analysis, and the properties which reflect them, are summarized in Table 15. Specific effects of the

Table 15. Pedogenic processes and soil properties of the Snyder and Nuttal Soils.

	Process or Process-Set	Process Category ^a	Process Trend ^b	Indicator Properties
I. Nuttal Soil				
Genetic Processes (ca. 1500-1000 B.P. to present)	Melanization	1	A	Dark color, high OM, high P of A horizons
	Translocation, Clay	3	A	High clay content in Bt, clay skins
	Translocation, P	3	A	Low P in Bt, relative to A
	Mineral Weathering	4	A	High CDE in Bt relative to A
II. Snyder Soil				
A. Genetic Processes (4000-2000 B.P.)	Cumulization	1	I	C-14 evidence of slow aggrada- tion; overthickened A horizons
	Homogenization (Pedoturbation)	3	I	Artifact dispersion; microfauna bones, burrows
	Melanization	1	A	Dark color of A horizons
	Translocation, P	3	A	Increases at depth, decreases in A horizons
	Transformations, P	4	A	P relative to pH at 14BU87
B. Post-genetic Processes (post-2000 B.P.)	Leucinization	2	I	Low OM in Snyder Ab
	Homogenization	3	I	Historic artifacts in Snyder Ab at 14BU9
	Translocation, P ^c	2,3	A	Increases at depth at 14BU9, 14BU87
	Erosion ^c	2	I	Absence of Ab accumulations of P
C. Interval uncertain	Transformation, P	4	A	P relative to pH in Cb, 14BU4

^a 1 = additions to pedon; 2 = losses from pedon; 3 = translocations within pedon; 4 = transformations within pedon (after the "four-fold categorization" of Buol et al. 1980:93-94 and Simonson 1959).

^b A = process promotes anisotropy; I = process promotes isotropy.

^c Conditional: the properties suggesting this process may also be accounted for by other processes.

various processes on soil properties are summarized according to whether the process: 1) adds substances to a pedon; 2) results in the loss of substances from a pedon; 3) results in the translocation or 4) transformation of substances within a pedon. The table also specifies whether the process promotes pedogenic isotropy or anisotropy.

Processes manifested by the Nuttall Soil promote anisotropy (Table 15). The A and B horizons of the Nuttall Soil are differentiated by color and textural contrasts, indicative of mineral weathering and clay translocation, respectively. The dominant trend in pedogenesis of the Nuttall Soil was horizonation, or the differentiation of soil properties to form distinct horizons.

By contrast, the formation of the Snyder Soil between 4000 and 2000 B.P. was dominated by cumulation and homogenization, which promote haploidization, or the maintenance of profile isotropism. Isotropism is also indicated by the lack of textural and color differentiation between horizons of the Snyder Soil at most sites. Despite the dominant trend toward haploidization, some anisotropic processes, including melanization and phosphorus translocation, were identified. Melanization, combined with cumulation, produced the overthickened A horizons of the Snyder Soil. Phosphorus translocation, through the vectors of plant uptake and leaching, depleted phosphorus in Snyder Soil A horizons.

Some Snyder Soil properties were altered after 2000 B.P. Organic matter in buried Snyder Soil A horizons was lost to post-burial degradation. Erosion of the Snyder Soil surface, or plant uptake from the Nuttall Soil surface may have removed phosphorus from buried pedons,

although it is equally plausible that the low phosphorus content of the Snyder Ab horizons represents intensive depletion prior to 2000 B.P., perhaps in the presence of higher water tables.

Three of the seven Snyder Soil pedons studied here are not buried. At 14BU81 (TP1 and TP2) and 14BU4 (Trench C), the Snyder Soil is the present-day surface soil, as it has been at these sites since 2000 B.P., when the remaining five pedons were buried. Because of their longer exposure at the surface, the unburied pedons might be expected to show a greater degree of profile development than the buried soils. However, properties of these pedons are similar to those of the buried pedons, differing only in the presence of surface accumulations of organic matter and available phosphorus not observed at the top of the buried soils. The unburied pedons, despite the greater age of their surfaces, have not developed the strong textural and color contrasts of the Nuttall Soil. A geomorphic explanation for the inhibition of pedogenesis in unburied Snyder Soil pedons is offered in the following chapter.

Prehistoric occupations contributed organic matter and phosphorus to both the Nuttall and Snyder Soils. Anthropogenic enhancement was particularly prominent at 14BU4 and 14BU9. Although not directly relevant to the argument of this thesis, a relationship was noted between available phosphorus content and bone preservation in buried sites, and observations were made on the potential utility of loss on ignition as a measure of anthropogenic contributions of organic carbon.

In summary, fundamental differences in pedogenesis characterized two soil-forming intervals in the East Branch Walnut valley. Since

pedogenic processes are influenced by environment, it follows that the Snyder and Nuttall Soils formed under different sets of environmental conditions. As discussed in Chapter 1, the extent of horizonation in alluvial soils is largely controlled by the potentially inhibitive influence of fluvial geomorphic processes. This concept is implicit in the definition of the process of cumulization, which assumes that pedogenic processes maintain an equilibrium with geomorphic (depositional) processes. The fact that cumulization is recognized in the Snyder Soil, but is not recognized in the overlying Nuttall Soil, constitutes direct evidence of geomorphic change at some loci in the East Branch Walnut alluvial landscape. Further evidence of geomorphic change is presented in the following chapter.

CHAPTER 5

SOIL-GEOMORPHOLOGY AND ALLUVIAL CHRONOLOGY

The individual Nuttal and Snyder Soil pedons discussed in the previous chapter did not, of course, form in isolation, but rather developed along with, and as an integral part of, the entire alluvial landscape. To determine what geomorphic parameters and processes influenced the differential development of the two soil-stratigraphic units, it is necessary to examine the soil-geomorphic and soil-stratigraphic contexts of the Nuttal and Snyder Soils.

The first section of this chapter discusses the alluvial soils of the East Branch Walnut valley from a soil-geomorphic perspective. From this perspective, the alluvial landscape of the study area is viewed as an aggregate of geomorphic surfaces, each surface differing from others in parameters such as surface age, flood frequency and drainage conditions. These parameters in turn determine the extent to which geomorphic processes inhibit pedogenesis beneath each surface, and therefore account for variations across the landscape in soil properties.

The second section of this chapter considers the stratigraphic relationships of the various soil-geomorphic surfaces. An alluvial chronology is developed which specifies the interval of time during which each surface formed. This chronology also suggests how the geomorphic parameters and processes controlling pedogenesis in the East Branch Walnut valley have changed during the past 4500 years.

SOIL-GEOMORPHIC RELATIONSHIPS

The principle geomorphic surfaces of the East Branch Walnut valley are the alluvial terrace surfaces identified in Chapter 2 (see Figs. 4, 5; also Fig. 31, this chapter). Summarizing, the floodplain (T-0) is restricted to relatively narrow surfaces adjacent to stream channels. At present, the channels of the East Branch Walnut and its tributaries are deeply entrenched beneath T-0 surfaces. The boundaries of T-0 surfaces are approximated by the distribution of the frequently flooded phase of the Verdigris series, as mapped by Penner et al. (1975).

First terrace (T-1) surfaces are situated 1.0-1.5 m above T-0 surfaces. Penner et al. (1975) identify the soils of T-1 as the Verdigris, Vanoss and Brewer series. The Verdigris series, the most extensively distributed of the three (Fig. 31), occupies broad, level or nearly level floodbasins. As discussed in a subsequent section of this chapter, the soils of this series also occur beneath levees. The Vanoss series occurs beneath slightly elevated surfaces including levees and footslopes. The Brewer series occurs in low-lying concave depressions including the backwater areas east of 14BU4 and channel scars at 14BU25 and 14BU81 (Fig. 31).

Second terrace (T-2) surfaces lie 3.5-6.0 m above T-1; soils of these surfaces are mapped as the Norge series. A third terrace (T-3) is tentatively identified from deposits of chert gravel which mantle the edges of tributary valleys (Fig. 31). These surfaces are not extensive in the study area, and are not further discussed in this chapter.

The Nuttall and Snyder Soils are soil-stratigraphic units which describe soils developed in sediments underlying the T-1 surfaces.

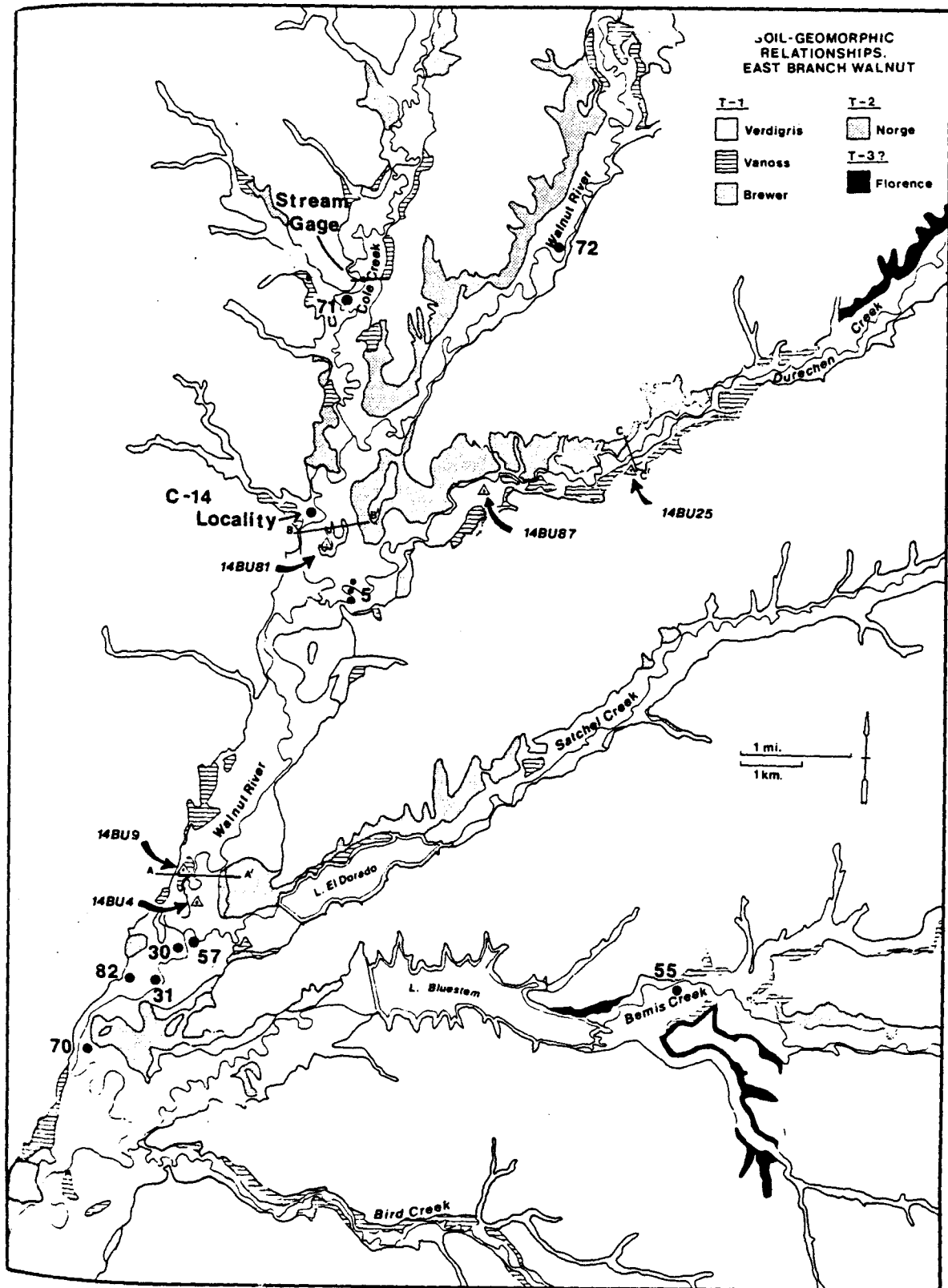


Fig. 31. Soil-geomorphic map showing location of sites discussed in text (after Penner et al. 1975).

These units were defined in Chapter 3 on the basis of pedologic comparisons and time-stratigraphic correlations among the soils of five archaeological sites. Although the two units were traced laterally at most of these sites, an attempt to map the distribution of the two units throughout the study area was not undertaken. However, both the Snyder and Nuttal Soils are comparable to specific soil series, the distributions of which are mapped by Penner et al. (1975). The Nuttal Soil, as discussed in Chapter 3, exhibits properties similar to those of the Vanoss series (Table 6). With the exception of the pedon at 14BU87, the Snyder Soil pedons described in Chapter 3 are similar to the Verdigris series (Table 7). In addition, the Brewer series is also identified with the Snyder Soil-Stratigraphic Unit on the basis of archaeological evidence from 14BU81. At this site, Late Archaic occupational strata within the Snyder Soil extend laterally from a floodbasin area where soils of the Verdigris series occur to a paleo-channel, the soils of which are mapped as the Brewer series. As discussed below, the Brewer and Verdigris series occur in specific depositional environments in the study area. In this study, the two series are considered to represent two "facies" (i.e., stratigraphically correlated but lithologically or pedologically dissimilar divisions) of the Snyder Soil-Stratigraphic Unit.

For purposes of this study, it is assumed that the mapped distribution of the Vanoss series approximates that of the Nuttal Soil-Stratigraphic Unit, and that the mapped distribution of the Verdigris and Brewer series approximates that of the Snyder Soil-Stratigraphic

Unit. This statement is an assumption and not a fact since soil stratigraphic investigations, confirming the presence of the Nuttall and Snyder Soils, were conducted in only a few small portions of the large area mapped by Penner et al. (1975). However, the assumption is supported by evidence, presented in Chapter 3, of the co-extensive distributions of the Nuttall and Snyder Soils (identified by time-stratigraphic correlation) with areas mapped as the Vanoss and Verdigris series, respectively.

In the East Branch Walnut valley, the degree of horizonation of alluvial soils tends to increase with increasing elevation above streams. Thus, the frequently flooded phase of the Verdigris series, beneath T-0 surfaces, often has thin layers of recently-deposited silt at the surface (Penner et al. 1975:21), indicating relatively recent accessions of sediment by floods. By contrast, the thick, well-developed argillic horizons of the Norge series indicate the long-term geomorphic stability of T-2 surfaces. On T-1 surfaces, at intermediate levels of the landscape, the cumulative A horizons of the Verdigris and Brewer series (the Snyder Soil-Stratigraphic Unit) reflect an equilibrium between pedogenesis and deposition. The minimally developed argillic horizons of the Vanoss series (Nuttall Soil-Stratigraphic Unit) reflect the relatively greater geomorphic stability of some landscape positions on the T-1 surface.

As discussed in Chapter 1, Walker and Coventry (1976) noted similar soil-geomorphic relationships in Australian alluvial soils (Table 1). They suggested that flood frequency is a critical parameter influencing

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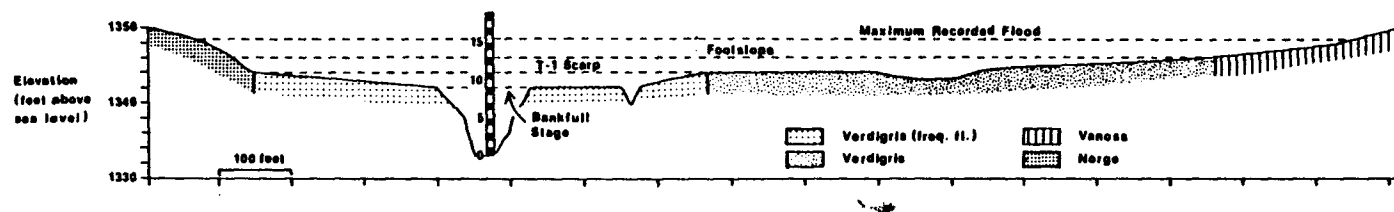


Fig. 32. Cross-section of Cole Creek valley at stream gage. For location of cross section, see Fig. 31. Source: U.S. Army Corps of Engineers, 1:200 topographic maps; Penner et al. (1975).

the creek is incised about 9 feet (2.7 m) below the T-0 surface. Penner et al. (1975) identify the soils of the T-0 surface as the frequently flooded phase of the Verdigris series (Fig. 32). East of the channel, a short scarp ca. 2 feet (0.6 m) high separates the T-0 and T-1 surfaces. At this topographic break, Penner et al. (1975) map the transition between the frequently flooded and infrequently flooded phases of the Verdigris series. The T-1 surface rises gradually towards the edge of the valley. The soils of the footslope at the edge of the valley are mapped as Vanoss series (Penner et al. 1975; Fig. 32). The geomorphic context of the Vanoss series at the Cole Creek cross-section is similar to that of the Vanoss series (and the Nuttall Soil) at 14BU25 on Durechen Creek (compare Fig. 32 with Fig. 5c). However, the presence of the Nuttall Soil was not verified at the Cole Creek cross-section. On the west side of the valley, T-0 is bounded by a relatively steeply sloping surface. This surface is probably underlain by an eroded remnant of T-2, since the soils beneath it are mapped as the Norge series.

At the Cole Creek gage, floods with stages greater than 9 feet (2.7 m) exceed the entrenched channel and spread over the T-0 surface (Fig. 32). Floods with stages exceeding 11 feet (3.4 m) inundate T-1 surfaces, while stages exceeding 13 feet (4.0 m) begin to inundate the area of Vanoss series beneath the footslope at the east edge of the valley.

The peak discharges associated with each of the above critical flood stages is estimated from a rating curve (Leopold et al. 1964:168), which plots stage against discharge (Fig. 33). The peak discharges

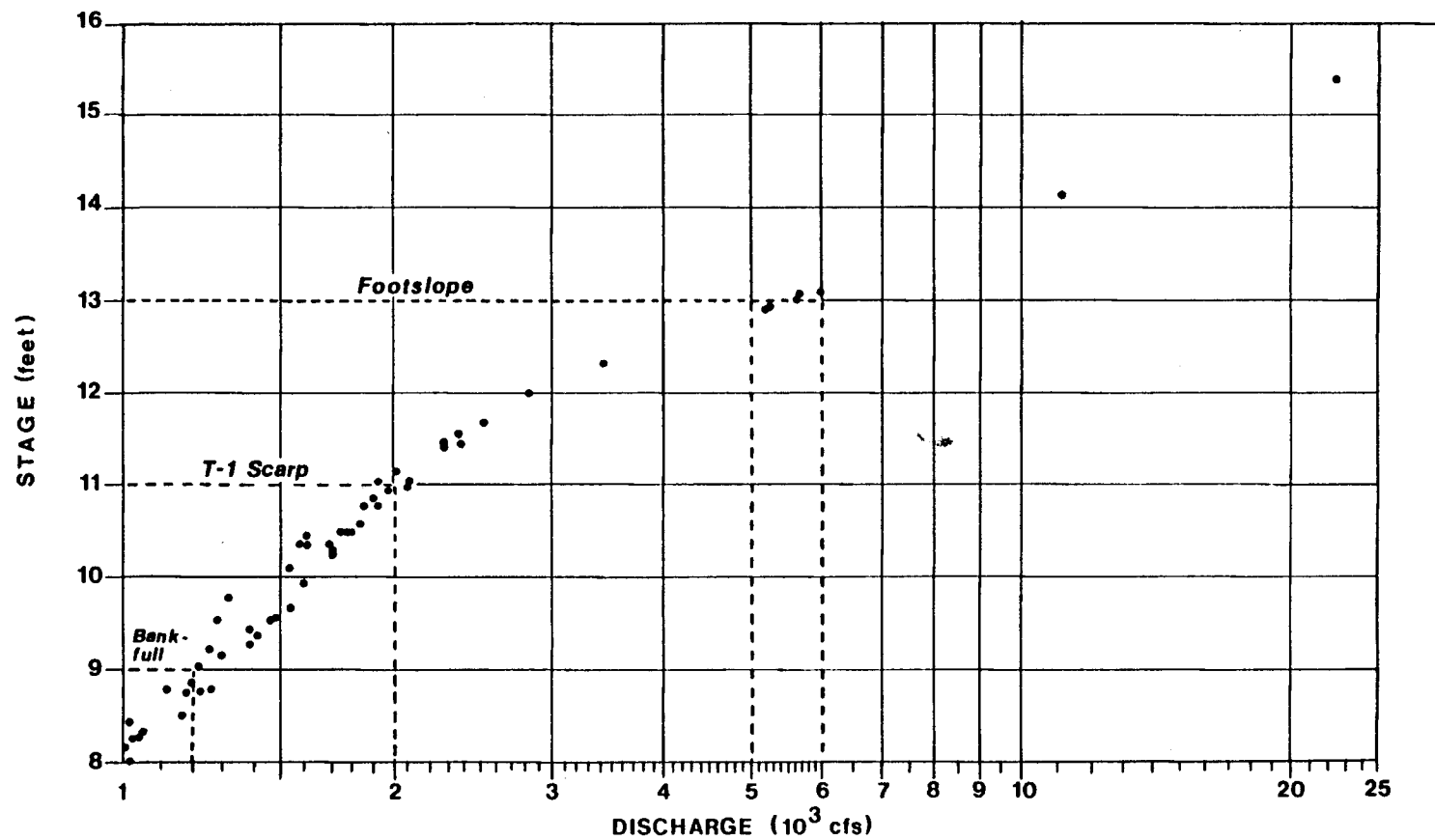


Fig. 33. Rating curve for Cole Creek stream gage.

expected for stages of 9 and 11 feet are read from the curve as 1200 and 2400 cubic feet per second (cfs), respectively. Discharges of 5000-6000 cfs are required to equal or exceed a stage of 13 feet (Fig. 33).

The frequency with which floods on Cole Creek equal or exceed the above discharges is expressed in terms of the recurrence interval of such floods. In its simplest form, recurrence interval is calculated by the formula $R.I. = N + 1 / M$, where N is the number of years of record and M is the rank of a given discharge in order of descending magnitude (M = 1 for the largest flood, M = 2 for the second largest, and so on). Thus, the seventh largest peak in a record of 20 years has a recurrence interval of $20 + 1 / 7$, or 3 years. This means that, during the 20 years of record, seven flood events equalled or exceeded the given discharge. Recurrence intervals are often used as the basis for making probabilistic statements: e.g., an R.I. of 3 years is taken to indicate that a flood of a given magnitude will probably recur, on the average, once every three years (Leopold et al. 1964:63-66). Ward (1978:83) discussed the problems involved in such interpretations of recurrence intervals.

At the Cole Creek gage, the estimated bankfull discharge of 1200 cfs has a recurrence interval of 0.44 years (Fig. 34). This means that the bankful stage should statistically have been equalled or exceeded about twice a year between 1960 and 1980. The 2400 cfs discharge necessary to overflow T-1 surfaces recurred on the average of once every 1.2 years. Because of the low frequency of floods reaching the

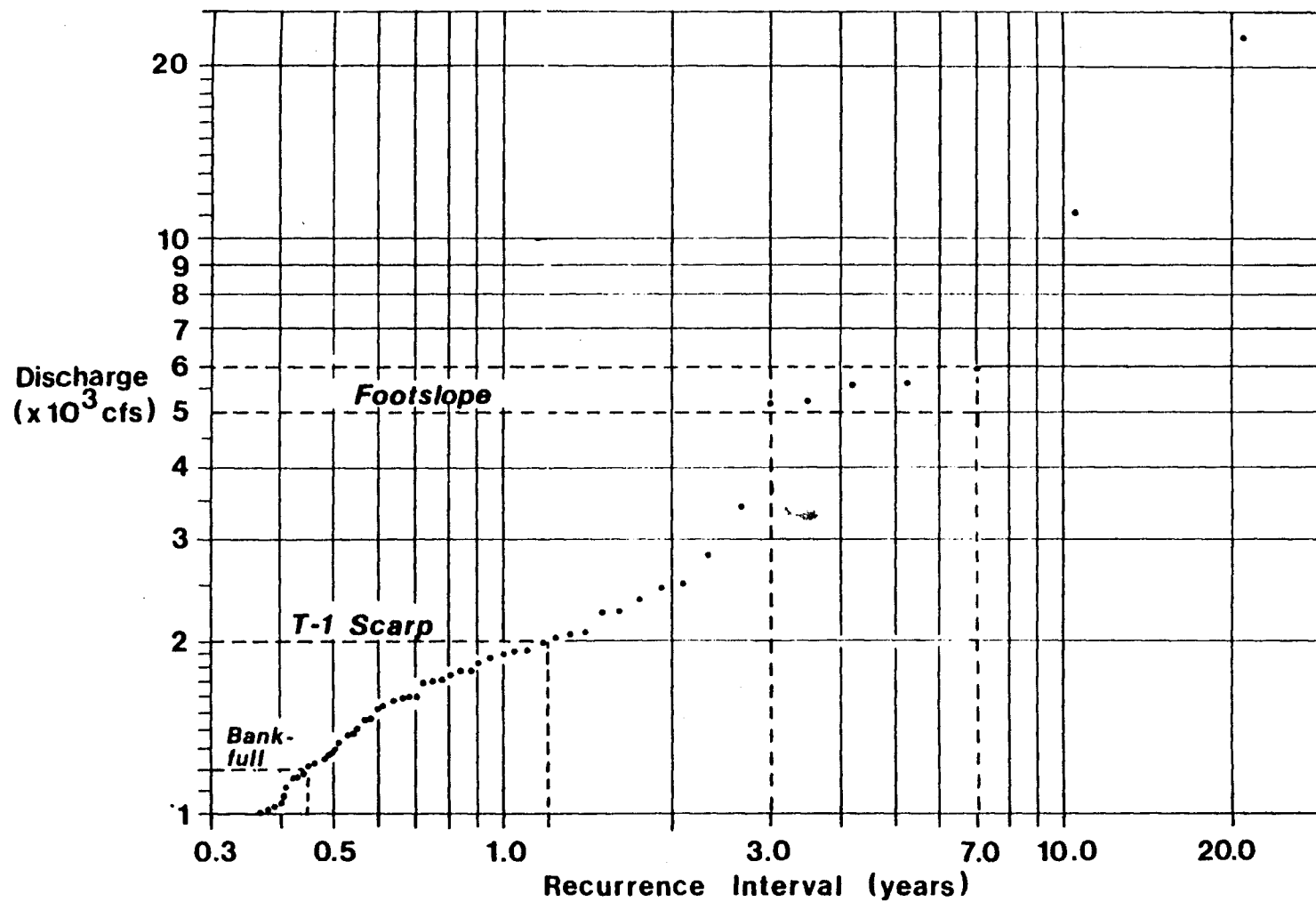


Fig. 34. Recurrence intervals of peak discharges above 1000 cfs at the Cole Creek stream gage.

valley edge footslope, the exact recurrence interval of such floods is uncertain, although a recurrence interval of 3 to 7 such floods is estimated (Fig. 34). These data suggest that, at present, T-0 surfaces near the Cole Creek gage flood at intervals of 0.44 to 1.2 years, while most T-1 surfaces flood at intervals of 1.2 to 7 years. The maximum flood of record, which reached a stage of 15.4 feet in 1979, did not completely inundate either the footslope at the east edge of the valley, or the T-2 remnant to the west (Fig. 32).

The above inferences are supported by observations made at the city of El Dorado, at the confluence of the East and West Branches of the Walnut River. According to L. A. Martindale, an engineer employed by the city (1982, personal communication), floods large enough to inundate the Walnut River valley bottoms (i.e., the T-1 surface) recurred at intervals of 4-5 years during the past 20 years. As at the Cole Creek gage, the 1979 flood at El Dorado was the largest on record; local informants claim the flood has not been exceeded since the town was founded in the 19th century (L. A. Martindale, personal communication, 1982). This flood, although large, did not inundate T-2 surfaces on the East Branch Walnut. Gary Harder, of Cassoday, Kansas, who farmed the East Branch valley in 1979, estimated that floodwaters were 8 feet (2.5 m) deep on T-1 surfaces near 14BU4 (personal communication, 1979). If his estimate is correct, the flood may have reached an elevation of 1312 feet on the T-2 scarp near 14BU4, 8 feet (2.5 m) below the T-2 surface (Fig. 6).

To summarize, informant accounts from the mainstem valley and records from the Cole Creek gage indicate that T-0 and T-1 surfaces

during the past 20 years have flooded relatively frequently, despite the depth of entrenchment of local streams. T-0 surfaces flood most frequently (as often as twice a year), while T-1 surfaces flood at intervals of 1.2-7 years. T-2 surfaces very rarely, if ever, flood.

These differences in flood frequency are reflected to some extent by soil properties. The frequent flooding of T-0 surfaces probably accounts for the layers of freshly-deposited sediment often observed at the surface of pedons of the frequently-flooded phase of the Verdigris series (Penner et al. 1975:21). The rarity of floods at the level of T-2 surfaces is undoubtedly one factor in the strongly developed horizonation of the Norge series associated with these surfaces. Flood frequency, however, does not adequately account for differences among soils of T-1 surfaces. As shown by topographic maps and cross-sections in Chapters 2 and 3 (Figs. 5, 6, 8, 10, 13), differences in elevation among areas of Vanoss, Verdigris and Brewer series are small. For example, at 14BU4, the crest of an abandoned levee, occupied by the moderately developed Nuttal Soil (Vanoss series), is only 25 cm higher than adjoining surfaces occupied by the weakly developed Snyder Soil (Figs. 6, 8). Parameters other than flood frequency must thus account for differences among T-1 soils.

Soil Drainage

Table 16 summarizes the relief, permeability, runoff and surface drainage characteristics of the three soil series of the T-1 surface, compiled from the series descriptions in Penner et al. (1975). Surface drainage reflects the collective influence of surface relief, runoff

Table 16. Soil drainage characteristics of T-1 soils.

Soil-Strati. Unit	Soil Series	Surface Drainage	Surface Runoff	Surface Relief	Subsoil Permeability
Snyder	Verdigris	Moderately good	Slow	Level and gently sloping (0-3%)	Moderate; 16.0-50.8 mm/hr
	Brewer	Moderately good	Slow	Concave; nearly level (0-2%)	Slow; 1.5-5.1 mm/hr
Nuttal	Vanoss	Good	Slow to Medium	Convex; level, gently sloping (0-3%)	Moderate; 16.0-50.8 mm/hr

Source: Penner et al. (1975: series descriptions and Table 6).

and permeability on the manner in which water flows across or soaks through a soil body. The Verdigris and Brewer series (Snyder Soil-Stratigraphic Unit) are characterized as moderately well-drained. As defined by the Soil Survey Staff (1951:171), this means that, "Water is removed from the soil somewhat slowly, so that the profile is wet for a small but significant part of the time." Slow drainage is potentially attributable to factors such as the presence of seasonally high water tables or slow permeability. The Vanoss series (Nuttal Soil) is characterized as well-drained, meaning that water drains freely, either by soaking through or running off the surface. The upper horizons of well-drained soils do not remain saturated for long periods of time (Soil Survey Staff 1951:171).

In the East Branch Walnut valley, good drainage is promoted by the convex shape and sloping surfaces of the landforms occupied by the Vanoss series (Nuttal Soil). This is reflected by surface runoff, which is slow to medium for the Vanoss series, and slow for the Brewer and Verdigris series (Snyder Soil) (Table 16). Slow runoff refers to a condition in which water, following inundation by floods or heavy rainfall, stands on the surface for significant periods before flowing away or soaking into the soil. Under medium runoff conditions, free water stands for only a short time (Soil Survey Staff 1951:167). Field observations made in 1979 during archaeological work at 14BU4 illustrate the influence of surface relief and runoff on drainage of Vanoss and Verdigris surfaces. On the day following a summer rainstorm, a field vehicle driven on the Verdigris series (Snyder Soil) surface at 14BU4 became mired in mud, while the archaeological excavations, situated on the slightly higher convex Vanoss (Nuttal Soil) surface a few meters away, were dry.

The moderately well-drained Brewer and Verdigris series (Snyder Soil) are differentiated by relief and permeability (Table 16). In the study area, Brewer soils occupy concave landscape elements such as channel scars and low-lying backwater areas. Because of their shape, many Brewer areas temporarily pond runoff from rainfall and floods. In aerial photographs of the study area dated April 3, 1973, areas of Brewer soil appear as dark patches, indicative of relatively high soil moisture, and in some places, have standing water. During the exceptionally wet summer of 1977, the Brewer areas east of 14BU4 remained

wet well into August (author's observation).

The Brewer series is also less permeable than either the Verdigris or Vanoss series (Table 16). Permeability, the rate at which water moves by gravity through saturated soil (Soil Survey Staff 1951:167), is 1.5-5.1 mm/hour in Brewer subsoil, compared to 16.0-50.8 mm/hour in Verdigris subsoil. The much lower permeability of the Brewer series subsoil reflects the clayier texture of the Brewer, relative to the Verdigris, series. The clayey texture of the Brewer series is probably a function of its concave landscape position. Brewer areas may act as seasonal "settling basins" in which clays are deposited from temporarily ponded waters. If so, then the clay textural B horizon (B2t) which characterizes the Brewer series (Appendix A) may reflect the depositional subenvironment of the soil rather than pedogenic accumulation of clay (cf. Parsons and Balster 1966). Such accumulations are not noted in the Verdigris series, which exhibits an A-C profile.

Summary

The influence of geomorphic parameters on alluvial soils of the East Branch Walnut valley is summarized in Fig. 35. Flood frequency accounts for differences in pedogenesis beneath the principle geomorphic surfaces. On frequently flooded T-0 surfaces, geomorphic processes dominate or balance pedogenic processes, depositing fresh alluvium on the cumulative profiles of the Verdigris (frequently flooded phase) soils. The Norge soils of T-2 surfaces, which rarely, if ever, flood exhibit strongly developed horizonation, reflecting long-term geomorphic stability. T-1 surfaces flood less frequently than T-0 surfaces.

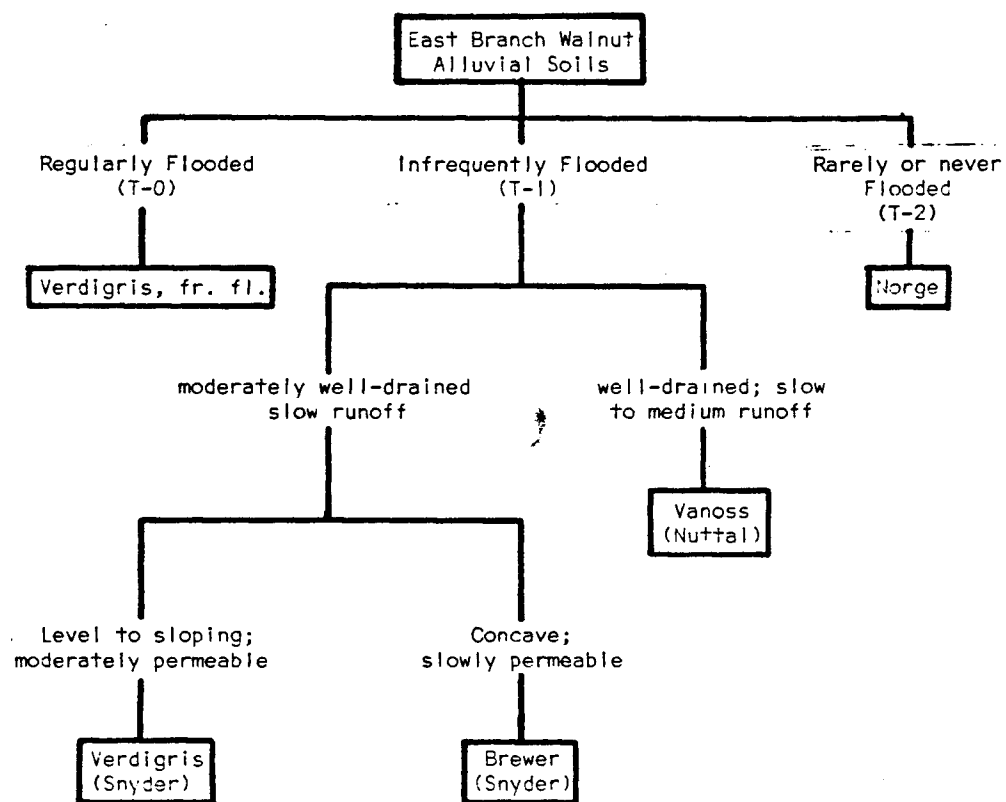


Fig. 35. Influence of geomorphology on pedogenesis in the East Branch Walnut valley.

Pedogenesis beneath T-1 surfaces at least balances and in the case of the Vanoss series (Nuttal Soil) dominates geomorphic processes. The dominant trends of pedogenesis in T-1 soils, however, are determined not by flood frequency, but by surface drainage. Soils of the Vanoss series (Nuttal Soil), associated with convex or sloping landforms, are sufficiently well-drained that when floods or precipitation occur, the pedons are not subject to long-term saturation. This allows pedogenesis to proceed relatively uninhibited. Adjacent areas of Verdigris series (Snyder Soil) are less well-drained and thus maintain weakly-developed profiles. The Brewer series (Snyder Soil) is rendered impermeable by the accumulation of clay in concave landscape positions.

ALLUVIAL CHRONOLOGY

Radiocarbon and archaeological evidence for inferring the ages of the various soil-geomorphic surfaces of the East Branch Walnut valley is discussed in this section. As shown in a schematic cross-section of the study area (Fig. 36), most of the available stratigraphic and chronological evidence relates to the formation of the T-1 surface. These data are from the five sites discussed in Chapters 3 and 4, supplemented by data from additional archaeological sites, discussed below. The soils and landforms of the latter sites have not been intensively studied but sufficient data are available from archaeological investigations to attempt correlation with the defined soil-stratigraphic units.

There is little direct evidence of the ages of the T-0 and T-2 surfaces and their underlying fills. Only one radiocarbon date was obtained from the T-0 fill. None were obtained from T-2 contexts. In

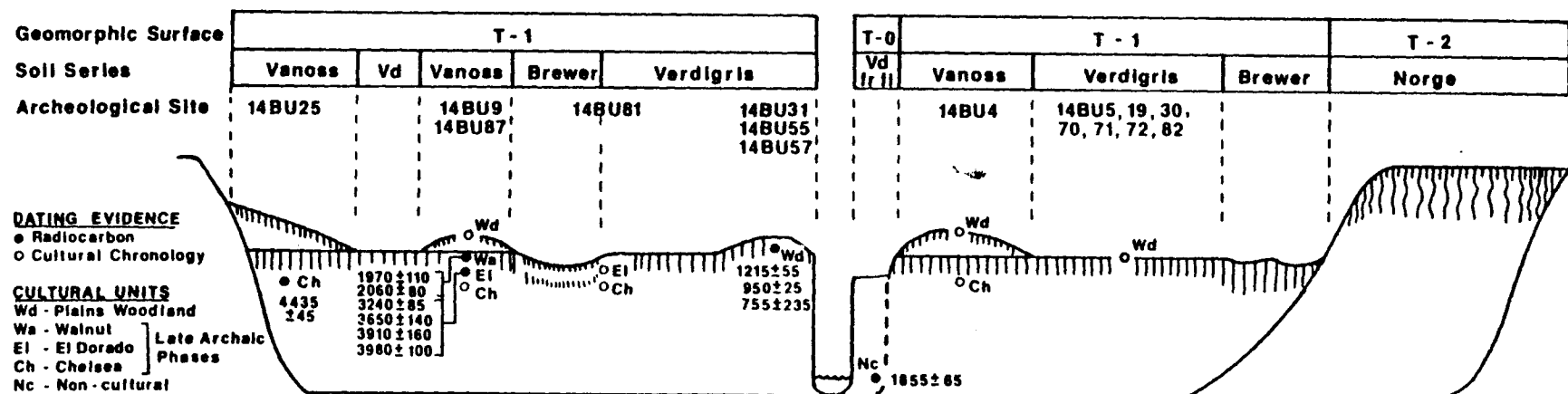


Fig. 36. Schematic cross section of East Branch Walnut valley, summarizing soil-geomorphic and soil-stratigraphic relationships.

addition, although the T-1 fill is up to eight m thick, archaeological investigations penetrated only the upper 2.0-2.5 m of the unit. The lower portions of the T-1 fill are essentially unknown.

The alluvial chronology presented below is therefore largely limited to events occurring after ca. 4500 B.P. which influenced the formation of the T-1 surface. These events are 1) valley-wide aggradation of the final 2.5-3.0 m of the T-1 fill (the Snyder Soil formed in this sediment) and 2) the areally-restricted formation of levees and colluvial footslopes on T-1 surfaces (the Nuttal Soil subsequently formed beneath some, but not all, of these surfaces). In addition, it is possible to make statements concerning 3) the incision of study area streams to their present levels, creating deep trenches. Evidence for each of these events is discussed in following sections.

Events which occurred prior to 4500 B.P. include the aggradation of the T-2 fill, the erosion of this fill and the subsequent aggradation of the T-1 fill. The quantities of sediment involved in these events is apparent in valley cross-sections (Fig. 36, and Chapter 2, Fig. 5). Based on comparisons with similar terraces on the Neosho River in eastern Kansas (Rogers 1979:250) and the Pomme de Terre River in southwestern Missouri (Brakenridge 1981), T-2 on the East Branch Walnut is undoubtedly Pleistocene, and possibly Sangamonian (cf. Brakenridge 1981:66), in age. The abandonment and erosion of T-2 and the subsequent aggradation of the T-1 fill may partially correlate with a sequence of Wisconsinan and early Holocene erosional and aggradational events posited by Brakenridge (1981:66-67, 72). Full

discussion of these events is beyond the scope of this thesis.

Valley-Wide Aggradation

Archaeological and pedological evidence from 14BU4, 14BU9, 14BU25 and 14BU81 (see Chapters 3-4) establishes that valley-wide aggradation of the T-1 fill was in progress at ca. 4500 B.P., proceeding until ca. 2000 B.P. Aggradation during this interval is assumed to have been valley-wide, given the extensive distribution of the Snyder Soil (formed prior to 2000 B.P.), as mapped by the distribution of the Verdigris and Brewer soil series.

Aggradation may have been more rapid at ca. 4500 B.P. than it was between 4000 and 2000 B.P. At 14BU4, 14BU25 and 14BU81, artifacts recovered from pre-4000 B.P. Late Archaic components tend to cluster in relatively thin strata, 20-30 cm thick (see Chapter 3). At 14BU4, 78% of all Archaic artifacts were recovered 135-155 cm below surface. In XU101 at 14BU25, 80% of all artifacts from a stratum dated at 4435 ± 100 B.P. (Table 3) were 150-180 cm below surface. All artifacts from the pre-4000 B.P. component in TP2 at 14BU81 were 130-150 cm below surface. By contrast, artifacts from the 4000-2000 B.P. components at 14BU9 showed more dispersed vertical distributions, attributed in Chapter 4 to pedoturbation. The fact that the pre-4000 B.P. deposits are not as vertically dispersed may indicate more rapid sedimentation, burying the deposits and protecting them from pedoturbation. Rapid burial is also suggested by the lack of evidence for soil formation on or beneath the pre-4000 B.P. surfaces. These surfaces, as stated in Chapter 3, were identified only from artifact distributions; they apparently were not

stable for a period of sufficient length to permit pedogenesis to occur. Artifacts deposited on these surfaces were apparently buried to a sufficient depth by the time the Snyder Soil began to form that they were not substantially disturbed by the pedoturbational processes which characterized the soil-forming interval.

The former land surfaces associated with the pre-4000 B.P. occupations of 14BU4, 14BU25 and 14BU81 all exhibit well-defined slopes (Chapter 3). At 14BU25, the 4500-year-old surface slopes ca. 60 cm in the 20 m separating XU101 from the 1969-1970 excavation block. At 14BU81, a surface of approximately equivalent age slopes ca. 50 cm in the 15 m separating TP1 and TP2 (Fig. 14). These observations are relevant to the interpretation of the radiocarbon sequence from 14BU9. Previous discussions (Grosser 1977:81) have considered the three lowest dates from the site as unreliable, due to the apparent stratigraphic inversion of a date of 4830 ± 105 B.P., recovered 178 cm below surface, with dates of 4150 ± 110 and 4600 ± 125 , recovered 250 cm below surface (Fig. 25, Table 3). Variations in slopes of former ground surfaces could easily account for the apparent 72 cm disparity in the age-depth relationships of these samples. It is conceivable that all three dates are accurate, and that the younger dates were recovered from a lower-lying surface, relative to the older date. Unfortunately, Grosser left no records of the provenience or stratigraphic contexts of the dated charcoal samples, other than their depth below surface.

Between 4000 and 2000 B.P., aggradation was relatively slow at 14BU9, with an average yearly increment of less than one millimeter of

sediment per year (Table 10). Aggradation was probably not constant, and may well have been episodic, with bursts of deposition separating intervals of non-deposition. The overall effect, however, was to provide a surface beneath which pedogenesis kept pace with additions of sediment, creating the homogeneous, cumulative A horizons of the Snyder Soil (Chapter 4).

Evidence from several archaeological sites demonstrates that valley-wide aggradation of the T-1 surface largely ceased at or shortly after 2000 B.P. For example, the Late Archaic El Dorado phase components at 14BU9 and 14BU81 are buried at nearly the same depths below the surface of Snyder Soil pedons, suggesting that the unburied pedon at 14BU81 experienced no more aggradation than the pedon at 14BU9 which has been buried 2000 years. At other sites in the study area, Plains Woodland (1500-1000 B.P.) and Plains Village (post-1000 B.P.) components occur in the upper 20 cm of unburied Snyder Soil pedons, further supporting the inference that no substantial valley-wide aggradation has occurred since 2000 B.P. The soils at these sites, although not the subject of extensive soil-geomorphic study, exhibit weakly-developed A-C profiles similar to the Verdigris series (Snyder Soil). Cultural and natural stratigraphy of these sites, located as shown in Fig. 31, are described in the following reports: 14BU30, 14BU82 (Leaf 1979b:144, 160-162); 14BU19 (Fulmer 1976:60-64); 14BU70, 14BU71 (Fulmer 1977:31, 40-43); 14BU5 and 14BU72 (Parisi 1982:19, 78, 85, 100).

The end of valley-wide aggradation of the T-1 surface probably indicates that at or soon after 2000 B.P., the Walnut River abandoned the T-1 surface, downcutting to its present level. Additional evidence

for stream entrenchment following 2000 B.P. is presented below.

Post-2000 B.P. Deposition on T-1

Evidence of post-2000 B.P. aggradation on T-1 surfaces is known only from specific, areally-restricted depositional subenvironments identified as levees and colluvial footslopes. At 14BU4, 14BU9, 14BU25 and 14BU87, aggradation occurred between 2000 B.P. (the age of the Snyder Soil surfaces buried at these sites) and ca. 1500-1000 B.P. The terminal date for the aggradation interval at these sites is inferred from the estimated age of Plains Woodland occupations on the Nuttall Soil surface at 14BU4 and 14BU9 (Chapter 3) and by the estimated time required for the formation of the minimally-developed argillic horizons of the Nuttall Soil (Chapter 4).

At none of the above sites is the Nuttall Soil areally extensive. As documented in Chapter 3, the Nuttall Soil at 14BU25 developed beneath the surface of a colluvial footslope at the edge of Durechen Creek valley. At 14BU4, the Nuttall Soil developed beneath the surface of an abandoned levee at the edge of T-1. At 14BU9, the Nuttall Soil occurs beneath a slightly elevated, convex surface located between a Walnut River paleochannel and an intermittent tributary. The depositional environment represented by this landform is not certain, but may represent a levee developed by either the mainstem Walnut or the tributary stream. The geomorphic context and areal extent of the Nuttall Soil at 14BU87 is also not precisely known, although the Nuttall Soil at this site may have developed in levee deposits of an abandoned stream channel which cross-cuts the site.

Sites 14BU31, 14BU55 and 14BU57 provide additional evidence of the post-2000 B.P. aggradation of levee surfaces. These sites, however, contrast with previously discussed examples of post-2000 B.P. aggradation in that moderately developed pedons of the Nuttall Soil are not present. Rather, excavations at these sites reveal Plains Woodland components buried at depths of 20-40 cm in Verdigris series profiles.

At 14BU55, most artifacts were recovered at depths of 30-42 cm below surface, in a soil similar to the Verdigris series. The existence of a buried prehistoric living surface at 30-42 cm is confirmed by the presence of the postmold pattern of a prehistoric dwelling in this depth interval (Adair and Brown 1981:244). Five C-14 dates from the 30-42 cm zone yielded an average age of 950 ± 25 B.P. (Adair and Brown 1981:251). The excavated areas of 14BU55 are situated on a low levee at the edge of T-1 on Bemis Creek (Adair and Brown 1981:240).

At 14BU31, on the Walnut River, the top of a limestone-filled pit, dated at 1215 ± 55 B.P., was encountered 42 cm below the surface of a levee developed at the edge of T-1 (Leaf 1979b:79, 84). In an excavation on the slope of this levee, Woodland artifacts were present only in the plowzone, indicating that the archaeological deposit is buried only beneath the crest of the levee, and is at or just beneath the surface of off-levee portions of the site. This is assumed to indicate that sediments were deposited on the levee, but not off the levee, following the Woodland occupation.

At 14BU57, a Plains Woodland component dated at 755 ± 235 B.P. was encountered at depths of 20-40 cm (Leaf 1979b:115; Brown 1981). Excavations at the site were placed on a levee developed on the upstream

side of a meander of Satchel Creek.

Since the soils observed at 14BU31, 14BU55 and 14BU57 exhibit the weakly-developed A-C profiles of the Verdigris series, they are assigned to the Snyder Soil-Stratigraphic Unit, and are assumed to have formed under the same pedogenic processes of cumulization and homogenization as other Snyder Soil pedons. The significance of this observation is that not all post-2000 B.P. aggradation created surfaces sufficiently stable to permit the formation of the moderately developed Nuttal Soil.

The reason why some but not all post-2000 B.P. depositional surfaces attained the requisite degree of surface stability is not clear. A variety of factors are potentially responsible, including variation in the age or the drainage characteristics of the surfaces. For instance, more than one episode of post-2000 B.P. aggradation may have occurred. The Nuttal Soil may have developed only beneath the oldest of these surfaces. Alternatively, subtle differences in flood frequency, runoff, relief or permeability may differentially inhibit soil formation beneath the surfaces of the various sites. Evaluating these alternatives requires additional data on the ages and geomorphic contexts of post-2000 B.P. depositional surfaces. For the purpose of the present analysis, it is sufficient to note that post-2000 B.P. aggradation on T-1 surfaces occurred only in specific areally-restricted depositional subenvironments, including colluvial footslopes and levees.

Stream Entrenchment

Since the T-0 surface and its underlying sediments have not been intensively studied, no precise chronology of its formation is possible.

As suggested above, the end of T-1 aggradation at or soon after 2000 B.P. probably signals the initiation of entrenchment. This conclusion is supported by a radiocarbon date of 1855 ± 65 B.P. (UGa-2716), obtained from a concentration of wood charcoal which was exposed in the bottom of a bulldozer trench excavated during railroad construction in 1978 (Drew 1983:10). Since no artifacts were found in or near the charcoal, it is assumed to be of natural origin. The stratigraphic section containing the charcoal was located approximately 600 m northwest of 14BU81 and 50-75 m west of the Walnut River (Fig. 31).

The charcoal concentration occurred in silty sediment only a few centimeters above limestone bedrock, and 40 cm beneath a channel fill of laminated sand and silt. The channel deposits had a maximum thickness of 78 cm and were overlain by 2-3 m of yellowish-brown silty clay. The surface layer of the section was a dark grayish brown A horizon.

Since the charcoal concentration was located immediately above bedrock, the date obtained from it is interpreted as an age for the base of the T-0 fill in the study area. This interpretation, if correct, supports the conclusion that entrenchment of study area streams occurred soon after ca. 2000 B.P., the suggested date for the end of valley-wide aggradation of T-1 surfaces. Unfortunately, the section described above provides the only chronometric evidence from T-0 contexts in the study area. For this reason, no precise chronology for the aggradation of the T-0 fill is possible. Accepting the above inferences concerning stream entrenchment, T-0 aggradation began sometime after 1855 ± 65 B.P. T-0 aggradation is perhaps still on-going, although episodic, since layers

of recently deposited silt often overlies the Verdigris (frequently flooded) pedons developed beneath T-0 surfaces (Penner et al. 1975:21).

Summary

Soil-stratigraphic and soil-geomorphic evidence, considered within a chronological framework provided by archaeological evidence, indicate that, during the past 4500 years, the East Branch Walnut valley experienced the following sequence of geomorphic events:

1) From at least 4500 B.P. to 2000 B.P., T-1 surfaces experienced valley-wide aggradation. By 4000 B.P., aggradation had slowed, permitting formation of the Snyder Soil.

2) At or soon after 2000 B.P., the East Branch Walnut and its tributaries entrenched, and perhaps flowed at nearly their present level in the landscape by ca. 1850 B.P. With entrenchment, valley-wide aggradation of T-1 surfaces ceased.

3) After 2000 B.P., deposition on T-1 surfaces occurred only on levees and colluvial footslopes. Post-2000 B.P. deposition at 14BU4, 14BU9, 14BU25 and 14BU87 occurred before 1500-1000 B.P., allowing for surface stabilization and the formation of the Nuttal Soil at these sites. Deposition occurred after ca. 1231 B.P. at 14BU31 and after ca. 1000 B.P. at 14BU55 and 14BU57. This evidence indicates that several episodes of post-2000 B.P. aggradation occurred on T-1 surfaces. However, at no time after 2000 B.P. was deposition on T-1 valley-wide.

4) T-0 surfaces and the sediments underlying them developed between ca. 1850 B.P. and the present, although detailed stratigraphic and chronometric evidence for these events are lacking.

This chronology, although sketchy, records a series of fundamental changes in the kinds and magnitudes of geomorphic processes at work in the East Branch Walnut valley. These processes shaped the alluvial landscape by eroding, transporting and depositing sediment, and in so doing, established the critical parameters of geomorphic stability and drainage to which pedogenic processes responded. Thus, for example, the pedogenic processes of cumulization and homogenization reflected in properties of the Snyder Soil represent a response to gradual, valley-wide aggradation between 4000 and 2000 B.P. Similarly, the minimally developed argillic horizons of the Nuttal Soil reflect a pedogenic response to the stabilization of convex, well-drained levees and footslopes between 2000 and 1500-1000 B.P.

In other instances, the relationship of soil properties, pedogenic processes and geomorphic parameters is less straight forward, requiring a careful consideration of soil-geomorphic contexts. As documented in this chapter, extensive areas of the Snyder Soil surface experienced little significant aggradation after 2000 B.P.; thus Snyder Soil pedons beneath these surfaces are no longer subject to the cumulization processes characterizing the interval 4000-2000 B.P. Yet, despite an additional 2000 years of surface exposure, these pedons exhibit weakly-developed profiles similar to those observed in Snyder Soil pedons buried since ca. 2000 B.P. In fact, as discussed in Chapter 4, the differences between buried and unburied Snyder Soil pedons largely reflect degradation of organic matter as a result of burial, and not increased horizonation in unburied pedons as a response to geomorphic stability.

This apparent contradiction is reasonably explained in terms of soil drainage. As discussed in this chapter, B horizon development occurred beneath some T-1 surfaces in less than 2000 years, but only in well-drained, convex landscape positions. The fact that unburied, but less well-drained Snyder Soil pedons lack B horizons suggests that drainage conditions, affecting the moisture regime of the soil, are responsible for the persistence of weakly developed profiles in Snyder Soil pedons no longer subject to cumulization processes.

A second and more troublesome paradox is presented by a consideration of soil-geomorphic and soil-stratigraphic evidence. Despite the depth of entrenchment of study area streams, and the apparent long-term persistence of this condition, the presence of levees developed since 2000 B.P. at the edge of T-1 surfaces indicates that the Walnut River and its tributaries are periodically capable of performing geomorphic work on T-1 surfaces. In fact, evidence presented in this chapter suggests that T-1 surfaces, at least during the last 20 years, have flooded at intervals of 4-5 years. Given this observation, it is perhaps surprising to find so little evidence of post-2000 B.P. aggradation on T-1 surfaces. Two explanations of this apparent paradox are possible: either the frequency of floods exceeding the T-1 terrace scarp has increased in recent times, or floodwater reaching T-1 surfaces either have not transported or have not deposited sediment. Pedologic and geomorphic evidence alone are not sufficient to resolve this problem. Evaluation of these alternatives requires a consideration of climatic, vegetational and hydrologic parameters influencing the magnitude, frequency

and sediment load of floods in the study area.

This chapter began with the simple observation that the Nuttall and Snyder Soils did not form in isolation, but rather developed as an integral part of the alluvial landscape of the East Branch Walnut valley. The chapter concludes with the observation that the alluvial landscape did not form in isolation, either, but rather was shaped by geomorphic processes which responded to a complex network of environmental processes and parameters.

Although soil-geomorphic and soil-stratigraphic evidence provide the basis for the alluvial chronology discussed above, this evidence does not directly reflect the causes of geomorphic changes recorded by the chronology. While pedogenic processes respond to geomorphic parameters inherent in the morphology of the landscape, these geomorphic parameters are themselves the product of geomorphic processes which shaped the landscape. Available pedologic evidence from the East Branch Walnut valley does not directly reflect the parameters which influenced these geomorphic processes. It is possible, however, given a model of fluvial geomorphic response to environmental change, to suggest that the East Branch Walnut alluvial chronology represents a series of responses to climatically-induced hydrologic change. This model is considered in the next chapter.

CHAPTER 6

GEOMORPHIC RESPONSE TO CLIMATE CHANGE

In this chapter, the alluvial chronology presented in Chapter 5 is viewed as a sequence of geomorphic responses to climatically-induced hydrologic change. It is argued that a shift in precipitation patterns, involving changes in the intensity and areal extent of storms, occurred in the study area at ca. 2000 B.P. This shift induced changes in the magnitude and frequency of floods, which in turn prompted morphological readjustment of study area streams. This readjustment is manifested geomorphically by evidence for stream entrenchment at or soon after 2000 B.P. Geomorphic events following 2000 B.P., including the construction of levees and colluvial footslopes and the aggradation of T-0 fill, are viewed as responses to variations in runoff and sediment yield within an essentially modern climatic and hydrologic regime.

Stream entrenchment can be induced by a number of causal agents in addition to climate. Entrenchment can occur in response to tectonic forces such as uplift and tilting which steepen the gradient of the stream, increasing its ability to erode and transport sediment (Leopold et al. 1964:474). Entrenchment can also ensue if the base-level of a stream is lowered. The lowering of base-level can occur as a consequence of a drop in sea level, or of the downcutting of a river to which the stream is a tributary (Leopold et al. 1964:258). In addition, entrenchment need not be induced by an external force, but can result from the exceedance of an internal threshold, beyond which the fluvial system is unstable and subject to change (Schumm 1973). Even when geomorphic change is induced

by change in an external variable (e.g., base-level, tectonics, climate), the specific timing of the geomorphic response may be determined by internal geomorphic thresholds (Schumm 1973:301).

These alternative explanations of geomorphic change are not considered at length in this chapter. Instead, emphasis is placed on presenting evidence which supports the hypothesis that geomorphic changes during the past 4500 years in the East Branch Walnut valley were climatically induced. Three lines of evidence, constituting the three major sections of this chapter, are discussed.

In the first section, the East Branch Walnut alluvial chronology is interpreted in terms of published models of climatically-induced geomorphic change. These models are based on theoretical concepts concerning the influence of climate and hydrology on geomorphic processes, and are supported by evidence from field studies.

In the second section, the influence of precipitation on streamflow in the present-day study area is discussed. The present-day landscape is shown to be adjusted to the currently prevailing hydrologic and climatic regimes. The different sets of geomorphic processes which were active in past landscapes of the study area are accounted for in terms of changes in the relative frequency of hydrologic and climatic events which occur at present.

The kinds of climatic changes inferred in the first two sections imply changes in upper atmospheric circulation which would have taken place on a regional scale. If the inferences are correct, then evidence of relatively synchronous geomorphic change should be found in stream

valleys throughout the surrounding region. This evidence is summarized and discussed in the third section of the chapter.

GEOMORPHIC RESPONSE

General Considerations

An alluvial landscape is primarily shaped by fluvial geomorphic processes which erode, transport and deposit sediment. Flowing water provides much of the energy for performing this geomorphic work; sediment is the material on which work is performed. Thus, the kinds of geomorphic processes at work in a landscape, and the rates at which they operate, are directly influenced by hydrologic parameters involving streamflow and sediment load. The supply of water and sediment is in turn influenced by climate, vegetation, geology and physiography (Schumm 1977:18-19).

At any one time in a fluvial system, geomorphic processes respond to the prevailing conditions of streamflow and sediment yield by adjusting, through erosion and deposition, the cross-sectional and planimetric forms of the stream channel. The particulars of these adjustments are succinctly summarized by Schumm (1969:224) and Allen (1965:94-99). Through these adjustments, streams tend to develop channel planforms and cross-sections in which the shearing (erosive) stress exerted by flowing water is balanced by the resistance to shear by sediment (Chorley 1962:5; Langbein and Leopold 1964:793).

As long as discharge and sediment load remain relatively constant, the morphology of the channel and the landscape through which the channel passes remains relatively stable. A persistent change in either parameter,

however, initiates morphological readjustment. Over time spans exceeding 100 years, alluvial landscapes experience relatively long episodes of geomorphic equilibrium, separated by relatively brief episodes of disequilibrium (Schumm 1976). During equilibrium episodes, a particular geomorphic response to prevailing hydrologic conditions is maintained. During the intervening periods of disequilibrium, morphological readjustment to changed conditions occurs (Knox 1976:182).

The alternating episodes of aggradation, stability and erosion documented by alluvial chronologies can be viewed in terms of shifts to and from geomorphic equilibrium. According to a model proposed by Schumm (1976), shown here as Fig. 37, geomorphic equilibrium is represented in alluvial landscapes by surface stability or slow aggradation. The transition to disequilibrium, signaled by erosion and downcutting in the landscape, is abrupt. The erosional interval ends in a period of relatively rapid deposition, during which the stream seeks a new equilibrium adjustment. The rate of aggradation slows as equilibrium is achieved, initiating a new episode of stability or slow aggradation (Fig. 37).

Floods play an important role in shaping alluvial landscapes. Much of the geomorphic work performed by streams is accomplished during floods, which exert forces of sufficient magnitude to move quantities of sediment (Wolman and Miller 1960:55-56). For this reason, numerous morphological parameters of alluvial landscapes, including the size, shape and planform of channels, as well as floodplain morphology, reflect adjustments to the prevailing magnitude/frequency relationships of floods (cf. Wolman and Leopold 1957:88-91); Wolman and Miller 1960:65-67; Leopold et al. 1964:319-322; Dury 1968, 1969; Woodyer 1968).

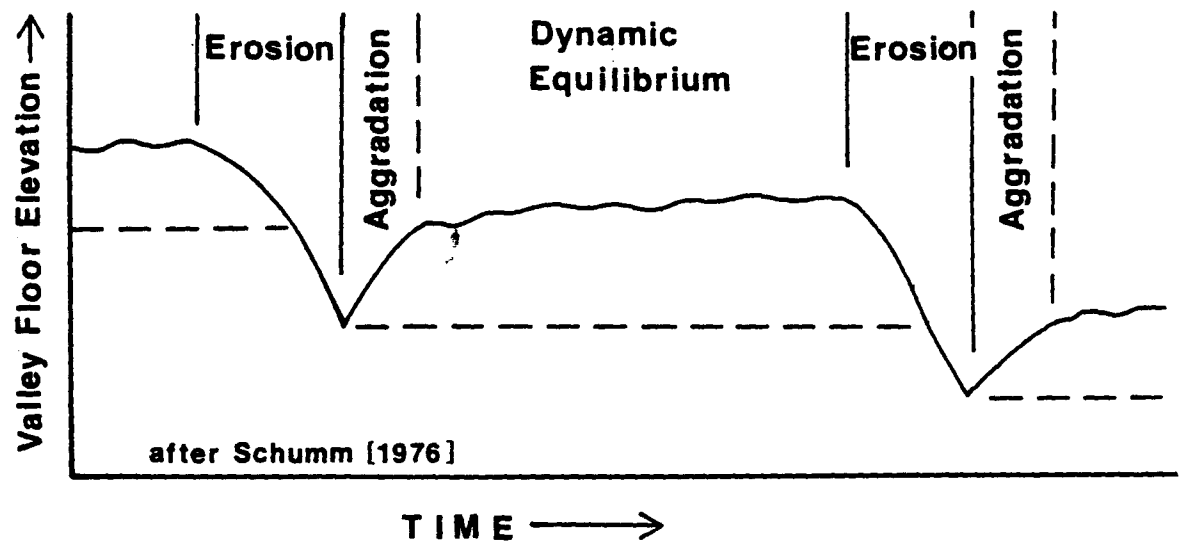


Fig. 37. Model of episodic geomorphic response in fluvial systems (after Schumm 1976:Fig. 5).

An increase in the frequency of large floods may prompt a transition from geomorphic equilibrium to disequilibrium. As suggested by Knox (1976:182), an increase in the frequency of large floods often results in the widening and/or deepening of stream channels to accommodate the larger volumes of water. This increase in channel dimension accomplishes the erosion of sediment which, according to Schumm (1976), accompanies disequilibrium.

Examples

Studies by Schumm and Lichty (1963) and Baker and Pentead-Orellana (1977) provide examples of past geomorphic change which occurred in response to changes in flood magnitude and frequency. Both studies postulate that changes in flood magnitude and frequency were induced by changes in precipitation patterns.

Schumm and Lichty (1963) studied the 20th century history of the Cimarron River in southwestern Kansas. Prior to A.D. 1914, the Cimarron flowed in a narrow, meandering channel across a grassy floodplain. In 1874, channel widths ranged from 10 to 304 feet, averaging 50 feet. Between 1914 and 1942, the Cimarron floodplain was severely eroded by a series of frequent large floods which created a wide, shallow, braided stream. In 1939, the average width of stream channels was 1200 feet (range, 200 to 3200 feet). The recurrence interval of large floods decreased after a final large flood in 1942, and the Cimarron River channel began to narrow, rebuilding a stable, vegetated floodplain. In 1954, the average channel width was 550 feet (range 100 to 2000 feet) (Schumm and Lichty 1963:73-76).

Schumm and Lichty (1963) correlated changes in flood magnitude and frequency on the Cimarron River with changes in precipitation patterns. The period of recurrent, intense floods between 1914 and 1942 was associated with a period of fluctuation between excessively wet and excessively dry years (1915-1930), followed by a period of severe drought (1931-1940). The period of channel narrowing and floodplain reconstruction after 1942 corresponded to a period of above average precipitation which did not produce large magnitude floods (Schumm and Lichty 1963:78-79). Although precipitation records for the period prior to 1915 in southwestern Kansas were not available, the climate was probably relatively moist, given evidence (Schumm and Lichty 1963:72) for perennial streamflow in the Cimarron during the late 19th century.

In explaining the influence of precipitation on channel widening and narrowing, Schumm and Lichty (1963) emphasize the importance of vegetation in promoting landscape stability. Increased rainfall after 1942 resulted in the re-vegetation of the floodplain and mid-channel islands. Vegetation stabilized the deposits by inhibiting erosion during floods and promoted floodplain construction and channel narrowing by trapping sediment. Likewise, the destruction of floodplain vegetation by large floods and droughts between 1914 and 1941 increased the susceptibility of the channel to widening (Schumm and Lichty 1963:79, 82).

Schumm and Lichty's study describes geomorphic responses to climatically-induced changes in flood magnitude and frequency during a relatively brief interval of ca. 40 years. Similar responses, occurring at time scales in excess of 1000 years, are documented by a study of the

Colorado River in central Texas (Baker and Pentead-Orellana 1977). Using empirical relationships of channel morphometry, discharge and sediment load developed for present day streams, Baker and P.-Orellana (1977) inferred changes over time in the paleohydrology of the Colorado River from variation in width, meander wavelength, sinuosity and sediment load of paleochannels preserved on geomorphic surfaces of known ages. Based on this evidence, they identified two complete cycles of landscape development: a Late Pleistocene cycle, extending from ca. 120,000 to 10,000 years B.P., and a Holocene cycle, extending from ca. 10,000 to 2,000 B.P. Consistent with Schumm's (1976) model, each cycle consisted of an episode of gradual aggradation, terminated abruptly by channel incision and erosion. In both cases, aggradational episodes began with the formation of a braided stream which transported coarse sediment. Over time, during each cycle, sediment load became progressively finer, and the braided stream developed into a sinuous (meandering) stream transporting silt and clay. Each cycle ended abruptly with an episode of erosion and stream entrenchment, during which a broad trough was scoured into the fine-grained deposits. Within this trough, a braided channel with a bed load of coarse sediment was created. This stream represented the beginning of a new aggradational episode, with deposition occurring within the newly-scoured fluvial trough (Baker and P.-Orellana 1977:415).

Comparing the Colorado River alluvial chronology to independently studied palynologic and paleontologic records from Texas, Baker and P.-Orellana (1977:415-418) found that braided, coarse-sediment streams

formed during relatively arid climatic episodes, while relatively moist conditions correlated with the formation of sinuous, fine-grained channels. They suggested that the braided channels formed in response to an increased seasonality of precipitation (i.e., precipitation occurring largely in spring and summer), which tended to concentrate discharges into infrequent, high magnitude flash floods. During moist episodes, Baker and P.-Orellana postulated that streamflow was more uniform, with frequent, moderate magnitude floods forming narrow, sinuous stream channels.

Both Schumm and Lichty (1963) and Baker and P.-Orellana (1977) associate changes in flood magnitude and frequency with changes from relatively moist to relatively dry climates. The climatic controls on flood characteristics involve not only the relative amounts of precipitation, but also the areal extent, duration and intensity of the storms that deliver precipitation. Wolman and Gerson (1978:191), citing meteorological studies, suggest that with increasing aridity, the areal extent of storms decreases, and that decreases in areal extent are accompanied by increases in rainfall intensity (i.e., the amount of rain falling in a given unit time). Thus, as noted by Leopold et al. (1951), rainfall of moderate intensity, falling over large areas, is less common in dry than moist climates.

The hydrologic response of fluvial systems to precipitation inputs differs from moist to dry climates. This difference appears in the shape of flood hydrographs, which plot discharge against elapsed time following a rainstorm. In response to moderate intensity, widespread rainfall,

hydrographs rise and fall gradually. Intense, localized storms, by contrast, generate hydrographs which rise and fall more rapidly. A flash flood is an example of such a response. Knox (1972:404) suggests that the peakedness of arid-climate hydrographs also reflects the decreased density of vegetation in arid climates. Vegetation intercepts falling rain and delays its removal to streams. Thus, the increased frequency of flash floods postulated by Baker and P.-Orellana (1977) for arid climatic regimes in central Texas reflects response to the intensity of precipitation as well as vegetation factors. Similarly, the moderate-magnitude floods of moist regimes occur in response to a moderation of rainfall intensity and an increase in vegetation cover.

With transition from moist to dry climates, Baker and P.-Orellana (1977) noted a change in sediment load from fine to coarse sediment. They postulate that devegetation associated with aridity, as well as the increased intensity of rainfall, increased sediment yields from the uplands. In addition, the large magnitude floods of dry climatic phases were able to transport coarser sediment than the more moderate floods of moist climatic phases.

Application

The alluvial chronology of the East Branch Walnut valley (Chapter 5) is consistent with Schumm's (1976) model of episodic geomorphic change (Fig. 37). Interpreted in terms of this model, three episodes of geomorphic activity are recognized, each characterized by different sets of geomorphic processes and hydrologic conditions. The three episodes are labelled Episodes A, B and C.

Episode A, the period of valley-wide aggradation which occurred between ca. 4500 and 2000 B.P., represents an interval of relative geomorphic equilibrium, manifested by the gradual aggradation of Snyder Soil (T-1) surfaces. This equilibrium adjustment was disrupted at or soon after 2000 B.P., since a radiocarbon date presumed to represent a basal date for the T-0 fill indicates that the Walnut River entrenched to its present level by ca. 1850 B.P. This episode of entrenchment, here labelled Episode B, corresponds to an erosional interval in Schumm's model (Fig. 37). By analogy with the models of Schumm and Lichty (1963) and Baker and P.-Orellana (1977), gradual aggradation during Episode A may represent an adjustment to frequent, moderate-magnitude floods.

Entrenchment, which occurred after 2000 B.P., represents an erosional interval in Schumm's (1976) model. During this interval, here termed Episode B, study area streams downcut to a lower level in the landscape, in the process eroding quantities of sediment. In the examples provided by Schumm and Lichty (1963) and Baker and P.-Orellana (1977), such episodes are associated with an increase in the frequency of large floods which widen and/or deepen stream channels to accommodate the increased discharge. It is postulated that, during Episode B, study area streams scoured broad, deep channels in which T-0 sediment was subsequently deposited. It must be emphasized that this inference is not supported by field evidence, since the stratigraphy of the T-0 fill was not investigated by the author. The inference is made on the assumption that the shift from gradual aggradation to entrenchment in

the study area conformed to Baker and P.-Orellana's (1977) model of geomorphic change on the Colorado River.

If a single date of 1855 ± 65 B.P. is correctly interpreted as an approximate date for the inception of T-0 aggradation, then Episode B was apparently quite brief. The brevity of Episode B is consistent with Schumm and Lichty's (1963) observation that major geomorphic change involving dramatic increases in channel dimension can occur very rapidly. In the case of the Cimarron River, such changes occurred in less than 30 years. Baker and P.-Orellana (1977) suggest that episodes of entrenchment on the Colorado River may have been induced by short-lived bursts of fluvial activity similar to those documented by Schumm and Lichty (1963).

In accordance with Schumm's (1976) model, the aggradation of the T-0 fill is tentatively interpreted as an episode of rapid deposition which ensued soon after entrenchment. The initiation of T-0 aggradation is here defined as marking the end of Episode B and the beginning of Episode C. As stated above, it is assumed that a broad trench was scoured during Episode B, and that this trench was partially filled by aggradation during Episode C. Again, the nature of both Episodes B and C are inferred largely by analogy to Schumm (1976) and are not based on field evidence.

Following Schumm and Lichty (1963) and Baker and P.-Orellana (1977), T-0 aggradation probably occurred in response to a decrease in the frequency of large floods, permitting the formation of a channel adjusted to lower magnitude discharges. For lack of evidence for further

subdivision, Episode C is here considered to be continuous from ca. 1850 B.P. to the present. According to Schumm's (1976) model, T-0 aggradation should have led to an episode of geomorphic stability. There is limited evidence both for and against such an episode. That some degree of stability has been maintained, at least in very recent times, is suggested by the fact that T-0 surfaces and streambanks in the study area are heavily vegetated with large trees, grass and shrubs. On the other hand, evidence for the deposition of levees at the edge of T-1 surfaces since 1000 B.P. indicates the occurrence of at least one interval within Episode C during which geomorphic work was performed on T-1, reflecting an increase in the effectiveness of large floods. Such floods probably had a destabilizing influence on T-0 deposits at lower levels of the landscape. It is thus likely that Episode C represents a complex sequence of geomorphic responses. Further insights into these responses are obtained by considering the specific hydrologic and climatic mechanisms which control the magnitude, frequency and geomorphic effectiveness of floods in the study area. These mechanisms are discussed in the next section.

HYDROLOGIC RESPONSE

Streamflow

At present, streamflow in the Kansas Flint Hills, including the drainage basin of the East Branch Walnut River, is relatively ephemeral, with a large proportion of annual streamflow occurring in a relatively few, moderate and high magnitude events. Metzler (1966:176) observed that "on a typical Flint Hills stream, average flows are equalled or

exceeded only about 15% of the time" and that "75 to 85% of the total annual flow occurs in about 15% of the time." Thus, while above-average discharges occur only during a total period equivalent to 55 days per year (i.e., 15% of 265 days), these flow events conduct over three-quarters of the total annual discharge. During the remainder of the year, streamflow is generally low. Complete cessation of streamflow is common during the summer months on many tributaries, and has occurred during severe droughts on most mainstem rivers (Kansas Water Resources Board 1960).

Most floods on Flint Hills streams are relatively flashy, with flood peaks rising and falling rapidly. The flashiness of these floods is due in part to geologic and physiographic factors. Flint Hills upland soils, derived from shale and limestone parent materials, are generally clayey and slowly permeable, or steep, stony and shallow (Bidwell 1966; Kollmorgen and Simonett 1965). The steep slopes and impermeable substrates promote rapid runoff, contributing to a rapid response of Flint Hills watersheds to precipitation (Metzler 1966:184).

The principal control of streamflow in the Flint Hills, however, is precipitation. In eastern Kansas, including the Flint Hills, precipitation is delivered by two types of storms. One type produces intense rains of short duration over a relatively small area, while the second type lasts for several days, delivering moderate intensity rainfall over thousands of square miles (K.W.R.B. 1961:36). Both types of storms, if of sufficient magnitude, produce floods which directly reflect characteristics of the rainfall input. The high

intensity, short-duration thunderstorms cause localized flash flooding, generally restricted to one or two tributaries. Areally extensive, long-duration storms, by contrast, generate basin-side floods on tributary and mainstem streams alike (K.W.R.B. 1960:6; Metzler 1966: 184).

The two types of storms, and by implication, the streamflow produced by each, vary in relative frequency during the year. Intense thunderstorms occur more frequently in late spring and summer, while moderate intensity, areally extensive storms are more frequent in early spring and fall (Lokke 1961:235). To explain this seasonal variation in precipitation type in eastern Kansas, it is necessary to consider climatic mechanisms operating on a continental scale, involving large-scale features of the upper atmospheric circulation. The most important feature of this circulation, as relevant to this study, are seasonal changes in the relative strength of the westerly, or zonal component of the upper atmospheric air flow. When this zonal component is strong, as it normally is in winter (Bryson 1966), a mild, dry Pacific air mass extends across the mid-continent. The southeastern boundary of this wedge of Pacific air is outlined in Fig. 38 by the mean winter position of the Pacific front. Fronts form at the convergence of unlike air masses, and thus the mean frontal position mapped in Fig. 38 shows a zone in which the Pacific air mass in winter commonly converges with Tropical air flowing from the Gulf of Mexico. In winter, the Pacific air mass effectively blocks most intrusions of moist air from the Gulf of Mexico and the Arctic Ocean,

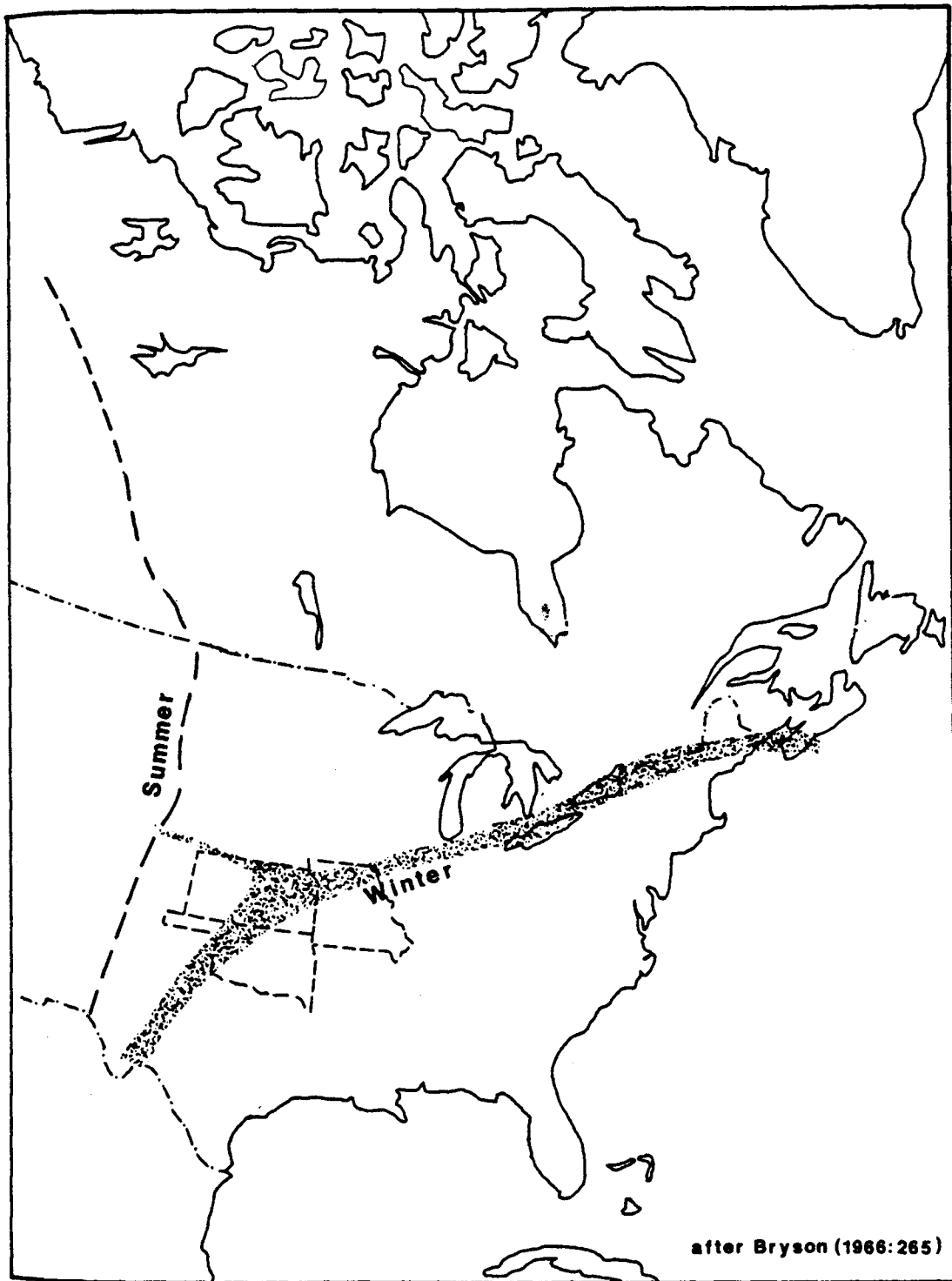


Fig. 38. Mean winter and summer positions of the Pacific front over North America (after Bryson 1966).

thus accounting for the characteristic winter droughts of the midwest (Borchert 1950; Bryson and Hare 1974:14).

In late spring and summer, the zonal component in the circulation weakens and is replaced by a strong meridional (north-south) component. The Pacific air mass is pinched out by southward extensions of cold Arctic air and by poleward surges of warm Tropical air (Bryson 1966:253). The convergence of these two air masses triggers the violent thunderstorms that characterize late spring and early summer in the mid-continent (Bryson and Hare 1974:16).

Early spring and fall are times of transition between the winter and late spring/summer circulations. As the zonal component strengthens in the fall, and as the meridional component strengthens in the spring, frontal storms frequently develop along the winter position of the Pacific front. Most of eastern Kansas is included in the zone of frequent frontal activity (Fig. 38), thus accounting for the increased occurrence of widespread, moderate intensity frontal storms in early spring and fall (Lokke 1961).

According to the Kansas Water Resources Board (1960:61), 70% of all floods exceeding bankful stages in the Walnut and Verdigris River basins in Kansas occur between April and July, when intense thunderstorm activity dominates Kansas rainfall patterns. The correlation of precipitation and floods in the East Branch Walnut valley is illustrated by the distribution per month of peak discharges exceeding 1000 cfs (cubic feet per second) at the U.S.G.S. stream gage on Cole Creek (Fig. 39). Between 1961 and 1980, 15 of 17 high magnitude floods

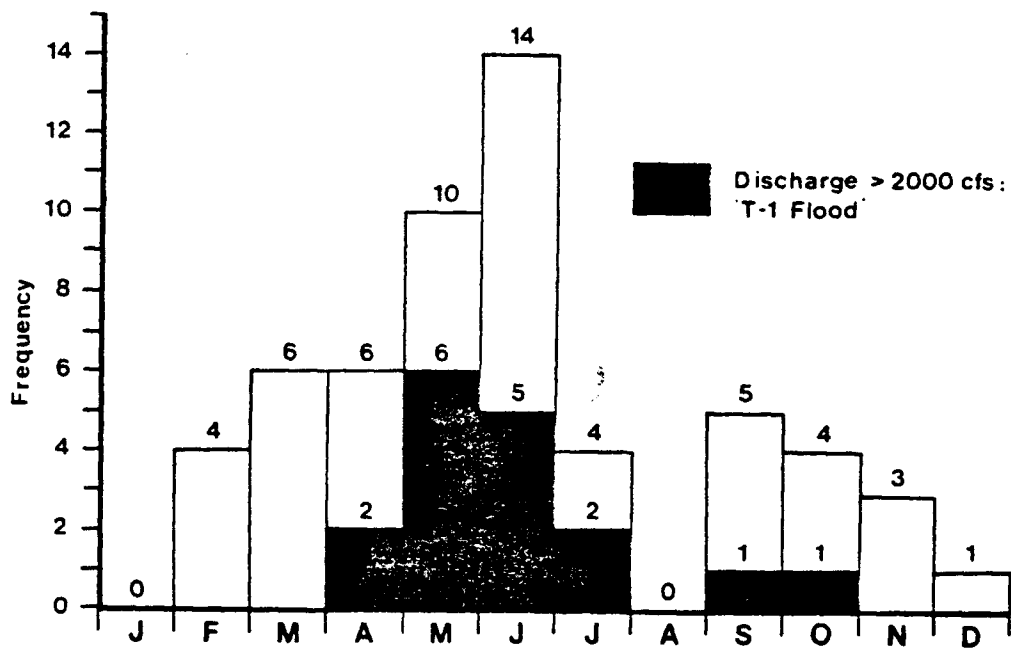


Fig. 39. Frequency of floods per month at Cole Creek stream gage (1961-1980).

(defined as floods which inundate T-1 surfaces at the Cole Creek gage; see Figs. 32, 33) occurred in April through July, during the late spring/summer period of intense thunderstorm activity. Low to moderate magnitude floods occur in all months of the year but August and January. The absence of floods in the latter months is probably due to the shortness of the record at the Cole Creek gage. Significantly, low to moderate magnitude floods are proportionately more frequent than large floods in spring and fall (Fig. 39), when rainfall in eastern Kansas is produced largely by moderate intensity frontal storms.

The flashiness of streamflow in the Flint Hills is thus largely attributed to the fact that a large proportion of the annual rainfall is delivered by intense thunderstorms which are generated by meridional clashes of Polar and Tropical air in late spring and summer. The flashy response of watersheds to rainfall events is heightened by the steep relief and shallow or impermeable soils of the Flint Hills upland.

Sediment Yield

Although a detailed study of sediment yield in the study area is not undertaken in this thesis, a consideration of edaphic and vegetational parameters suggests that sediment yields from upland surfaces in the Flint Hills are relatively low. Since, as mentioned previously, upland soils in the Flint Hills are either clayey or stony, they are probably relatively resistant to erosion. Resistance to erosion is

further enhanced by the tall-grass prairie vegetation cover of the uplands. The root system of prairie grasses create a cohesive sod which holds soil and inhibits erosion. In addition, prairie grasses also break the impact of rainfall and reduce to some extent the velocity of surface runoff (Tricart and Callieux 1973:79).

In many regions of North America, drastic increases in sediment yield have occurred in historic times as a result of clearance and cultivation of uplands (cf. Knox 1977; Johnson et al. 1980). The Flint Hills upland, valued by Euroamerican agriculturalists primarily as rangeland for livestock grazing, has remained largely uncultivated. As of 1973, 80% or more of the uplands drained by the East Branch Walnut River were uncultivated range (Williams and Barker 1973). Although over-grazing and the once-popular range management practice of annual burning have reduced the vegetation cover from its native condition, thereby increasing sediment production (Anderson 1967; Kollmorgen and Simonett 1965:287-288), historic increases in sediment yield in the Flint Hills are undoubtedly far below those experienced elsewhere in North America.

Because of low upland sediment yields, Flint Hills streams probably receive and transport relatively low quantities of sediment. In addition, the heavily vegetated streambanks and the cohesiveness of the fine-grained alluvial fills undoubtedly inhibit erosion of sediment. The low concentration of sediment in Flint Hills streams is supported by Cross (1954:303), who noted the presence in Flint Hills streams of several species of fish with "Ozarkian distributions", i.e.,

fish requiring clear, non-sluggish habitats more commonly associated with the Ozark Highlands than the Plains. Cross and Braasch (1968), however, note the recent decline of such fish as a result of pollution from cattle feedlots.

Geomorphic Effectiveness of Floods

The alluvial landscape of the East Branch Walnut appears to reflect an equilibrium adjustment to the relatively ephemeral streamflow of the present day. It is suggested here that the deeply entrenched stream channels are formed and maintained by the flashy, moderate magnitude floods, which, according to Metzler (1966:176), conduct a large percentage of the total annual discharge. The low discharges which characterize streamflow during the remainder of the year lack the power to alter the established channel pattern, and, furthermore, permit the establishment of dense vegetation on streambanks and floodplain surfaces. This vegetation, in turn, inhibits erosion by larger magnitude floods.

Despite the depth of entrenchment of study area streams, T-1 surfaces flood at relatively frequent intervals of 1.2 to 7 years (Chapter 5). These large floods, however, perform little geomorphic work on T-1 surfaces, for the following reasons. First, because of low upland sediment yields and the low erodibility of streambank and floodplain sediments, floods exceeding T-1 scarps probably transport little sediment. Second, and perhaps most importantly, the flashiness which characterizes most Flint Hills floods implies that flood peaks rise and fall rapidly, and that floods inundate T-1 surfaces for only brief periods of time (cf. Metzler 1966:184). In addition, by abandoning the sinuous low-flow

channel for a more direct, down-valley course, "T-1 floods" probably maintain relatively high velocities. Both the brevity of inundation and the velocity of flood waves reduce the probability that sediment carried by the floods will be deposited (Wolman and Leopold 1957:101-103). The velocity maintained by the floods implies that the net geomorphic effect should tend toward erosion (Metzler 1966). Evidence of erosion was noted at 14BU25 (see pedon description in Chapter 3). At this site, the plowzone was apparently stripped from portions of the site by the large flood of 1979 (Root 1981b).

The numerous abandoned stream channels noted on T-1 surfaces in the study area may serve an integral function in the landscape's adjustment to relatively frequent, large floods by acting as drainage channels for overbank flows. The presence of such channels undoubtedly further increases the swift down-valley passage of flood waves (Wolman and Leopold 1957:103).

Past Responses

The preceding section discussed climatic and vegetational parameters which at present influence streamflow and sediment yield in the study area. Assuming that these same relationships have held in the study area during the past 4500 years, then the geomorphic changes recorded in the East Branch Walnut alluvial chronology may be explained in terms of variations in the magnitude and frequency of climatic and hydrologic processes observed in the area today.

Episode A

As suggested by the model of episodic geomorphic response discussed in the first section of this chapter, Episode A (ca. 4500-2000 B.P.) probably experienced a streamflow regime dominated by low to moderate magnitude floods. Large floods were probably less frequent than at present. The Snyder Soil surface, which today occupies T-1, was the floodplain (T-0) of Episode A streams. Because study area streams were not entrenched during Episode A, the floodplain was probably subject to inundation by moderate magnitude floods. These floods deposited fine-grained sediment on Snyder Soil surfaces, suggesting that, relative to the present, the velocity of overbank flow was lower, and flood hydrographs were less peaked (less flashy). As a consequence of these two factors, floodwaters probably inundated Snyder Soil surfaces for longer periods of time, permitting the deposition of suspended sediment.

Aggradation of the Snyder Soil surface during Episode A was slow. An average aggradation rate of less than a centimeter a year was estimated in Chapter 4. This perhaps suggests that the concentration of sediment in Episode A streams was not significantly higher than at present; this, in turn, may indicate that vegetation in the valley and in the uplands was relatively dense.

In the southern Flint Hills, as discussed previously, moderate magnitude floods are proportionately more frequent (relative to large floods) during spring and fall, when rainfall occurs largely in the form of widespread frontal storms of moderate intensity. In Episode A, it is suggested that such storms occurred more frequently than at present; similarly, it is postulated that the meridional clashes of Polar and Tropical

Fig. 40. Rainfall, flood frequency and flood magnitude on Cole Creek (1962-1979).

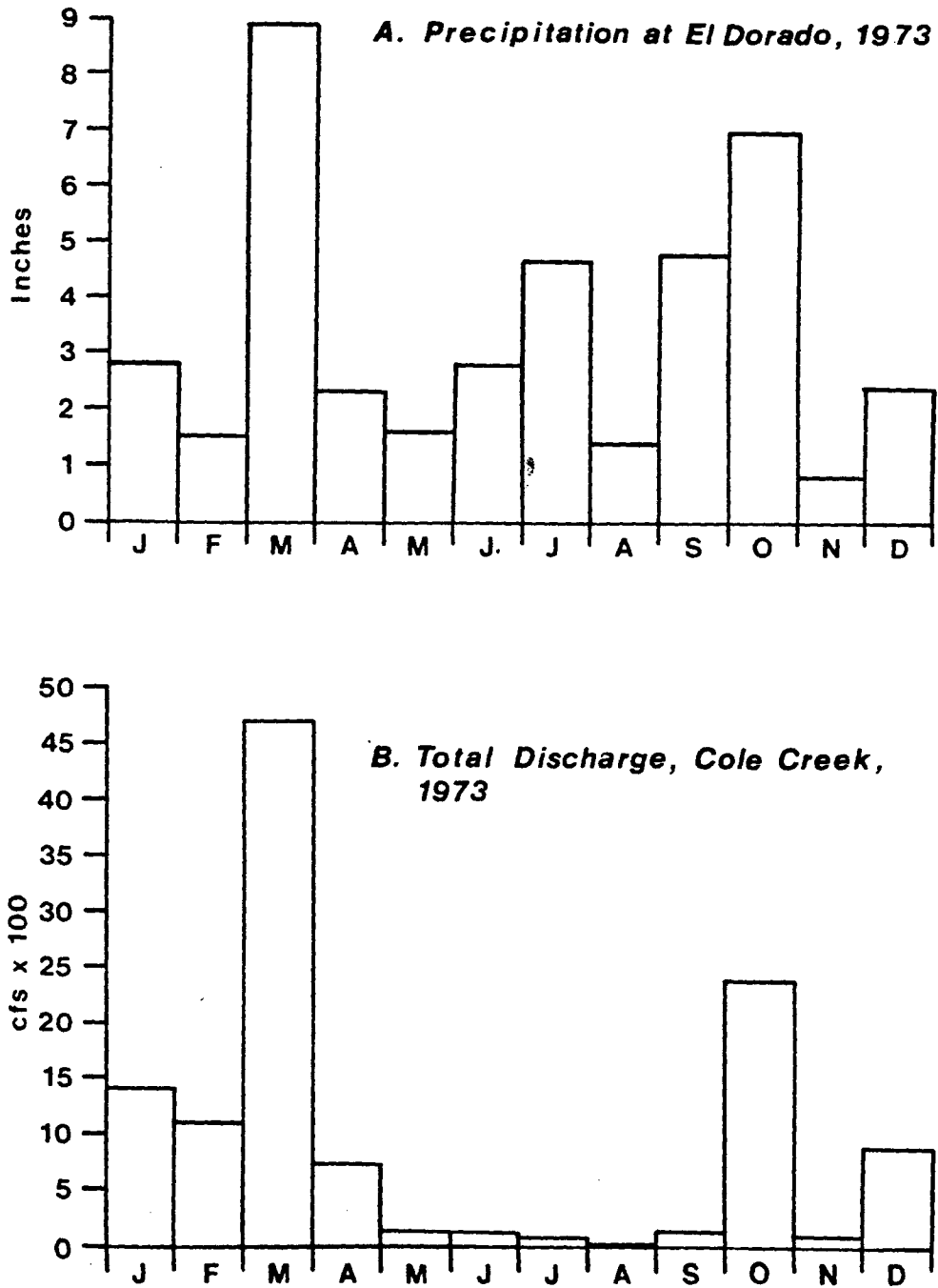


Fig. 41. Monthly precipitation at El Dorado compared to monthly total discharge at Cole Creek gage in 1973.

air which at present produce intense rains and large floods were less frequent during Episode A.

Schumm and Lichty (1963) and Baker and P.-Orellana (1977) found that geomorphic episodes dominated by moderate-magnitude floods tended to correlate with relatively moist climatic episodes. Fig. 40 compares year-to-year variation in annual precipitation and flood characteristics in the study area from 1962-1979. Of interest to the present discussion is the year 1973. This year was relatively wet, with the third highest rainfall in the 18 year series (Fig. 40a). Responding to this rainfall, Cole Creek experienced a relatively large number of floods (Fig. 40b). However, as shown in Fig. 40c, the magnitude of these floods was low, since the maximum annual flood (that is, the largest peak discharge observed during the year) was less than 2000 cfs.

In 1973, most of the precipitation fell in March, September and October (Fig. 41), times of frequent, moderate intensity rainfall. The summer months, which are normally rainiest, were quite dry, with the exception of July. Discharge totals at the Cole Creek stream gage reflect precipitation inputs, with high streamflow occurring in spring and fall, with dwindling discharges through the summer.

This example demonstrates that a condition of frequent, moderate magnitude floods, as postulated for Episode A, can be induced by atmospheric circulation patterns which favor the formation of moderate-intensity frontal storms over the southern Flint Hills. Although increased precipitation is one possible consequence of such a climatic pattern, as suggested by 1973, it is not necessarily the only possibility.

Such an interpretation, however, is consistent with the models proposed by Schumm and Lichty (1963) and Baker and P.-Orellana (1977), and is thus tentatively accepted. Further evidence supporting this inference is presented in the next section.

Episode B

During Episode B (ca. 2000 to ca. 1850 B.P.), study area streams entrenched. The previously cited models of geomorphic response suggest that entrenchment occurred in response to an increased frequency of large floods. Also at 2000 B.P., the valley-wide aggradation of Snyder Soil surfaces ceased. This suggests that the large floods of Episode B were characterized by flashy hydrographs with narrow bases and high peaks. Such floods, at present, inhibit deposition on T-1 surfaces.

It is postulated that the increased frequency of large, flashy floods was prompted by a sudden shift to an atmospheric circulation pattern favoring intense, meridional clashes of Polar and Tropical air masses. Such a pattern, at present, accounts for the high frequency of large, flashy floods in the southern Flint Hills.

Schumm and Lichty (1963) and Baker and P.-Orellana (1977) suggest that increases in the frequency of large floods correlate with increases in relative aridity, or with a series of rapid fluctuations between very wet and very dry years. The levees and colluvial footslopes at 14BU4, 14BU9, 14BU25 and 14BU87 possibly provide evidence for the occurrence of severe drought during Episode B. As established from pedologic considerations in Chapter 4, these landforms were deposited prior to 1000 B.P., and most likely prior to 1500 B.P. Both levees and colluvial

footslopes may be interpreted as evidence for increases in sediment yield. In the case of the colluvial footslope at 14BU25, the sediment was derived from adjacent upland surfaces. Levees, such as those observed at 14BU4, 14BU9 and 14BU87, were formed by overbank flow. Levees may be constructed as a result of increased sediment loads or an increased frequency of overbank flow. Wolman and Leopold (1957:99) observed that accelerated erosion and runoff from unpaved streets and construction sites near Washington D.C. led to the formation of levees on streams which had previously lacked levees. Thus, increases in upland sediment yield may have increased the sediment load of study area streams. Increased sediment yield may reflect a decrease in the density of vegetation in response to drought during Episode B. The increased intensity of precipitation inputs in Episode B may also have accelerated the erosion of relatively devegetated upland surfaces.

The vegetational and hydrologic responses discussed above need not reflect a continuous drought. Rather, as suggested by Schumm and Lichty (1963), Episode B may have experienced a rapid fluctuation between wet and dry years. Vegetation, in general, dies back in response to drought faster than it recovers from drought. Thus, severe, short-term droughts during Episode B may have reduced vegetation cover, increasing the potential of hillslope erosion with the return of rains in subsequent seasons (Baker and P.-Orellana 1977; Knox 1972).

Episode C

Episode C, extending from ca. 1850 B.P. to the present, reflects the stabilization of the East Branch Walnut valley following entrenchment.

The deposition of the T-0 fill within the Episode B trench occurred in response to moderate magnitude floods, possibly in the presence of an increasingly stable riparian vegetation. This vegetation, by stabilizing streambanks, may have decreased the geomorphic effectiveness of large floods. It is postulated that Episode C marks the development and maintenance of a geomorphic adjustment to relatively ephemeral stream-flow which is continuous to the present.

This adjustment, however, was disturbed at least once during Episode C. Levees constructed on T-1 surfaces during Episode C at 14BU31, 14BU55 and 14BU57 may represent relatively short-term increases in the frequency of large floods and in sediment load. These episodes may represent brief intervals of drought. It is likely that these minor episodes of disequilibrium produced changes in the stream channel and possibly in T-0 deposits, although no evidence of such changes is available.

Summary

The geomorphic events represented in the East Branch Walnut alluvial chronology have been accounted for in the preceding sections in terms of climatically-induced hydrologic change. The events, and the hydrologic and climatic conditions inferred from them, are summarized in Table 17. Major geomorphic changes in the study area were induced by hydrologic changes involving shifts in flood magnitude and frequency. Shifts in flood magnitude and frequency, in turn, were induced by climatic changes involving the areal extent, duration and intensity of storms delivering precipitation. Climatic change is also reflected in increases in sediment yield which occurred in response to decreases in vegetation

Table 17. Geomorphic, hydrologic and climate changes inferred from the East Branch Walnut alluvial chronology.

Episode	Time Interval (yrs. B.P.)	Geomorphic Response	Hydrologic Conditions	Climatic Conditions
A	4500-2000	Slow, valley-wide aggradation of T-1 surfaces.	Frequent, low to moderate magnitude floods. Infrequent large floods. Broad-based flood hydrographs.	Increase in relative frequency of frontal storms, bringing widespread, moderate intensity rainfall.
B	2000-ca. 1850	Entrenchment	Sudden increase in frequency of large, flash floods, prompting channel widening and deepening.	Increase in frequency of intense, localized thunderstorms.
C	post-1850	T-0 aggradation, persistent entrenchment.	Decrease in frequency of large floods. Channel narrowing enhanced by establishment of riparian vegetation.	Similar to present. Relatively ephemeral streamflow, frequent flash floods.
B and C	episodic	Deposition of levees, colluvial footslopes	Increase in sediment yield, probably reflects devegetation of upland and valley surfaces.	Short-term drought, probably accompanied by increase in frequency of large floods.

cover during short-term droughts.

As shown in Table 17, streamflow in the East Branch Walnut valley was dominated by frequent, moderate magnitude floods during Episode A (4500-2000 B.P.). The lower overbank velocities and less peaked hydrographs of these floods accounts for gradual valley-wide aggradation during Episode A. The pattern of frequent, moderate magnitude floods reflects an increase in the relative frequency of widespread, moderate intensity frontal storms.

During Episode B (2000-1850 B.P.), the frequency of large floods increased, and study area streams entrenched, thereby increasing the dimensions of their channels. The flashiness of floods also increased, inhibiting the deposition of sediment on Snyder Soil (T-1) surfaces. Colluvial footslopes and levees, constructed on T-1 surfaces during Episode B, indicate increases in sediment yield, possibly in response to relative short-term droughts which decreased upland vegetation cover and increased the susceptibility of hillslopes to erosion. The increased frequency of large, flashy floods may indicate an increased meridionality in atmospheric circulation, a condition which at present causes intense thunderstorms in eastern Kansas, including the study area.

Episode C (1850 B.P. to the present) represents an adjustment to an ephemeral streamflow regime dominated by frequent, large magnitude floods. An important part of this adjustment is the heavy vegetation cover of streambanks and floodplains, which inhibits erosional damage and permits the maintenance of relatively narrow, although deep, channels. The equilibrium adjustment of Episode C was interrupted at least once

by short-term increases in sediment load, during which levees were constructed on T-1 surfaces. These landforms provide evidence of brief episodes of geomorphic instability, perhaps induced by severe, short-term droughts.

REGIONAL SYNCHRONEITY OF GEOMORPHIC RESPONSES

The previous sections suggest that the most important factor inducing the shift from valley-wide aggradation to entrenchment at ca. 2000 B.P. in the East Branch Walnut valley was an increase in the frequency of intense thunderstorms, relative to more moderate intensity frontal storms. Obviously a change in precipitation patterns of this nature would have occurred on a regional, rather than local, scale, since both types of storms are produced by the movement of air masses on a continental scale. Therefore, if climatic change is actually reflected in the East Branch Walnut alluvial chronology, it is reasonable to expect that evidence of this change should also be reflected in the alluvial chronologies of other nearby streams.

This section presents evidence for the synchronous occurrence of geomorphic change at 15 localities in western Missouri, Kansas and northeastern and southwestern Oklahoma. This three-state region is cross-cut by the southeastern periphery of the Central North American Grassland, as this biome is defined by Borchert (1950). The edge of the Grasslands is more accurately described in terms of climatic and vegetational gradients than discrete geographic boundaries. Vegetation changes from southeast to northwest across the region from deciduous forest to prairie/forest mosaic to prairie (Fig. 42). There is a

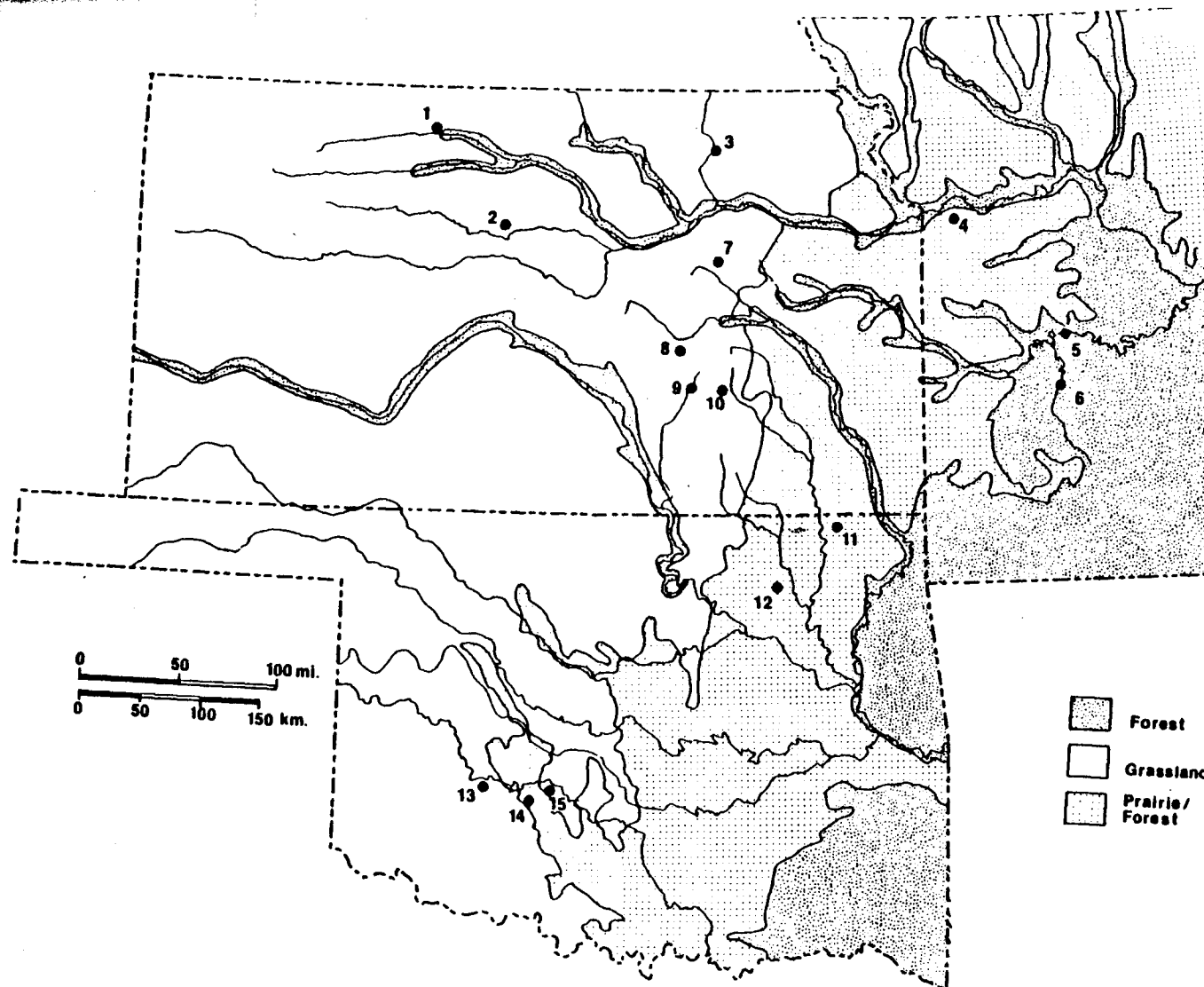


Fig. 42. The southeastern edge of the Grasslands in Kansas, Oklahoma and Missouri (after Kuchler 1964). Numbers refer to localities listed in Table 18.

corresponding southeast-northwest decrease in average annual precipitation.

The southeastern edge of the boundary between prairie-forest mosaic and deciduous forest (Kuchler 1964) presently lies to the south and east of the mean winter position of the Pacific front, as mapped by Bryson (1966:235; compare also Fig. 38 and 42). The frontal zone, as discussed previously, marks the area within which Tropical and Pacific air masses frequently converge in winter. The importance of this climatic boundary is not confined to winter, however, since outbreaks of Pacific air in summer tend to follow the general form of the circulation pattern of winter (Bryson 1966). The frontal zone establishes southeast to northwest gradients in precipitation which importantly influence both vegetation and streamflow.

The position of the Pacific front is largely responsible for the increasing seasonality of precipitation and runoff from southeast to northwest across the southeastern periphery of the Grasslands. Because the Pacific front blocks intrusions of moist Tropical air in winter, there is a southeast-to-northwest increase in the percentage of total annual precipitation which falls in summer, and a decrease in the percentage which falls in winter. For example, in southeastern Oklahoma, 22-24% of the total precipitation falls in June through August, while 18-20% falls in December through January. In the panhandle of western Oklahoma, by contrast, 42-48% of the total precipitation falls in summer, and only 6-10% in winter (Sutherland 1977:50-51). In eastern Kansas, 65-70% of the annual precipitation falls during April to September (the

growing season); in western Kansas, 75-80% of the annual total falls during these months (Bark 1961:8). Thus, precipitation is more uniformly distributed through the year in the southeastern portions of Kansas and Oklahoma, where the flow of moist air from the Gulf is less frequently blocked during the year by Pacific air. As a consequence of the increasingly uniform seasonal distribution of precipitation, streamflow becomes less ephemeral from northwest to southeast. More uniform runoff is also prompted by denser vegetation and less intensive moisture stress in the more humid southeast.

The intensity of summer rainfall, measured as the amount of rain delivered by single storms, increases in a general east-west direction. Borchert (1950:7) noted that, from east to west across most Plains states, "the probability of an inch or more of rain falling from one storm doubles." Increasing rainfall intensity undoubtedly accounts for a southeast-northwest increase in the flashiness of floods across Kansas and Oklahoma (Beard 1975).

Finally, the frequency of, and probability of, short-term drought increases from southeast to northwest across Kansas and Oklahoma (Borchert 1950), due largely to air mass characteristics. Drought in the Grasslands is generally attributed to strong summer westerlies, which prevent intrusions of moist maritime air into the continental interior (Borchert 1950). In other words, drought occurs when the circulation pattern of winter persists into the summer months. Borchert (1950) and Kuchler (1967) observe that the grassland vegetation is adapted to the frequent recurrence of moisture stress. The capacity of grassland vegetation to

rapidly respond to and recover from severe climatic stress was documented during and after the droughts of the 1930's (cf. Weaver and Albertson 1956).

The southeast-northwest gradients in the seasonality, intensity and variability of precipitation emphasize a point noted frequently by botanists and climatologists of the Plains: the unique character of the region's climate is not reflected in averages, but in the seasonal and year-to-year variance of conditions around those averages.

Analysis

The 15 localities considered in this analysis are located as shown in Fig. 42; Table 18 provides further information on each locality. This analysis uses only geomorphic events which were approximately contemporaneous with the 4500-year alluvial chronology of the East Branch Walnut valley; only data from alluvial landscapes are used. Data concerning rockshelter, eolian and hillslope deposition are not included. The resultant data set represents a comprehensive sample of geomorphically-relevant studies undertaken in the three-state region during the relevant time range. The list is not complete, since several studies are still in progress.

Data from the 15 localities are in many ways not directly comparable among localities, since the studies which report the data have varying purposes and scopes of analysis. The kinds of geomorphic data reported for each locality reflect the purpose of individual researchers. Schmits (1980), Brakenridge (1981) and Kopsick (1981), for instance, focus largely on geomorphic processes of erosion and deposition and make little

Table 18. Localities providing geomorphic data from ca. 5000 B.P. to the present on the southeastern periphery of the Central Plains.

Locality ^a	Relationship in Major Drainage System	Scale of Study ^b	Intent of Study ^c	Source
1. Deer Creek	Kansas, via Solomon	V	SG	Johnson 1981
2. Wolf Creek	Kansas, via Saline	V	SG	Johnson 1981
3. Big Blue River at 14P01	Kansas	S	G	Schmits 1980
4. Little Blue River	Missouri	B	G	Kopsick 1981
5. Osage River at 23BE1010	Missouri	S	SG	Lees et al. 1981
6. Pomme de Terre River	Missouri, via Osage	B	G	Brakenridge 1981
7. Munkers Creek at 14M0304	Arkansas via Neosho	S	A	Witty 1982
8. Cedar Creek at 14CS355	Arkansas via Cottonwood, Neosho	S	A	Wood 1977
9. East Branch Walnut River	Arkansas	B	SG	This study
10. Fall River at 14GR38	Arkansas, via Verdigris	S	A	Johnson 1971
11. Cotton Creek	Arkansas, via Little Caney, Caney, Verdigris	V	SG	Artz 1982
12. Candy Creek	Arkansas, via Bird, Verdigris	V	A	Saunders 1980
13. Carnegie Canyon	Red, via Washita	V	SG	Lintz and Hall 1983
14. Delaware Canyon	Red, via Washita	V	SG	Hall 1980a, 1982c
15. Central Washita River	Red	V	G	Goss et al. 1972

^a Number is keyed to location in Fig. 42.

^b B - sequence for drainage basin; V - sequence for one stream valley or valley segment;

S - site-specific sequence.

^c G - emphasis on geomorphology; SG - emphasis on soils and geomorphology; A - emphasis on archaeology.

reference to soils. Soil geomorphic relationships, in other studies, have been a principal focus of research. Other studies have a primarily archaeological focus: the geomorphic implications of the latter studies are frequently embedded in discussions of cultural stratigraphy and have been carefully extracted by this writer. The research focus of each study is listed in Table 18.

The studies also vary in the geographic scope of analysis, as shown in Table 18. Three studies consider geomorphic changes which are generalized from an entire drainage basin, including tributaries confluent to a mainstem stream. Six studies document geomorphic changes occurring primarily in a single tributary or a specified reach of a larger stream valley. Seven studies document geomorphic change at single sites, but do not discuss, in detail, the site's geomorphic context within a valley.

For purposes of the present analysis, which searches for general patterns in the geomorphic data, the most useful data are from studies of basins and valleys. These studies present sequences which are generalized from single-site and single-event data, and thus include a range of dates and geomorphic contexts. The problem with single-site data is that the specific sequence encountered at a site may not be typical of the general trend of events in a valley or basin. In the East Branch Walnut valley, for example, sites varied considerably in the kinds of geomorphic events documented at each. Only by considering a number of sites was the general trend of the alluvial chronology established. In the present analysis, site-specific data are used primarily to support patterns suggested by the more generalized sequences.

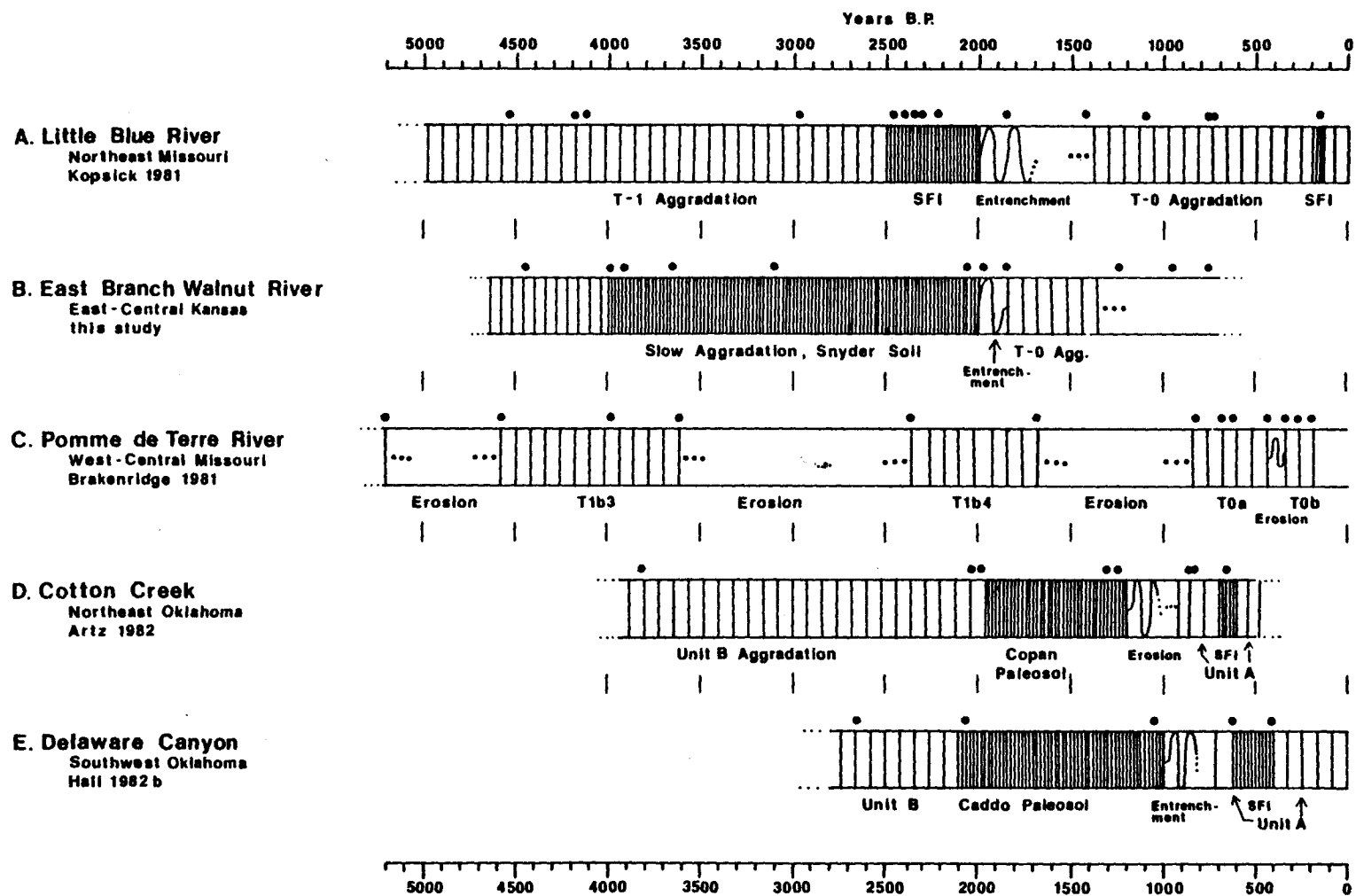


Fig. 43. Five alluvial chronologies from the southeastern periphery of the Grasslands.

Fig. 43 illustrates alluvial chronologies from three drainage basins and two valleys located on the southeastern periphery of the Grasslands. Of the studies listed in Table 18, these provide the most complete and best-dated generalized chronologies. Episodes of aggradation, soil formation and erosion (including stream entrenchment) are represented in this diagram. The temporal limits of the episodes shown in the diagram are those specified by the person reporting each chronology, with exceptions as noted below. Ellipses in the diagram represent cases in which the beginning or end of an episode was either not stated or not known by the author. All five chronologies are supported by radiocarbon evidence; these dates are shown on the diagram by large dots. Specific information on the Little Blue, Pomme de Terre, Cotton Creek and Delaware Canyon chronologies are provided below.

Little Blue River. In the Little Blue River drainage of northwest Missouri (Fig. 42, loc. 4), the modern floodplain (T-0) is typically 2 m above the present low-water channel; T-1 is typically 4-5 m above the channel (Kopsick 1981:56). Radiocarbon dates show that the upper 2.9 m of the T-1 fill aggraded between ca. 4500 and 2000 B.P. Aggradation either slowed or ceased between ca. 2500 and 2000 B.P. (Kopsick 1981:62), as inferred from the large number of unburied Woodland and post-Woodland sites found on present-day T-1 surfaces (Kopsick 1981:67). Fig. 43a shows an interval of soil formation from 2500-2000 B.P., associated with the stabilization of T-1 surfaces. Kopsick (1981) does not describe such an interval. However, given evidence that radiocarbon dated archaeological sites within 20 cm of T-1 surfaces range in age

from 2400 ± 85 to 1850 ± 140 B.P., it is considered likely that T-1 surfaces were sufficiently stable at this time to permit soil formation.

After 2000 B.P., but before 1420-1460 B.P., the Little Blue incised and the present floodplain (T-0) began to aggrade. The upper one meter of the T-0 fill at 23JA43 aggraded between ca. 750 and 150 B.P. The date of 150 B.P. is from a soil which is believed to have formed before Euro-American settlement in the midnineteenth century. Alluvium burying this soil reflects increased runoff and sediment yield due to land clearance and cultivation (Kopsick 1981:69).

Pomme de Terre River. The Pomme de Terre River, a tributary of the Osage River, is located in west-central Missouri (Fig. 42, loc. 6). The alluvial chronology for this river (Brakenridge 1981) contains both Pleistocene and Holocene events. The Holocene portion of the chronology is manifested in T-0 and T-1 terrace fills. T-1, formed between ca. 38,000 and 1680 B.P., is comprised of five lithologically and time-stratigraphically distinct units. T-0, formed between ca. 840 and 140 B.P., consists of two such units. Brakenridge (1981) presents the Pomme de Terre alluvial chronology as a sequence of alternating episodes of aggradation, stability and erosion (entrenchment). Radiocarbon dates from each terrace fill are used to suggest the minimum interval of time during which the fill was deposited. The interval of time separating the oldest date of one fill from the youngest date of the succeeding fill is considered to bracket an erosional episode during which the stream entrenched, abandoning the previous floodplain level. In the sequence of dates used by Brakenridge (1981), four aggradational episodes are

represented since ca. 5000 B.P.: 4585-3610 B.P. (T1b3); 2360-1680 B.P. (T1b4); 840-430 B.P. (T0a) and 330-140 B.P. (T0b). Erosion intervals, separating the four aggradational episodes, occurred at sometime between 5200 and 4585 B.P., 3610 and 2360 B.P., 1680 and 840 B.P. and 430 and 330 B.P. These intervals do not state the absolute limits of aggradational and erosional episodes on the Pomme de Terre. For example, erosion did not necessarily occur throughout the interval 1680-840 B.P., but may have occurred only during one or two centuries, with aggradation or stability characterizing the remainder of the interval. Although Brakenridge (1980:655; 1981:72), in illustrations of this alluvial chronology, implies that entrenchment did indeed occur in brief intervals, he presents no evidence for such intervals, and they are thus not included in Fig. 43.

Cotton Creek. The alluvial stratigraphy of Cotton Creek, a tributary of the Little Caney River in northeastern Oklahoma (Fig. 42, loc. 11) includes two lithostratigraphic units (Units A and B) and a soil-stratigraphic unit, designated the Copan Paleosol (Hall 1977). The Copan Paleosol is developed in the upper 40-100 cm of Unit B. Unit A occurs as floodbasin or levee deposits burying the Copan Paleosol, or as the fill of abandoned channels incised into Unit B (Artz 1982). This sequence, although recognizable in the valleys of both the Little Caney and Cotton Creek (Hall 1977), has been most intensively studied and dated on the latter stream (Artz 1982).

Radiocarbon dates from depths of 1-3 m in the B or C horizons of the Copan Paleosol suggest that Unit B was aggrading without sufficient

stability for pedogenesis between 3810 and 1981 B.P. The initiation of pedogenesis in Unit B, forming the Copan Paleosol, probably occurred as aggradation slowed after ca. 2000 B.P. Radiocarbon dates from the upper 20-30 cm of Copan Paleosol A horizons suggests that soil formation, accompanied by slow aggradation, continued until at least ca. 1250-1300 B.P. Unit B aggradation ended between 1300 and 830-850 B.P., at which time Unit A was being deposited on top of Unit B and in channels incised into the Unit B surface. The 400 year gap between Units A and B probably witnessed erosion in the form of the entrenchment and/or lateral movement of stream channels. Brief intervals of soil formation, represented by thin, weakly-developed A horizons, interrupted Unit A aggradation at some, but not all localities in the valley (Artz 1982).

Delaware Canyon. The west fork of Delaware Creek, a tributary of the Washita River in southwestern Oklahoma (Fig. 42, loc. 14) flows in a deeply eroded, canyon-like valley, referred to locally as Delaware Canyon. Two alluvial stratigraphic units (Units A and B) are described by Hall (1982c). The Caddo Paleosol, referred to as the Kickai Paleosol in preliminary reports (Hall 1980a), is a weakly-developed cumulative soil developed in the top of Unit B. A date of ca. 2650 B.P. at a depth of 3 m beneath the Caddo Paleosol suggests rapid aggradation at this time; nine dates from the A horizon of the Caddo Paleosol indicate that it formed, in conjunction with slow aggradation of Unit B surfaces, between ca. 2065 and 1050 B.P. At ca. 1000 B.P., the stream incised into the Unit B surface. Unit A deposition subsequently mantled the Caddo Paleosol as well as filling channels eroded into Unit B during

incision. A weakly-developed paleosol is observed in Unit A, ca. 1.5 m above the Caddo Paleosol and 50-100 cm beneath the modern surface (Hall 1982c).

The five chronologies summarized in Fig. 43 each contain at least one major episode of stream entrenchment, which ended a preceding episode of aggradation or soil formation. On the Little Blue and East Branch Walnut Rivers (Fig. 43a,b), the transition from T-1 aggradation or stability to entrenchment occurred at ca. 2000 B.P. On the lower Pomme de Terre River (Fig. 43c), T-1 was abandoned sometime between 1680 and 840 B.P. On Cotton Creek and Delaware Canyon, a major geomorphic transition involving stream entrenchment or lateral movement occurred at ca. 1200 and ca. 1000 B.P., respectively (Fig. 43d,e). In four of the five chronologies, an episode of soil formation immediately preceded the shift. Brakenridge (1981) does not discuss soil-geomorphic relationships; thus it is not known if a soil formed beneath T1b4 surfaces prior to entrenchment. Despite this uncertainty, the direction of the shift on the Pomme de Terre, (from aggradation to entrenchment) is the same as in the other four chronologies. Thus, Fig. 43 shows that within an interval of 1000 years, streams experienced approximately similar changes over a relatively large geographic area.

In addition, the figure hints that these changes were time-transgressive. In the Little Blue and East Branch Walnut valleys, the transition occurred at ca. 2000 B.P. To the south, in northeast and southwest Oklahoma, a similar shift occurred on Cotton and Delaware Creeks, but not until 700-1000 years later. On the Pomme de Terre, in west-central Missouri, the shift from aggradation to entrenchment also

occurred after similar shifts on drainages to the west and north. If stream entrenchment, as suggested previously for the East Branch Walnut valley, was due to a climatic change over the southeastern periphery of the plains, then Fig. 43 indicates the possibility that the change was time-transgressive from northwest to southeast, occurring earlier in central Kansas and northeastern Missouri than in west-central Missouri, southwest Oklahoma or northeastern Oklahoma.

A number of studies from Kansas, Missouri and Oklahoma document geomorphic events similar to, and synchronous with those documented in Fig. 43. Detailed discussion of this evidence, obtained from studies listed in Table 18, is not presented, but an attempt is made to identify evidence for significant geomorphic discontinuities which are approximately synchronous with the shifts from aggradation or soil formation to entrenchment shown at 1000-2000 B.P. in Fig. 43. Table 19 lists dates associated with either soil formation or the end of an aggradational episode from 15 basins, valleys and sites. Supporting evidence for the discontinuities shown in Table 19 are summarized briefly below.

On Deer Creek and Wolf Creek (Fig. 42, locs. 1-2), radiocarbon dates from soils indicate that soil formation occurred between ca. 2000 and 2500 B.P. (Johnson 1981).

At 14P01 (the Coffey site) on the Big Blue River (Fig. 42, loc 3), a soil formed at the top of a channel fill deposit (Unit IV). Dates of 2360 ± 60 and 2480 ± 55 B.P. were recovered from archaeological strata ca. 1.0 and 2.0 m, respectively, below the top of Unit IV. From these dates, Schmits (1980) estimated that soil formation occurred between

Table 19. Dates defining soil forming intervals and/or ends of aggradational episodes.

Locality ^a	Dates (yrs. B.P.) ^b	Reference
1. Deer Creek	>2000; <2500	Johnson 1981
2. Wolf Creek	>2000; <2500	Johnson 1981
3. Big Blue River at 14P01	<2360; ca. 2000 (est.)	Schmits 1980
4. Little Blue River	ca. 2000-2500	Kopsick 1981
5. Osage River at 23BE1010	>1480; <2940	Lees et al. 1982
6. Pomme de Terre River	<1680	Brakenridge 1981
7. Munkers Creek at 14M0304	ca. 4000-2000 (est.)	Witty 1982
8. Cedar Creek at 14CS355	ca. 2000-1500 (est.)	Wood 1977
9. East Branch Walnut River	ca. 4000-2000	This study
10. Fall River at 14GR38	ca. 2000-1500 (est.)	Johnson 1971
11. Cotton Creek	2000-1200	Artz 1982
12. Candy Creek	<1631; <1758	Saunders 1980
13. Carnegie Canyon	2000-1000	Lintz and Hall 1983
14. Delaware Canyon	2065-1065	Hall 1982b
15. Central Washita River	<1000; <1760	Goss et al. 1972

^a Numbers refer to locations shown in Figs. 42, 44.

^b See text for discussion; ">" implies "older than", "<" implies "younger than".

ca. 2300 and 2000 B.P. Soil formation, according to Schmits (1980) was ended by renewed aggradation (Unit V) after ca. 2000 B.P.

At 23BE1010 on the Osage River (Fig. 42, loc. 5), a radiocarbon date of 1480 ± 70 B.P. was obtained from charcoal from alluvial sediment overlying a buried soil, suggesting that a transition from soil formation to aggradation occurred prior to this date. Soil formation occurred after 2940 ± 90 B.P., a date recovered from a buried hearth 140 cm beneath the buried soil surface (Lees et al. 1982).

At 14M0304, on Munkers Creek, a tributary of the upper Neosho River (Fig. 42, loc. 7), aggradation occurred between 5340 ± 80 B.P., slowing at ca. 3000-4000 B.P., and ceasing at ca. 2500-1500 B.P. The latter time intervals are inferred from the presence of El Dorado phase and Late Archaic and/or early Plains Woodland artifact forms (Witty 1982). Since the El Dorado phase material occurred within 50 cm of the present-day surface, rates of aggradation since ca. 4000 B.P. were probably slow, possibly accompanied by concurrent soil formation.

At 14CS355, on a tributary of the Cottonwood River (Fig. 42, loc. 8) and at 14GR38, on the upper Fall River (Fig. 42, loc. 10), archaeological strata associated with the buried A horizons of weakly-developed alluvial soils yielded large, corner-notched dart points and plain-surfaced, thick pot sherds tempered with grit and sand. Wood (1977) and Johnson (1971) inferred dates of ca. 2000-1500 B.P. (i.e., early Plains Woodland) from this evidence.

On Candy Creek, a tributary of Bird Creek in the Verdigris River basin (Fig. 43, loc. 12), archaeological strata in a soil similar to the

Copan Paleosol of nearby Cotton Creek valley produced dates of 1758 ± 82 and 1631 ± 89 B.P. (Saunders 1980). This suggests that the end of the soil-forming interval in this drainage occurred some time after ca. 1760-1630 B.P.

In Carnegie Canyon (Fig. 43, loc. 13), a deeply incised tributary of the Washita River, a floodplain soil similar to the Caddo Paleosol identified in Delaware Canyon (Hall 1982c) formed between 2000 and 1000 B.P. (Lintz and Hall 1983), and was buried after 1000 B.P. by renewed aggradation.

On the central Washita River, Goss et al. (1972:240) report radio-carbon dates on humates from 2 exposures of buried A horizons, one dating to 1760 ± 150 and the other to 1000 ± 100 B.P., indicating the end of soil formation intervals after these dates.

Using dates from Table 19, the estimated dates of major fluvial geomorphic discontinuity which occurred between ca. 1000-2000 B.P. are shown in Fig. 44. The general synchronicity of these responses supports the hypothesis that these geomorphic changes were climatically induced. The pattern is especially significant in that the alluvial chronologies used in this analysis represent responses in drainages of widely ranging sizes, from three major drainage systems, including the Missouri-Kansas, Arkansas and Red River systems. Synchronous responses over such a range is difficult to explain by other mechanisms, such as base-level changes.

The tendency, first noted in Fig. 43, for similar geomorphic responses to be time-transgressive from northwest to southeast is also apparent in Fig. 44. Discontinuities in Kansas and northwestern Missouri

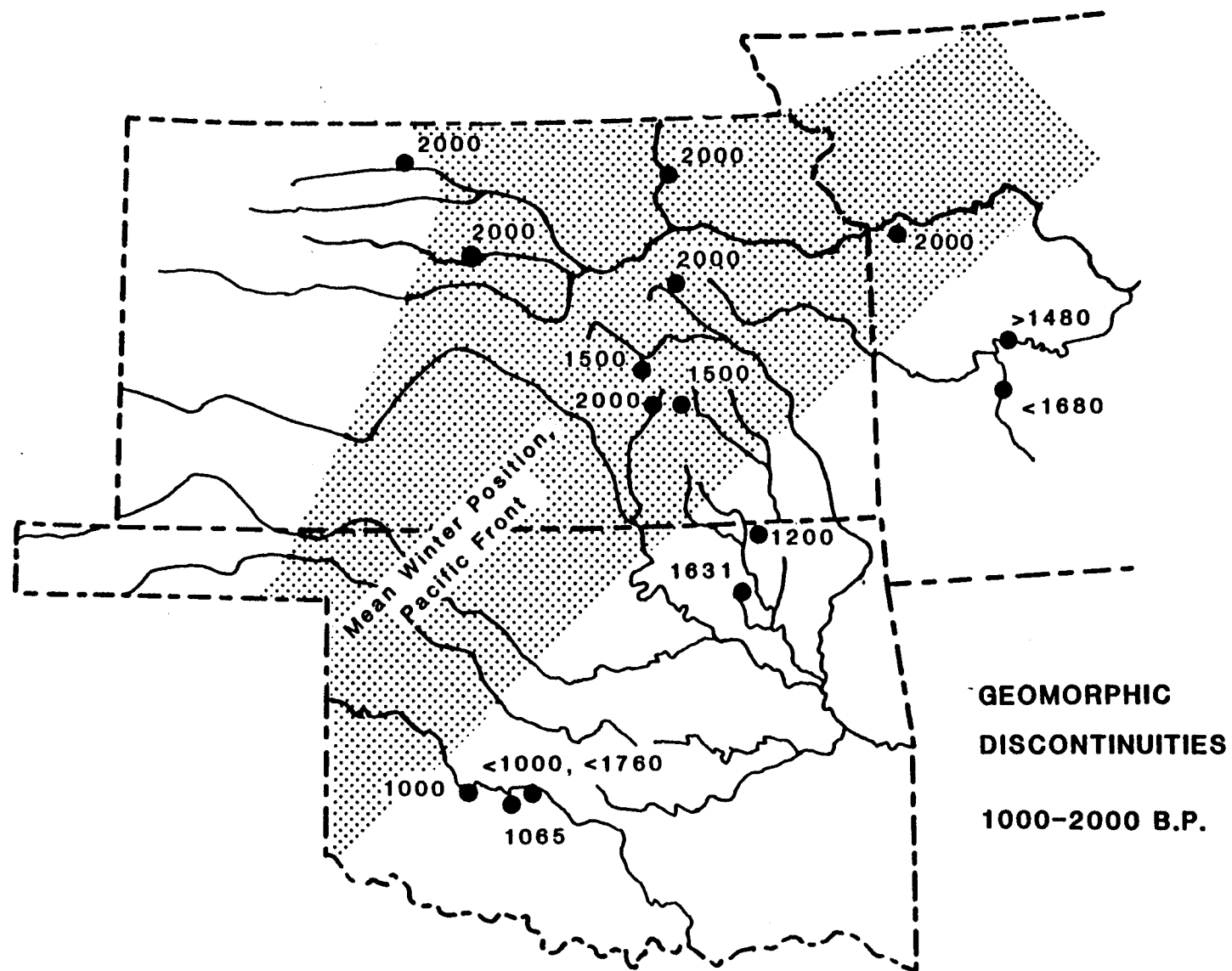


Fig. 44. Dates of major fluvial geomorphic discontinuities occurring 2000-1000 B.P. in Kansas, western Missouri and Oklahoma.

occurred at ca. 2000 B.P., with the exception of two sites (14CS355 and 14GR38), where the discontinuity is placed at ca. 1500 B.P. In west-central Missouri and northeastern and southwestern Oklahoma, discontinuities occurred between 1000 and 1500 B.P.

Correlation with Paleobiotic Records

Because geomorphic processes do not respond directly to climate it is difficult to infer the magnitude of a climate change from geomorphic evidence alone (Baker and P.-Orellana 1977). The model of geomorphic response developed earlier in this chapter suggested that shifts from geomorphic equilibrium to disequilibrium occurred with increasing aridity but did not attempt to specify how large an increase was required to prompt the geomorphic response. By considering paleobiotic evidence from the three-state region, the approximate magnitude of climatic shifts which prompted geomorphic change can be suggested.

The late Pleistocene and early-to-middle Holocene vegetational history of Missouri, Iowa and northeastern Kansas are documented by several palynological studies (e.g., Gruger 1973; King 1973; King and Allen 1977; Van Zant 1979). These studies, however, provide little or no detail on vegetation change during the past 5000 years. The pollen records from Muscotah Marsh, in northeastern Kansas (Gruger 1973), Lake West Okoboji in northwestern Iowa (Van Zant 1979) and Oldfield Swamp in southeastern Missouri (King and Allen 1977) document a gradual recovery during the late Holocene from a mid-Holocene interval of aridity (the Hypsithermal). Recovery began at these three sites between 6300 and ca. 5000 B.P., and was probably complete within ca. 2000 years, by which time a flora similar

to that of the present day had been established. These studies document no prominent late Holocene fluctuations in vegetation until those associated with land clearance and cultivation in the nineteenth century.

Pollen and land snail successions from the Cross Timbers of northeastern Oklahoma (Hall 1982b) do not extend deeper into the past than ca. 2000 B.P. These records show that a biota similar to the present was established in the area by that time. The records do, however, provide evidence of minor climatic fluctuations after 2000 B.P. These are discussed below. A number of pollen cores from the Quachita Mountains of southeastern Oklahoma, studied by Albert (1981) also provide evidence of late Holocene climatic change. Unfortunately, because of inconsistencies in the dating and correlation of strata among cores, Albert's interpretations have been challenged (Hall 1983). This study is not further discussed here.

The longest and best-preserved pollen record from northeastern Oklahoma is that recovered from Big Hawk Shelter, an archaeological site located on Hominy Creek, Osage County. The most important feature of this record, in terms of paleoclimatic change, is the increase in frequency of hickory (*Carya* spp.) pollen from 2% at ca. 1550 B.P. to a peak of 12% at 1350 B.P., followed by a gradual decline to 3% at 900 B.P. Since hickory is a mesic component of area forests, the increase in hickory provides evidence for increasing moisture availability from ca. 1550 to 1350 B.P. (Hall 1980b:44).

Land snail evidence from Big Hawk Shelter supports the palynological data. *Anguispira alternata*, the dominant moist-habitat snail, increased

in frequency from 15% at ca. 1900-2000 B.P. to a peak of 35% at 1450-1550 B.P., then decreased to 5% by ca. 450 B.P. Coincident with the decrease in both A. alternata and hickory was the appearance of the dry-habitat snail Triodopsis cragini at 1340 B.P. Although T. cragini never attained frequencies of over 1%, its appearance supports the inference of a shift toward drier conditions at ca. 1350 B.P. (Hall 1980b).

Although decreasing available moisture is noted as early as 1550 B.P. at Big Hawk Shelter, the full effect of drying was not felt until ca. 1000 B.P. At this time, the frequency of snails from dry wood habitats, principally of the genera Triodopsis and Polygyra, increases, while moist woods and moist prairie snails decrease. Similar increases in dry habitat snails occur at other rockshelters in Hominy Creek and nearby Birch Creek valleys at dates ranging from ca. 900 to post 880 B.P. (Hall 1982b:404). The early appearance of evidence of drying at Big Hawk Shelter is probably due to sample size (12,841 shells were analyzed from Big Hawk Shelter, compared to 736 to 1176 shells from other sites). The larger sample is probably better able to reflect more subtle changes than the smaller samples, which show only the dominant trends (Hall 1980b:45).

The pollen and land snail evidence summarized above documents changes in the relative frequency of species in an essentially present-day biota. Major climatic change, causing extirpation or replacement of species, is not indicated. The climatic change probably represents a relatively minor perturbation. The inferred decreases in available moisture between ca. 1000 and 1550 B.P. may reflect not so much a decrease in average

precipitation, but an increase in the variability of precipitation, perhaps resulting in an increased frequency of short-term droughts.

Despite the relatively subtle effects on local biota, the post-2000 B.P. climatic fluctuations in Oklahoma significantly influenced local fluvial systems. The peaks in abundance of hickory pollen and Anguispira alternata, both indicating moist habitats, occur during the interval within which the Copan Paleosol formed in the nearby Cotton Creek and Candy Creek valleys (Fig. 43c). Snails from the Caddo Paleosol in southwestern Oklahoma also indicate moist habitats during this interval (Hall 1982a,c; Lintz and Hall 1983). This suggests that geomorphic equilibrium and soil formation occurred in response to a greater availability of moisture. Similarly, the geomorphic discontinuities which ended the soil-forming interval at ca. 1250 B.P. in northeastern Oklahoma and at ca. 1000 B.P. in southwestern Oklahoma occurred during an interval of declining moisture availability, indicated by decreases in moist habitat snails and hickory pollen in the Cross Timbers rockshelters.

This correlation of geomorphic and biotic records in Oklahoma is consistent with the models of geomorphic response discussed earlier in this chapter. As noted by Schumm and Lichty (1963), an increase in the frequency of large magnitude floods may occur as a result of increased variability of precipitation (i.e., fluctuation between relatively moist and relatively dry years). Geomorphic evidence from Cotton Creek (Artz 1982) indicates that such a shift in flood characteristics occurred at approximately the same time as decreases in moist habitat plants and snails. This supports the above suggestion that climatic change at ca.

1550-1000 B.P. was induced by an increasing recurrence of short-term drought.

Similar geomorphic changes occurred at ca. 2000 B.P. at localities in Kansas and northeastern Missouri. Consistent with the models of geomorphic response discussed earlier in this chapter, it is postulated that soil formation prior to 2000 B.P. at these localities accompanied a geomorphic response to increased moisture availability. Entrenchment was initiated after 2000 B.P. in response to decreases in moisture availability.

Few studies provide evidence for inferring the characteristics of the late Holocene climate of Kansas and Missouri. However, three studies which provide such evidence all indicate increases in available moisture from ca. 3000 to at least 2000 B.P. At the Coffey site on the Big Blue River in northeastern Kansas, Schmits (1980:102) found that a paleo-channel of the river, probably formed a few centuries prior to 2500 B.P., was only 95 m wide, compared to the 140 m width of the 19th century channel. Schmits considered the narrower channel to be indicative of a narrow, deep, sinuous channel, in contrast to the wide, shallow, straight channel of historic time. Such a channel was probably formed by stream-flow dominated by frequent, moderate magnitude floods.

At Rodgers Shelter, on the Pomme de Terre River in west-central Missouri, an episode of intense upland erosion, indicated by an accumulation of coarse alluvial fan gravels, ended at ca. 3000 B.P., and was followed by the deposition of fine-grained colluvium from ca. 3000 to 1750 B.P. Ahler (1976:137) interpreted the transition at 3000 B.P. as

reflecting the recovery of vegetation with increased moisture, following an episode of relative aridity.

Molluscs from archaeological strata at Rodgers Shelter support Ahler's inference (Klippel et al. 1978). The predominant mussel species from Early and Middle Archaic layers (ca. 8600-3000 B.P.) prefer shallow, low-discharge stream habitats. Mussels from Late Archaic and Woodland layers (3000-1750 B.P.) prefer deeper, larger streams (Klippel et al. 1978:265-268), indicating increased streamflow during this interval.

The 1750-3000 B.P. interval of increased streamflow on the Pomme de Terre River includes the Tlb4 aggradational episode defined by Brakenridge (1981; see Fig. 6.7c), suggesting the correlation of geomorphic response with climate change. Unfortunately, there is not sedimentological or biotic evidence for climate change after ca. 1750 B.P. at Rodgers Shelter.

Conclusions

The preceding pages examined fluvial geomorphic sequences from the southeastern margin of the central plains to determine if other streams in this region experienced changes similar to and synchronous with those documented in the East Branch Walnut valley. Evidence for widespread geomorphic change between 1000 and 2000 B.P. was found, supporting the hypothesis that hydrologic and geomorphic change on the East Branch Walnut was climatically induced. In addition, the correlation of geomorphic and paleobiotic records from Oklahoma suggest that these climatic changes were consistent with those posited by the model of geomorphic response applied to the East Branch Walnut alluvial chronology. Specifi-

cally, episodes of soil formation, indicating geomorphic equilibrium, correlate with paleobiotic evidence for relatively moist conditions, while entrenchment occurred during intervals of decreasing moisture availability. The geomorphic shift is interpreted as a response to an increase in the frequency of large floods, which in turn is probably a product of increased rainfall intensity, and perhaps a greater variability of precipitation, producing a greater frequency of short-term drought.

Based on evidence discussed in the second section of this chapter, these geomorphic responses may have been induced by shifts in atmospheric circulation which influenced the seasonality, intensity and areal extent of precipitation. Geomorphic equilibrium is thought to represent an adjustment to an increased frequency of widespread, moderate intensity frontal storms. Disequilibrium, reflected by entrenchment, occurred in response to an increased frequency of large floods induced by intense thunderstorms.

The reason that geomorphic equilibrium was achieved earlier and ended sooner at some localities than others in the three-state region is not presently known. The timing of similar geomorphic responses was apparently time-transgressive across the region, occurring prior to 2000 B.P. in Kansas and northwestern Missouri and after 2000 B.P. in west-central Missouri and Oklahoma. Interestingly, the southeast-northwest direction of this time-transgressive response is the same as the direction of important vegetational and climatic gradients which characterize the southeastern periphery of the plains. As discussed previously, many of these gradients are importantly influenced by the relative frequencies

of Tropical and Pacific air masses over the region, and by the mean position of their convergence. Fig. 44 shows the mean position of the Pacific front in winter over the three-state region. The southeastern edge of this broad zone, as mapped by Bryson (1966), separates those localities at which geomorphic discontinuities occurred after ca. 1500 B.P. from those at which discontinuity occurred at ca. 2000 B.P. This correspondence may be coincidental; on the other hand, it may indicate that differences in geomorphic response on either side of this boundary are in some presently unknown manner related to the boundary.

It is likely that past geomorphic responses could be partially explained by models seeking modern-day analogues for past climatic conditions. The development of such models requires further study of the meteorological processes which at present produce gradients in precipitation characteristics across the southeastern periphery of the plains, and of the influence of these gradients on vegetation and hydrology. For example, it was suggested in a previous section of this chapter that the year 1973 provides an analogue for climatic and hydrologic conditions which may have characterized Episode A in the East Branch Walnut valley. A search for analogues of this type on a regional basis would be a first step toward explaining the patterns of climatically-induced geomorphic response documented in this chapter.

Additional study of alluvial chronologies and paleobiotic records from the region is also required to confirm and elaborate upon the patterns of synchronous response suggested in this chapter. Geomorphic response to Late Holocene climate change in the southeastern Central Plains is

undoubtedly not as concise as the pattern in Fig. 44 indicates. By emphasizing major events of geomorphic change, such as shifts from geomorphic equilibrium to disequilibrium, the present analysis overlooks many differences in the nature and timing of geomorphic episodes among the localities discussed. As chronologies from other localities become available, or as the existing chronologies are compared in greater detail, examples of non-synchronous responses will undoubtedly be recognized.

Differential response, however, is expected. The nature and timing of geomorphic responses are importantly influenced by local watershed parameters, internal to individual valleys and drainage basins. An external influence, such as climate, is capable of inducing geomorphic change only when internal thresholds, established by local conditions, are exceeded (Schumm 1973). These internal parameters include the size and shape of the drainage basin, and the relief, vegetation and lithology of the basin. All of these parameters vary considerably across the three-state region considered above. In fact, given the range of watershed parameters represented in the localities shown in Fig. 44, the degree of regional synchronicity evident is even more noteworthy.

Thus, evidence for non-synchronous response, perhaps including the time-transgressive aspects of response noted above, do not necessarily contradict generalized patterns suggesting synchronicity. The significance and goodness-of-fit of any single chronology within a generalized regional pattern can only be evaluated when the processes of geomorphic response at the local level have been considered. At the same time, as the present analysis demonstrates, patterns of regional synchronicity are evident, but are presently definable only in terms of major geomorphic

discontinuities or trends observed at relatively broad time scales of 500-1000 years.

The necessity of viewing geomorphic response to climatic change at both the local and regional levels is illustrated by the following analogy. Viewed from afar, an impressionistic painting exhibits a discernible pattern. However, the viewer, rushing in for a closer look, may be dismayed to see the pattern dissolve into a confusion of individual dots and dashes of color. At this point, the viewer may either leave the gallery in disgust, or he may experience the fundamental revelation that an impressionist's work can be viewed from two perspectives. One can stand back and appreciate the overall effect of the artist's work, or one can step closer and appreciate, in detail, the techniques and processes by which the master achieved the overall effect. In the long run, to properly appreciate the painting, it is necessary to do both.

Like an impressionistic painting, the geomorphic record of the southeastern central plains is comprised of an aggregate of individual valleys, each of which has responded to regional climate change within the constraints of local watershed parameters. Although from a regional perspective, generalized patterns of synchronous response are discernible, it is also possible, and necessary, to consider the manner in which individual basins responded to region-wide change.

The present study has used both perspectives. The bulk of this thesis has attempted to provide a detailed reconstruction of the pedologic and geomorphic processes reflected in the soil-geomorphic records of a single drainage basin. In the last section, however, the perspective was enlarged in order to demonstrate the position of the East Branch

Walnut alluvial chronology in a broader, regional pattern of geomorphic response to climate change.

CHAPTER 7

SUMMARY AND CONCLUSIONS

SUMMARY

This thesis investigated the paleoenvironmental implications of the alluvial soils and landforms of the East Branch Walnut valley, in the southern Flint Hills of eastern Kansas. To accomplish this objective, the soils and landforms of the study area were viewed as the products of pedogenic and geomorphic processes, respectively. Each set of processes, in turn, are influenced by complex systems of environmental parameters and processes. Pedogenic processes in alluvial landscapes are importantly influenced by the age, geomorphic stability and drainage characteristics of the surfaces beneath which soils form. These variables, in turn, are the product of geomorphic processes which shape the alluvial landscape by eroding, transporting and depositing sediment. Since flowing water provides the energy for this geomorphic work, and sediment the material on which work is performed, geomorphic processes respond to variation in the hydrologic parameters of streamflow and sediment yield. The supply of water and sediment to the stream is in turn influenced by environmental phenomena including climate, vegetation, physiography and geology.

The thesis traced this hierarchical chain of process and response, beginning with the analysis of individual soil pedons and ending with a consideration of regional geomorphic responses to climatic change. Each stage in the analysis consisted of "taking a step back", looking at the phenomena discussed in the preceding stage from a broader, more inclusive

and more generalized perspective.

In Chapters 3 and 4, the analysis focused on the pedogenic processes manifested by soil properties in pedons at five archaeological sites. Using a chronological framework provided by archaeological evidence, two soil-stratigraphic units were recognized. The dominant pedogenic trend manifested in the Nuttal Soil, formed between 1000-1500 B.P. and the present, was horizonation, or the differentiation of soil properties with depth into discrete layers or horizons. Mineral weathering and clay illuviation contributed to the development of minimal argillic horizons in the Nuttal Soil. The dominant pedogenic trend manifested in the Snyder Soil, formed between 4000 and 2000 B.P., was haploidization, the maintenance of weakly differentiated soil properties with depth. Haploidization of the weakly-developed Snyder Soil was accomplished by the processes of cumulization and homogenization.

In Chapter 5, the Nuttal and Snyder Soils were considered from soil-geomorphic and soil-stratigraphic perspectives within the broader context of the East Branch Walnut alluvial landscape. The different pedogenic trends manifested by the soils were explained in terms of their different geomorphic contexts. The Nuttal Soil occurs beneath the well-drained, convex or gently sloping surfaces of abandoned levees and colluvial footslopes, while the Snyder Soil formed beneath less well-drained, level or depressional surfaces of floodbasins and backwaters.

From a soil-stratigraphic perspective, the Nuttal and Snyder Soils were viewed as stratigraphic markers and as indicators of depositional environments. Consideration of soil-stratigraphic and soil-geomorphic

evidence permitted the formulation of an alluvial chronology which documented a sequence of episodic geomorphic change in the East Branch Walnut valley since ca. 4500 B.P.

In Chapter 6, the geomorphic processes manifested in the alluvial chronology were viewed as responses to climatically-induced changes in streamflow and sediment yield. Consistent with models of fluvial geomorphic response proposed by Schumm and Lichty (1963) and Baker and P.-Orellana (1977), the three episodes of the chronology are reasonably interpreted as different adjustments to changes in flood magnitude and frequency. These changes, in turn, may have been induced by changes in precipitation patterns. At present, the East Branch Walnut valley is apparently adjusted to the prevailing flood magnitude-frequency relationships, which in turn are importantly influenced by precipitation patterns. Past geomorphic responses manifested by the East Branch Walnut alluvial chronology may be explained in terms of variations from the modern hydrologic regime. The following interpretations are tentatively offered.

During Episode A (4500-2000 B.P.), the East Branch Walnut valley experienced a greater frequency of moderate magnitude floods, relative to large magnitude floods. At present, similar floods are generated by the widespread, moderate intensity frontal storms of early spring and fall.

During Episode B (2000-1850 B.P.), the frequency of large floods increased, perhaps in response to an increased frequency of intense thunderstorms. It is assumed that stream channels widened and deepened to accomodate the larger discharges, forming a broad, deep trench. The

deposition of colluvial footslopes and levees during Episode B reflects increases in sediment yield, perhaps in response to increased upland erosion. Increased upland erosion may reflect a greater frequency of short-term, severe droughts during Episode B.

During Episode C (1850 B.P. to present), the frequency of large floods decreased, and the T-0 fill aggraded within the Episode B trench. It is postulated that Episode C marks the development and maintenance of a geomorphic adjustment to the present hydrologic regime. Evidence for levee building during Episode C indicates that this adjustment was disrupted at least once by a short-term increase in the frequency of large floods and in sediment yield, possibly in response to severe drought.

The final stage of the analysis considered regional implications of the climatic changes posited in the above sequence. Since the postulated shifts in precipitation patterns involve changes in the frequency and position of air masses on a continental scale, it was suggested that geomorphic changes like those documented in the East Branch Walnut valley should have occurred throughout the region. Geomorphic data from 15 localities in Kansas, Oklahoma and Missouri indicated that major geomorphic discontinuities similar to that separating Episodes A and B in the East Branch Walnut valley were widespread along the southeastern periphery of the Central Plains between 2000 and 1000 B.P., occurring earlier in the northwest than in the southeast. Although the reason for this time-transgressive response is not known, it is postulated that the climate change involved shifts in southeast-northwest gradients in the

seasonality, intensity and variability of rainfall observed at present in the region.

A review of the limited paleobiotic evidence from the region indicates that alluvial soil formation, representing geomorphic equilibrium, occurred during episodes of increased moisture availability, and that entrenchment and erosion accompanied decreases in moisture availability. The paleobiotic records indicate that Late Holocene climatic changes represent relatively minor perturbations, and are perhaps best interpreted as changes in the variability of precipitations (e.g., changes in the frequency of short-term droughts) rather than changes in average or absolute amounts of rainfall.

CONCLUSIONS

Climatic change is not directly reflected in the soils and landforms of the East Branch Walnut valley, but it is possible to infer such changes, at least tentatively, if the processes of hydrologic, geomorphic and pedologic response to environmental change are understood, and if the manifestations of those responses in the stratigraphic record are properly interpreted. Caution is required, since the response of these phenomena to climate is extremely complex. Consideration of multiple lines of evidence is necessary before inferences of past climate can be accepted.

In attempting to infer past climates of the East Branch Walnut valley from soil-geomorphic evidence, the present study relied on models of geomorphic response which were originally developed to account for geomorphic change on streams which are quite different from the East Branch Walnut

(e.g., the Colorado River, in Texas, and the Cimarron River, in southwestern Kansas). The applicability of these models to the specific case of the East Branch Walnut was supported by three lines of evidence. First, the present-day hydrology of Flint Hills streams suggests similar correlations between streamflow and precipitation patterns as those posited by the models. Second, similar geomorphic responses occurred over a large area at approximately the same time, supporting the notion that the responses were climatically induced. Third, the correlation of paleobiotic and geomorphic records from northeastern Oklahoma and west-central Missouri reveals correspondence between moisture regime and geomorphic episodes similar to those posited by the models.

Despite the support lent by these lines of evidence, the climatic changes inferred for the East Branch Walnut are considered tentative for two reasons. First, the present-day processes of hydrologic and geomorphic response to climate in the study area require further investigation, in an attempt to identify processes and responses which might be analogous to past conditions. Second, paleobiotic evidence from the southern Flint Hills, which would provide independent evidence for climate change in the immediate vicinity of the study area, is not available.

These limitations, however, apply primarily to the latter stages of analysis, in which the East Branch Walnut alluvial chronology was related to processes of climatic and hydrologic change. A degree of uncertainty is expected in such an analysis, given the complexity of geomorphic responses to climate and hydrology. Relatively more confidence is placed in the results of the earlier stages of analysis, in which pedogenic

responses to geomorphic change were inferred, and a chronology for geomorphic change established. The soil-geomorphic relationships discussed in Chapters 3-5 are perhaps the most important contribution of this thesis.

The discussion of soil-geomorphic relationships and alluvial chronology in the East Branch Walnut valley will hopefully be of value to future researchers in developing and evaluating more refined models of fluvial geomorphic response in the southeastern Central Plains.

ARCHAEOLOGICAL CONTRIBUTIONS

Although this thesis deals primarily with pedologic, geomorphic, hydrologic and climatic phenomena, archaeological data contributed to all stages of the analysis. The most important of these contributions was the provision of a chronological framework. The relative and absolute ages of archaeological deposits from various soil-stratigraphic and soil-geomorphic contexts contributed to the definition of soil-stratigraphic units in Chapter 3 and to the establishment of an alluvial chronology in Chapter 5. Also, much of the geomorphic data used in Chapter 6 to document regional geomorphic response relied on chronological and stratigraphic data from archaeological sites.

Archaeological data also contributed to the inference of pedogenic processes operative during formation of the Snyder Soil. Radiocarbon dates from the stratified Late Archaic occupations at 14BU9 provided evidence for the slow, gradual aggradation of Snyder Soil surfaces, which suggested the importance of the pedogenic process of cumulation. The process of homogenization was inferred from the tendency for artifacts

buried in the Snyder Soil at 14BU9 to gradually decline in frequency above and below levels of maximum frequency. These data suggested that pedoturbational processes had dispersed material from originally discrete occupational layers. Cumulization and homogenization, although important processes in the genesis of alluvial soils (cf. Riecken and Poetsch 1960; Hoeksama 1953), are difficult to infer from pedologic evidence alone. This study suggests the utility of archaeological data for recognizing the effects of these processes in soil genesis.

Archaeology benefitted from, as well as contributed to, this study. Several of these benefits, although essentially "spin-offs" from the study's primary goal of environmental reconstruction, are worth noting, since they demonstrate the broad range of specifically archaeological problems to which pedologic and geomorphic data are potentially applicable.

Preservation Potential of Bone

Comparison of the depth distributions of animal bone and available phosphorous at 14BU9 and 14BU25 indicates that bone was most abundant at depths below the zone of maximum intensity of phosphorous leaching. This suggests that the decomposition of bone (a principle constituent of which are calcium phosphate minerals) is accelerated in horizons in which soil phosphorus is most subject to use by plants or translocation by soil water. Bone is preserved deeper in the profile, where the drain on soil phosphorus is less intensive.

These observations are important to archaeologists, who are commonly confronted with changing frequencies of animal bone, both within sites

(with depth) and among sites. Before basing inferences of prehistoric behavior on these frequencies, it is necessary to consider the extent to which the frequencies may reflect differential preservation.

Detection of Anthropogenic Organic Matter

In discussing the organic matter content of the Snyder Soil in Chapter 4, it was noted that loss-on-ignition revealed more pronounced patterns of anthropogenic additions of organic matter than the Walkley-Black method. The latter method, although widely used by soil scientists, does not measure "inert" or elemental forms of organic carbon such as charcoal, while LOI achieves complete oxidation of all carbon. The increases in LOI within occupational levels in the Snyder Soil probably reflect the concentration of finely-divided charcoal particles. Since charcoal is a principal constituent of many archaeological residues, a simple, inexpensive technique such as LOI might be used to define the presence and occupational intensity of buried archaeological strata. Such applications, however, are most reliable in non-calcareous soils, since carbonates are also lost during ignition at high temperatures.

Site Selection

A number of Plains Woodland sites in the East Branch Walnut valley are situated on levees on the T-1 surface (e.g., 14BU4, 14BU9, 14BU31, 14BU55, 14BU57 and 14BU87, discussed in Chapter 5). Artz (1983), analyzing artifact distributions in a systematic surface collection of the Woodland component at 14BU4, suggested that habitation at the site concentrated on the crest of the levee, while off-levee areas were used

primarily for refuse dumping and specialized activities such as flint-knapping. Although the levees are relatively subtle relief features, the moderately-developed profile of the Nuttall Soil developed beneath some of the levees (e.g., 14BU4, 14BU9, 14BU87) indicates that the convex levee surfaces are substantially better-drained than adjacent off-levee surfaces (Chapter 5). This pedologic evidence perhaps indicates that the levees were attractive as loci for prehistoric occupation because of drainage characteristics. Simply stated, the Plains Woodland occupants at 14BU4 may have selected that particular site location for the same reason that a Bt horizon was forming beneath their feet--good surface drainage.

Culture History and Prehistoric Human Ecology

A review of the prehistory of the East Branch Walnut valley (Chapter 2) indicated that prehistoric occupations of the valley were most extensive between 4500 and 2000 B.P. in the Late Archaic period, and from ca. 1200-900 B.P. in the Late Plains Woodland period. Occupation of the area from 2000-1200 B.P. and after 900 B.P. was apparently not extensive. Sites pre-dating 4500 B.P., if present, are deeply buried beneath T-1 surfaces and are unknown.

The 800 year period of non-extensive occupation from 2000-1200 B.P. includes Episode B in the East Branch Walnut alluvial chronology. This episode may have been a time of frequent, large floods, perhaps induced by increased frequencies of short-term drought and probably increased climatic variability. For prehistoric human groups inhabiting the area, or moving through the area within a seasonal round, increased climatic

variability would have meant decreased predictability of climate-dependent resources. The risks attending environmental unpredictability may have been sufficient to preclude extensive habitation of the southern Flint Hills during this interval.

A comparison of prehistoric cultural chronologies with the regional patterns of environmental change posited in this thesis is beyond the scope of this study. Such a comparison, if undertaken, would undoubtedly confront many of same general problems confronted in this thesis. Archaeological evidence does not directly reflect environmental change, but rather reflects human responses to such change. Like the soils, landforms and streams discussed in this thesis, the people of the East Branch Walnut valley responded to environment through a unique set of behavioral processes. An analytical approach to understanding prehistoric human adaptation to changing environments in the southern Flint Hills would need to infer the behavioral processes reflected in the archaeological record of the region, and then consider how these processes might be interpreted in terms of human responses to environmental change.

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APPENDIX A

REPRESENTATIVE PROFILES OF ALLUVIAL SOIL SERIES

The following profiles describe pedons which Penner et al. (1975) consider to be representative of the Verdigris, Brewer, Vanoss and Norge series in Butler County, Kansas. Profile descriptions are quoted from Penner et al. (1975); measurements, reported in inches in the original, are converted to centimeters to aid comparison with pedons described in Chapter 3.

VERDIGRIS SILT LOAM

- Ap -- 0 to 20 cm, dark grayish brown (10YR4/2) silt loam, very dark grayish brown (10YR3/2) when moist; weak, medium, granular structure; slightly hard when dry, friable when moist; slightly acid; clear, smooth boundary.
- A11 -- 20-41 cm, dark-gray (10YR4/1) silty clay loam, black (10YR2/1) when moist; strong, fine, granular structure; slightly hard when dry, friable when moist; few worm casts; numerous open pores; slightly acid; gradual, smooth boundary.
- A12 -- 41-84 cm, dark-gray (10YR4/1) silty clay loam, dark grayish brown (10YR4/2) when crushed, black (10YR2/1) when moist; strong, very fine, irregular, blocky structure that has a subangular blocky component; hard when dry, friable when moist, numerous open pores; slightly acid; gradual, smooth boundary.
- AC -- 84-144 cm, very dark grayish brown (10YR3/2) silty clay loam; very dark brown (10YR2/2) when moist; common, fine, distinct mottles of strong brown (7.5YR5/6) when moist; moderate, fine, blocky structure; hard when dry, friable when moist; few open pores; slightly acid.

The Ap, A11 and A12 horizons are silt loam or silty clay loam, and the combined thickness is 51-89 cm. The color ranges from dark gray to very dark grayish brown. Reaction is slightly acid to medium acid. The AC horizon is silt loam or silty clay loam. Its color ranges from dark gray to very dark grayish brown. Reaction ranges from slightly acid to medium acid. In some places, mottles of strong brown are below a depth of 76 cm.

BREWER SILTY CLAY LOAM

- A1 -- 0-36 cm, dark gray (10YR4/1) silty clay loam, black (10YR2/1) when moist; moderate, medium, granular structure; hard when dry, friable when moist; many open pores; many roots; slightly acid; gradual, smooth boundary.

- B1 -- 36-53 cm, very dark gray (10YR3/1) heavy silty clay loam, black (10YR2/1) when moist; moderate, fine, subangular blocky structure; extremely hard when dry, firm when moist; few open pores; thin, patchy clay films; common roots; slightly acid; gradual, smooth boundary.
- B2t -- 53-104 cm, dark gray (10YR4/1) silty clay, very dark gray (10YR3/1) when moist; common, medium, distinct, dark brown (7.5YR4/4) mottles; moderate, medium, blocky structure; extremely hard when dry, very firm when moist; clay films cover all peds; few weak slickensides; few roots; few open pores; neutral; gradual, smooth boundary.
- C -- 104-168 cm, dark gray (10YR4/1) silty clay, very dark gray (10YR3/1) when moist; common, medium, distinct, dark brown (7.5YR4/4) mottles; massive; extremely hard when dry, very firm when moist; few open pores; neutral.

The A1 horizon is silt loam or silty clay loam and ranges from 20 to 46 cm in thickness. Its color ranges from dark gray to very dark grayish brown. Reaction is slightly acid to neutral. The B2t horizon is silty clay or clay 48-89 cm thick. Reaction is slightly acid to neutral, and the color range is the same as that for the A1 horizon.

VANOSS SILT LOAM

- A1 -- 0-30 cm, dark gray (10YR4/1) silt loam, very dark gray (10YR3/1) when moist; moderate, fine, granular structure; slightly hard when dry, very friable when moist; numerous open pores; slight grayish coatings on peds in lower part; slightly acid; gradual smooth boundary.
- B1 -- 30-46 cm, dark grayish brown (10YR4/2) light silty clay loam, very dark grayish brown (10YR3/2) when moist; moderate, fine and very fine, subangular blocky structure; hard when dry, friable when moist; many open pores; slightly acid; gradual, smooth boundary.
- B2t -- 46-89 cm, brown (7.5YR5/3) silty clay loam, dark brown (7.5YR4/3) when moist; strong, fine and very fine, blocky structure; hard when dry, friable when moist; many open pores; few, thin, patchy clay films; few very fine black concretions; slightly acid; gradual, smooth boundary.
- B3 -- 89-145 cm, brown (7.5YR5/4) light silty clay loam, dark brown (7.5YR4/4) when moist; weak, medium blocky structure; hard when dry, friable when moist; few, thin, patchy clay films; few very fine black concretions; slightly acid.

The A1 horizon is silt loam or light silty clay loam and is 20 to 36 cm thick. Its color ranges from dark gray to very dark grayish brown. Reaction is slightly acid to medium acid. The B1 horizon is heavy silt loam or silty clay loam and is 10-20 cm thick. Its color is dark grayish brown to dark brown. Reaction is neutral to medium acid. The B2t horizon is silty clay loam and is 38 to 64 cm thick. Its color

ranges from dark yellowish brown to brown. Reaction is neutral to medium acid. The B3 horizon is light silty clay loam or heavy silt loam, and it ranges from brown to yellowish brown in color. Reaction is neutral to medium acid.

NORGE SILT LOAM

- A1 -- 0-23 cm, dark brown (7.5YR4/2) heavy silt loam, dark brown (7.5YR3/2) when moist; moderate, fine and medium granular structure; slightly hard when dry, very friable when moist; few open pores; roots common; slightly acid; gradual, smooth boundary.
- B1 -- 23-46 cm, reddish brown (5YR4/3) silty clay loam, dark reddish brown (5YR3/4) when moist; strong, fine, subangular blocky structure; hard when dry, friable when moist, few open pores; thin and patchy clay films; slightly acid; gradual, smooth boundary.
- B2lt -- 46-81 cm, reddish brown (5YR4/5) silty clay loam, dark reddish brown (5YR4/5) when moist; strong, medium, blocky structure; very hard when dry, firm when moist; many open pores; thin and continuous clay films; few fine iron-manganese concretions; slightly acid; gradual, smooth boundary.
- B22t -- 81-107 cm, reddish-brown (5YR5/5) heavy silty clay loam, yellowish red (5YR5/6) when moist; weak, medium, blocky structure; very hard when dry, firm when moist; thin and continuous clay films; few open pores; few iron-manganese concretions; slightly acid; gradual smooth boundary.
- B31 -- 107-142 cm, yellowish red (5YR5/6) heavy silty clay loam, yellowish red (5YR4/6) when moist; weak, medium, blocky structure; very hard when dry, firm when moist; few small iron-manganese concretions, few roots; slightly acid; gradual, smooth boundary.
- B32 -- 142-183 cm, yellowish red (5YR5/6) silty clay loam, yellowish red (5YR4/6) when moist; weak, coarse blocky structure; very hard when dry, friable when moist; few roots; slightly acid.

The A horizon ranges from brown to dark brown and is silt loam or light silty clay loam 18 to 30 cm thick. Reaction is slightly acid to medium acid. The B1 horizon ranges from dark brown to dark reddish brown and is heavy silt loam or silty clay loam 13 to 30 cm thick. Reaction is slightly acid to medium acid. The B2lt and B22t horizons are silty clay loam or heavy silty clay loam 51 to 102 cm thick, and they range from reddish brown to yellowish red and dark reddish brown. Reaction is neutral to slightly acid in these horizons. In most places, the B3 horizon is silty clay loam, but it ranges to heavy silt loam in places. Its color ranges from yellowish red to reddish brown. Reaction is neutral to slightly acid. In some places chert pebbles are in thin lenses.

APPENDIX B
SOIL PROPERTIES

Table B.1. Physical and chemical properties, 14BU4 (XU102).

Horizon	Depth (cm)	pH	Avail. P (ppm)	Org. Matter %	LOI		Sand %	Texture	
					400° C %	600° C %		Silt %	Clay %
Ap/A	15-25	6.5	110	2.6	2.39	5.07	18.2	56.2	25.6
Bt	45-55	7.2	74	2.7	1.29	3.38	15.2	52.7	32.1
Ab	75-85	7.3	55	1.0	1.39	3.44	15.8	56.6	27.6
ACb	130-140	7.3	61	0.9	1.30	3.54	14.4	51.5	34.1
	140-150	7.3	63	1.1	1.67	3.62	18.2	46.1	35.7
	150-160	7.2	58	1.0	n.d.	4.01	15.8	48.6	35.6
	160-170	7.3	59	0.9	1.70	3.94	15.8	48.2	36.0
	170-180	7.3	62	0.8	1.56	3.74	16.8	48.5	34.7
Cb	180-190	6.0	43	0.7	1.59	3.35	17.3	47.5	35.2
	200-210	6.4	43	0.7	1.59	3.34	14.8	50.7	34.5
	220-230	6.6	46	0.7	1.65	3.55	13.4	51.6	35.0

Table B.2. Physical and chemical properties, 14BU4, Trench C.

Horizon	Depth (cm)	pH	Avail. P (ppm)	Org. Matter %	LOI		Sand %	Texture	
					400° C %	600° C %		Silt %	Clay %
A11	24-40	6.9	63	2.6	3.18	5.08	15.4	60.6	24.0
	45-55	7.2	43	1.9	2.70	4.12	14.4	54.6	31.0
A12	70-80	7.1	54	1.2	2.58	4.03	15.4	51.0	33.6
AC	100-110	6.9	49	1.3	2.42	3.86	13.9	51.5	34.6

Table B.3. Physical and chemical properties, 14BU9.

Horizon	Depth (cm)	pH	Avail. P (ppm)	Org. Matter %	LOI		Sand %	Texture	
					400° C %	600° C %		Silt %	Clay %
A	20-30	6.4	80	1.6	2.64	4.29	15.4	57.0	27.6
Bt	35-40	6.8	52	0.8	1.83	3.63	15.4	51.0	33.6
	45-50	6.7	45	1.0	1.83	3.43	14.4	53.0	32.6
Ab	55-60	6.8	39	1.0	1.82	3.56	15.0	59.0	26.0
	65-70	6.8	38	1.0	1.81	3.50	13.8	53.5	32.7
ACb	75-80	6.8	39	1.0	1.84	3.53	13.3	50.5	36.2
	85-90	6.8	40	1.0	1.68	3.66	13.4	52.4	34.2
	95-100	6.8	42	1.0	1.60	3.31	13.4	52.4	34.2
	105-110	6.7	53	1.0	1.60	3.43	15.4	51.4	33.2
	115-120	6.7	69	1.15	1.19	3.41	12.4	51.0	36.6
	125-130	6.6	59	1.3	1.42	2.85	13.4	50.4	36.2
	135-140	6.7	50	1.1	1.31	2.91	13.1	50.5	36.4

Table B.4. Physical and chemical properties, 14BU25.

Horizon	Depth (cm)	pH	Avail. P (ppm)	Org. Matter %	LOI		Sand %	Texture	
					400° C %	600° C %		Silt %	Clay %
A	00-10	5.9	49	2.8	4.59	6.72	24.8	40.0	35.6
	10-20	6.6	9	3.0	4.88	7.12	26.6	37.6	35.8
	20-30	6.7	6	2.6	5.89	7.48	24.2	36.6	39.2
B1	30-40	6.8	1.5	2.4	5.05	7.48	22.8	38.0	39.2
	40-50	6.9	1	2.1	4.57	7.28	23.8	38.0	38.2
	50-60	6.9	1.5	1.8	4.23	6.44	21.0	38.4	40.6
	60-70	6.9	1	1.5	3.73	5.33	15.8	41.6	42.6
B2t	70-80	7.0	1	1.3	3.09	5.07	16.8	42.6	40.6
	80-90	7.0	2	1.2	2.83	4.64	15.0	46.5	38.5
Ab	90-100	7.0	1	1.0	2.60	4.26	17.6	47.0	35.4
	100-110	7.0	2	1.0	2.50	4.05	16.6	48.0	35.4
	110-120	7.0	1	0.9	2.35	3.98	16.6	46.0	37.4
	120-130	7.1	1.5	0.9	2.08	3.80	17.8	44.0	38.2
	130-140	7.0	2	0.8	2.01	3.75	17.0	46.5	36.5
Cb	140-150	7.0	1.5	0.7	1.94	3.67	17.0	46.0	37.0
	150-160	7.1	1.5	0.7	1.75	3.49	15.8	49.2	35.0
	160-170	7.1	1.5	1.6	1.72	3.43	15.8	48.6	35.6
	170-180	7.1	4	0.5	1.61	3.19	15.0	52.0	33.0
	180-190	7.1	6	0.5	1.48	3.00	14.8	52.6	32.6
	190-200	7.1	10	0.4	1.26	2.68	15.8	51.0	33.2
	200-210	7.2	13	0.3	1.13	2.51	16.2	52.6	31.2
	210-220	7.2	15	0.4	1.22	2.67	17.2	48.5	34.3
	220-230	7.2	24	0.5	1.42	2.77	22.3	40.9	36.8

Table B.5. Physical and chemical properties, 14BU81, TP1.

Horizon	Depth (cm)	pH	Avail. P (ppm)	Org. Matter %	LOI		Sand %	Texture	
					400° C %	600° C %		Silt %	Clay %
Ap	0-20	5.9	11	1.8	1.10	4.46	14.2	57.4	28.2
A	20-30	6.0	7	1.5	2.20	4.19	16.3	53.4	30.4
	30-40	6.2	8	1.1	2.44	3.87	16.4	54.8	28.8
	40-50	6.1	9	1.1	1.80	3.46	15.2	54.0	30.8
AC	50-60	6.2	11	1.1	1.56	3.30	17.4	52.5	30.1
	60-70	6.2	13	0.8	1.37	3.09	15.0	55.6	29.4
	70-80	6.3	16	0.8	1.37	2.76	16.0	53.9	30.1
	80-90	6.3	20	0.9	1.26	2.82	14.4	56.0	29.6
	90-100	6.3	24	0.7	1.12	2.78	14.4	54.6	31.0

Table B.6. Physical and chemical properties, 14BU81, TP2.

Horizon	Depth (cm)	pH	Avail. P (ppm)	Org. Matter %	LOI		Sand %	Texture	
					400° C %	600° C %		Silt %	Clay %
Ap	0-20	6.0	13	1.8	2.68	4.89	14.0	59.0	27.0
A	20-30	6.1	13	1.2	2.19	3.94	16.5	53.5	30.0
	30-40	6.1	16	1.1	2.05	3.72	13.5	54.3	32.2
	40-50	6.1	16	1.2	1.87	3.44	14.0	54.0	32.0
AC	50-60	6.0	22	1.0	1.74	3.26	16.0	51.4	32.6
	60-70	6.2	24	1.1	1.66	3.28	16.8	53.0	30.2
	70-80	6.3	21	0.9	1.47	3.22	13.8	54.2	32.0
	80-90	6.4	24	0.7	1.44	3.17	13.8	54.5	31.7
	90-100	6.3	28	0.7	1.41	2.95	18.8	50.0	31.2
	100-110	6.4	28	0.7	1.36	3.07	15.4	54.4	30.2
	110-120	6.2	29	0.6	1.30	3.04	16.4	52.0	31.6
	120-130	6.5	30	0.6	1.22	2.92	15.0	54.3	30.7
	130-140	6.2	35	0.7	nd	2.67	12.8	57.0	30.2
	140-150	6.3	40	0.6	1.05	2.74	13.0	57.0	30.0
	150-160	6.6	42	0.6	1.03	2.55	13.0	71.0	16.0

Table B.7. Physical and chemical properties, 14BU87.

Horizon	Depth (cm)	pH	Avail. P (ppm)	Org. Matter %	LOI		Sand %	Texture	
					400° C %	600° C %		Silt %	Clay %
Ap	0-10	6.1	47	1.4	2.50	4.27	17.0	55.0	28.0
Bt	10-20	6.4	34	1.0	2.26	4.19	15.0	51.0	34.0
	20-30	6.4	38	1.3	2.17	4.02	14.4	52.6	33.0
	30-40	6.4	29	1.1	2.13	3.87	14.4	54.0	31.6
ACb	40-50	6.4	25	1.1	2.08	3.75	13.0	54.3	32.7
	50-60	6.5	19	0.9	1.83	3.38	15.0	52.8	32.2
	60-70	6.5	19	0.9	1.69	3.00	13.8	56.2	30.0
	70-80	7.6	7	0.8	1.53	2.94	13.3	59.2	27.2
	80-90	6.3	25	0.7	1.56	2.95	15.5	56.9	27.6
	90-100	6.0	33	0.6	1.42	2.40	14.0	56.0	30.0