THE GEOARCHAEOLOGY OF THE LITTLE MISSOURI BADLANDS:
THE LATE QUATERNARY STRATIGRAPHIC AND PALEOENVIRONMENTAL
CONTEXT OF THE ARCHAEOLOGICAL RECORD

A Dissertation
by
DAVID DUANE KUEHN

Submitted to the Office of Graduate Studies of
Texas A&M University
in partial fulfillment of the requirements for the degree of
DOCTOR OF PHILOSOPHY

May 1995

Major Subject: Anthropology
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May 1995

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ABSTRACT

The Geoarchaeology of the Little Missouri Badlands:

The Late Quaternary Stratigraphic and Paleoenvironmental Context of the Archaeological Record. (May 1995)

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Geoarchaeological studies undertaken in the South Unit of Theodore Roosevelt National Park in the Little Missouri Badlands of western North Dakota provide a geological and environmental framework for the interpretation of the archaeological record. Within this framework, archaeological materials are viewed not as static objects, but as interactive components of a highly dynamic natural environment. An archaeological survey of 4840 ha provides evidence for prehistoric utilization of the region from Paleoindian through Late Prehistoric times, however the material remains associated with this utilization are highly skewed in time and place. Probable factors behind this incomplete data base are revealed through late Quaternary stratigraphic and pedologic investigations. Lowland landforms are comprised primarily of fluvial and slopewash sediments of Holocene age. In the tributary valleys, as many as four depositional terraces are present, ranging in age from 7000 B.P. to present, while in the valley of the Little Missouri River, extant alluvial sediments were deposited almost entirely within the last 400 years. Upland landforms contain a much longer stratigraphic sequence, comprised primarily of eolian loess deposition going back over 12,000 years. The stratigraphic and pedologic data are integrated with stable carbon isotope analysis in a reconstruction of the late Quaternary geomorphic and paleoenvironmental history of the badlands region.
ACKNOWLEDGMENTS

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A very special thanks goes to my wife, Castle McLaughlin, for her encouragement and patience during the many long months of studying and writing. Castle, who also conducted research in Theodore Roosevelt National Park, shares my deep fondness for this special place and for the wild things that dwell within.

Finally, I dedicate this dissertation to my parents, Duane and Dorothy Kuehn, for their countless acts of kindness and support. Their faith in me will not be forgotten.
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CHAPTER I
INTRODUCTION

Geoarchaeology and Contemporary Archaeology

In the past several decades a number of significant changes have occurred in the way archaeologists approach the study of the human past. These have included the rise of neo-evolutionary and processual, neo-historical, neo-Marxist, behavioral, post-processual, contextual, and relativistic paradigms (Trigger 1989). Although often theoretically quite divergent, all of these approaches depend upon the use of inference in their development of hypotheses concerning past human behavior. The use of inference, of course, is rendered necessary by the fact that archaeology as a discipline cannot depend solely upon direct observation of human behavior in cultural systems that are still operational. Whether it is called "translating", "decoding", or "reading" the past (Binford 1983; Hodder 1986), inference is a necessary challenge and unifying theme in contemporary archaeology.

Of the various recent theoretical approaches to archaeological inference, few have proven as useful during these times of increased field research and federally-mandated cultural resource management, as have certain elements of the behavioral (Schiffer 1976) and contextual (Butzer 1978, 1980, 1982; Schoenwetter 1981) paradigms. In particular, these elements are transformation theory, or the study of site formation processes, (Schiffer 1987), and the subdiscipline of geoarchaeology (Butzer 1982; Davidson and Shackley 1976; Gladfelter 1977, 1981; Hassan 1979; Rapp and Gifford 1985; Waters 1992).

The usefulness of transformation theory stems from the fact that most archaeological inference is derived from the study of the material remains of past human activity (i.e., artifacts and ecofacts), and that these remain seldom retain a static relationship with the natural environment after they cease to be part of a behavioral system (Schiffer 1976, 1987; Trigger 1989:19). Therefore, unless one is dealing with rare and exceptionally well-preserved sites, most archaeologists have come to realize that meaningful inferences normally must be drawn after a thorough study of site formation processes (Trigger 1989:359-360).

This dissertation follows the style and format of American Antiquity.
While geoarchaeology, the application of geological concepts and principles to archaeological research, is an important constituent in the analysis of environmental, or natural, site formation processes (Butzer 1982; Gladfelter 1981; Schiffer 1987), it also contributes to archaeological inference in a number of other critical areas.

Geoarchaeology is normally regarded as a subdiscipline of archaeology that has emerged only recently in terms of legitimacy and widespread acceptability (Waters 1992:3). Yet, the discipline of archaeology itself began in the late nineteenth century with a strong dependence upon geological principles, especially as they were applied to the dating of sites and to the establishment of human antiquity in Europe (Daniel 1963; Gifford and Rapp 1985; Grayson 1983; Hassan 1979; Renfrew 1976; Trigger 1989). In the United States, geology played a very similar role, as evidenced by the impact that geologist Kirk Bryan (1929) had on archaeological thought by conclusively demonstrating the antiquity of the Folsom site in New Mexico (i.e., late Pleistocene).

Although the determination of temporal placement and site chronology remains one of the most important contributions of geoarchaeology, the subdiscipline has expanded to include all aspects of archaeological context. This expansion has made it a key element in the contextual archaeology paradigm, as espoused by Butzer (1978, 1980, 1982), Schoenwetter (1981), Stein and Farrand (1985), Waters (1992), and others. This is not to be confused with the contextual approach of Ian Hodder (1982), which stresses the role of social interaction, structuralism, and symbolism in the production of material culture, rather than exclusive reliance on factors of ecological adaptation and sociopolitical organization (Trigger 1989:348-356).

The contextual archaeology of Butzer and others is an ecosystems approach to the study of the human past that embraces space, scale, complexity, interaction, and stability as its central themes (Butzer 1982). It requires the input of not only geoarchaeology, but also zooarchaeology, archaeobotany, and archaeometry. Geoarchaeology, according to Butzer (1982:38), contains five principle "study components", these being: (1) landscape context, (2) stratigraphic context, (3) site formation, (4) site modification, and (5) landscape modification.

The five "study components" of Butzer are similar to the five major geoarchaeological "interests" elucidated by Gladfelter (1981:347). These are: (1) site location, (2) site formation processes, (3) cultural and natural post-occupational site disturbances, (4) temporal context, and (5) paleoenvironmental landscape reconstruction.
Somewhat earlier, Hassan (1979) listed nine principal topics in geoarchaeology. The list included many of the same research interests as those just mentioned, but also contained elements of archaeometry and archaeological resource conservation (Hassan 1979:268).

In a more recent and succinct melding of previous discussions on geoarchaeological method and theory, Waters (1992:7-13) argues that the principle research objectives of geoarchaeology are concerned with: (1) temporal context, (2) spatial context/site formation processes, and (3) landscape reconstruction. Like Butzer, Waters (1992) embraces a contextual and systemic approach to archaeology that stresses the identification of all components of the human ecosystem. Such an approach is, by necessity, highly interdisciplinary, involving input from the biological and geological sciences in addition to traditional archaeological methods of research.

While established themes in geoarchaeology have tended to center on questions of context (especially the temporal, spatial, and environmental contexts of sites and regions), contemporary geoarchaeology has contributed to other aspects of archaeological research as well. The description and explanation of past human behavior is (and has been) one of the principal goals of archaeology. According to Binford (1968) and Schoenwetter (1981:369), the other principal objectives of archaeology are "the description and explanation of ethnographic phenomena of the past (paleoethnography or reconstruction of lifeways); and ...the study of cultural process." In a theoretical sense, geoarchaeology has the potential to contribute to all three of these goals, although its strongest contribution has been to the understanding of past behavior via culture change and cultural ecology.

Butzer's (1976a, 1980) landmark research into the paleohydrology of the Nile River and its impact on pharaonic Egypt; his (1981) research into environmental degradation and the decline of the Axum civilization in Ethiopia; Jacobsen and Adams' (1958) thesis on the role of soil salinity and sedimentation in the breakup of complex societies in ancient Mesopotamia; and Waters (1988) research into the impact of fluvial processes on Hohokam settlement patterns, are but four examples of the contribution geoarchaeology can make to the archaeological understanding of culture change. Likewise, the relationship between human cultures and the biophysical environment is the central theme of Butzer's contextual archaeology paradigm: a paradigm which strives "...to arrive at a realistic appreciation of the environmental matrix and of its potential spatial, economic, and social interactions with the subsistence-settlement system" (Butzer 1982:12). This approach is remarkably similar to the anthropological paradigm of cultural ecology, a methodology defined as "the study of human populations, and their culturally patterned behavior, as components within ecosystems."
(Keesing 1976:552). It is ironic, however, that the pioneering figure in the anthropological study of both cultural ecology and culture change, Julian Steward (1955), is not cited in Butzer's most comprehensive work on the contextual paradigm. "Archaeology as Human Ecology" (1982), nor are some of the other more significant anthropological contributions to ecology (i.e., Jochim 1981; Rappaport 1968; Vayda 1969).

In addition to its emphasis on context and the reconstruction of human ecosystems, another recurrent theme in the geoarchaeological literature is the call for improved academic training (Gifford and Rapp 1985; Gladfelter 1981). For example, Gladfelter (1981:356-357) argues for changes in university curricula, with the objective of producing not specialists in each of the associated disciplines, but rather "...to instruct the archaeologist in the contribution that can be made to archaeology so that as a discipline it can better pursue the questions it raises."

A little over a decade later, Gladfelter's plea is beginning to become reality, and increasing numbers of archaeologists are receiving substantial training in the geosciences. It is only natural that as this trend continues, the subdiscipline of geoarchaeology will address more and more of the principal goals of contemporary archaeology, including the description of paleoethnography and the study of cultural process (Schoenwetter 1981:369). This increased focus, by necessity, cannot afford to overlook relevant concepts and principles of anthropology, the parent discipline of archaeology in the United States for over four decades (Schoenwetter 1981:369). As a result, most archaeologists with Ph.D.'s in the U.S. have their degrees in anthropology, and therefore have had extensive training in anthropological method and theory. It is for this reason that geoarchaeology will inevitably begin to adopt a perspective that does not exclude relevant concepts from the social sciences. Indeed, in a recent discussion on the goals and definition of geoarchaeology, Leach (1992:415) argues that geoarchaeology should "...develop a series of integrative models that build on the strengths of both geology and anthropology."

As an example, while contextual archaeology has a tendency to identify environmental or evolutionary factors as the catalysts behind culture change (cf. Butzer 1982), anthropological research indicates that such change can also occur as the result of ideational development through processes of innovation, invention, and culture contact (Rosman and Rubel 1992:264). In addition, recent anthropological approaches also recognize the possibility of holistic relationships between a wide array of cultural components and the environment, including such non-materialistic aspects as ritual and marriage patterns (Jochim 1981; Vayda and Rappaport 1968). Other paradigms like general systems theory, suggest that social and political variation can result from a number of non-
ecological factors like social unit relationships (Johnson 1978). In other words, anthropology recognizes that the natural environment "...is of course not the only source of influences upon cultural behavior" (Vayda 1969:xi).

In current archaeology, the various approaches to the study of culture change and its relationship to the environment stem, in no small way, from the very nature of archaeological data, which exhibit an ascending scale of difficulty as far as interpretation is concerned (Hawkes 1954). The easiest cultural elements to identify are technological, followed with increasing difficulty by economic, social, political, and finally, ideological (Hawkes 1954; Trigger 1989). The progressively higher levels contain more cultural features that cannot be explained by ecological factors alone, especially features such as ideology, religion, and social organization (Trigger 1989:394-395). Thus, paradigms which rely too heavily on ecological or evolutionary explanations tend to emphasize the external constraints on human behavior at the risk of overlooking other equally significant components of human culture. For this reason, the difficulty associated with the drawing of inference from archaeological data has been the source of much archaeological debate since at least the 1950's, and has been responsible for the development of multiple, and often radically divergent, archaeological paradigms. These include processual archaeology, which argues that these difficulties can be overcome by proper methodological advances (cf. Binford 1972), and the contextual approach of Butzer and others that appears to concede, albeit not explicitly, that the higher levels of culture are not available to archaeological understanding. It is ironic than an alternative paradigm under the same title, but with quite a different perspective, has also recently arisen. That is the contextual approach of Hodder (1982), which considers context to be all of the constituents of culture, including meaning, style, cosmology, religion, and ideology. Rather than strictly environmental, evolutionary, or materialistic approaches, Hodder and other post-processual archaeologists, draw on elements of structuralism, spatial analysis, Marxism, historicism, and cultural dichotomy (Hodder 1986).

While the basic tenets of Butzer's contextual archaeology have, in fact, formed an integral part of the following study, it would be a mistake for the contextual paradigm, and one of its principle constituents, geoarchaeology, to ignore relevant aspects of the other approaches to the study of human behavior. Indeed, Butzer himself (1982:12) states that "no one paradigm deserves to be enshrined as superlative; alternative viewpoints are essential to good scientific practice." It is in this spirit that I urge geoarchaeologists to consider, and attempt to incorporate when possible, some of the more appropriate concepts of anthropology and the other social sciences when drawing
inferences about past human behavior on the basis of geoarchaeological data. Such consideration will help insure that geoarchaeology remains relevant not only to future generations of archaeologists, but also to its place as a viable subdiscipline of archaeology.

Delineation of Project Research Goals and Study Area

The Project Research Goals

Few places are as well suited to the application of geoarchaeological and site formation principles as are the Little Missouri Badlands of western North Dakota, a highly dynamic landscape where natural impacts to the archaeological record have been profound (Figure 1). It is only recently that archaeologists in the area have become aware of the need to more fully understand these impacts before drawing inferences based solely on the study of extant material remains (Kuehn 1990, 1993). Such an understanding can best be facilitated by elucidating the stratigraphic and paleoenvironmental context of the badlands archaeological record. The identification of such a context comes none too soon, considering that, in spite of Colin Renfrew’s (1976:2) profound statement, "...every archaeological problem starts as a problem in geoarchaeology", hypotheses concerning prehistoric settlement and subsistence behavior in the badlands have been developed (and even published) for over a decade in lieu of an adequate understanding of site formation processes and landscape evolution. It is for this reason that the following study is intended to provide a starting point for the drawing of meaningful archaeological inference in the Little Missouri Badlands by systematically identifying some of the more significant non-cultural factors involved in the creation of the extant archaeological record. In other words, it will utilize an interdisciplinary and contextual approach in order to demonstrate that there are significant temporal and spatial voids in the badlands archaeological record that have been formed by natural geomorphic processes, and that these voids must be identified before true progress in the development of archaeological inference can be made. The primary research orientation therefore involves the elucidation of noncultural site formation processes at the regional level (Schiffer 1987), rather than a comprehensive discussion on precontact human behavior and paleoethnography in the badlands region. This lack of emphasis stems from: (1) the explicitly stated project research goals; (2) the limitations inherent in the known archaeological data base; and (3) the preliminary nature of archaeological research in the Little Missouri Badlands. The latter are deficiencies that hinder the drawing of meaningful inference about many aspects of past human behavior. Nevertheless, some of the data that will be discussed do have
relevance to questions of culture change and human/environmental relationships, and therefore will be addressed in rather modest detail.

This dissertation, however, is intended to accomplish more than just establish a starting point for future research. Although the principal goal is indeed evident in its title (i.e., to utilize geoarchaeological methods and principles to identify the late Quaternary stratigraphic and paleoenvironmental context of the archaeological record), it will also attempt to contribute to the question of site location in the badlands region. This additional archaeological objective will be addressed through the presentation of a series of predictive hypotheses (cf. Sebastian and Judge 1988:1) founded on the well-known principle that an archaeological site in primary context (i.e., a site that accurately reflects its systemic context), cannot be younger than the sedimentary matrix overlying it (Schiffer 1972; Stein 1987). In other words, it will be based on the spatial distribution of stratigraphic units and associated landforms of known age and archaeological composition within a well-defined and representative study area. These in turn will serve as analogues to equivalent units and landforms in the badlands as a whole. This is an increasingly popular approach to site location grounded on sound empirical geoarchaeological research (cf. Artz 1985; Bettis and Benn 1984; Gardner and Donahue 1985; Haynes 1982; Waters 1992).

Just as there is a need to better understand the archaeology of the badlands region, there is also a need to better understand the late Quaternary stratigraphy, paleoenvironment, and geologic history of this portion of the Northern Great Plains. It is therefore the principal non-archaeological objective of this dissertation to provide significant data of relevance to the earth sciences.

The Project Study Area

The badlands encompass a large region surrounding the Little Missouri River, from its entry into the state near the South Dakota border, to its confluence with the Missouri River in west-central North Dakota. Badland topography, therefore, covers over 5,400 km2. In addition, the current badland archaeological data base is comprised of hundreds of separate archaeological contract and research reports of varying size, scope, complexity, and rigor (cf. Kuehn and Gregg 1985:196-198). Many of these are outdated, inadequate, and inconsistent in their methodology, artifact identification, and description of sedimentary environments, landform setting, and raw material descriptions. These inconsistencies and deficiencies, coupled with the sheer volume of available reports, makes synthesis and utilization of the entire data base at face value logistically impractical and methodologically unsound. Therefore, the huge size of the badlands, and the inconsistent nature of
the archaeological data base, precluded any attempt at complete archaeological, stratigraphic, and paleoenvironmental data recovery from the region as a whole. Instead, it was decided to focus the study on a relatively small, but well defined area; one which had been at least partially geologically and geomorphologically investigated; and one with a consistently recorded archaeological data base. The only portion of the badlands which meets all of these criteria is the South Unit of Theodore Roosevelt National Park (THRO), a 19,000 ha federal preserve located in the heart of the badlands (Figures 1 and 2). The THRO South Unit is indeed well defined and contains representative portions of all major badland depositional environments and landform categories. It has also been the focus of previous geomorphological research, in fact the only substantive research into badland fluvial geomorphology and stratigraphy (Everitt 1968; Gonzalez 1987; Hamilton 1967). In addition, it was the object of a three-year archaeological survey conducted under the direction of the author (Kuehn 1989, 1990). For these reasons, this dissertation on the geoarchaeology of the Little Missouri Badlands will draw primarily on data recovered from the South Unit of Theodore Roosevelt National Park. Some portions of the study, particularly sections dealing with synthesis, interpretation, and site location, will however, utilize archaeological and stratigraphic data that were gathered during professional research in other portions of the badlands. Most of this research has focused on upland settings, although limited stratigraphic information collected from lowland settings in the southern badlands will also be included (Figure 1).
CHAPTER II
NATURAL AND CULTURAL BACKGROUND

Physiographic and Topographic Setting

The Little Missouri Badlands are located entirely within the Great Plains physiographic province as delineated by Fenneman (1931). The Great Plains province, a generally flat landscape created by the erosion of flat-lying sedimentary rocks, encompasses approximately 1.5 million square kilometers in the center of the North American continent. The Great Plains, in turn, have been subdivided into nine sections; these being: (1) the Missouri Plateau; (2) the Black Hills; (3) the High Plains; (4) the Plains Border; (5) the Colorado Piedmont; (6) the Raton section; (7) the Pecos Valley; (8) the Edwards Plateau; and (9) the Central Texas section (Pirkle and Yoho 1977). The Missouri Plateau section includes the western and southwestern two-thirds of North Dakota, the western three-fourths of South Dakota, all of Montana east of the Rocky Mountains, and the northeastern one-quarter of Wyoming.

In North Dakota, the Missouri Plateau is further divided into the glaciated and unglaciated subsections, and into the Coteau Slope, Missouri Slope Upland, McKenzie Upland, and Little Missouri Badland regions. The badlands are located primarily in the unglaciated subsection, although the maximum southern limit of Pleistocene glaciation bisects their eastern and northernmost margins (Figure 2). As will be shown, the badlands themselves began to form primarily as the result of Pleistocene glacial advances onto the Missouri Plateau.

The badlands region is described by Bluemle (1991:4) as a "rugged, deeply eroded, hilly area along the Little Missouri River; gentle slopes characterize 20 to 50 percent of the area and local relief is commonly over 500 feet." They were created by the downcutting of the Little Missouri River into the poorly cemented sediments of the Missouri Plateau and by other associated degradational processes. The Missouri Plateau sediments and their geologic history will be briefly described in the subsequent section. In any case, the dissection and topographic relief associated with the badlands increases along with the base level gradient of the Little Missouri (i.e., in a north/northeastern direction). In the southern portion of the badlands, the Little Missouri has downcut approximately 25 m into the preglacial Missouri Plateau surface, while in the northeastern portion, the depth of the downcutting is over 150 m (Bluemle 1977:11). In the South Unit of...
Figure 2. Portion of geologic map of North Dakota showing Little Missouri River and adjacent badlands (from Clayton 1980).
Theodore Roosevelt National Park, the Little Missouri base level is currently 90 m below the prebadland Pleistocene surface.

The badlands are comprised of a series of distinctive landforms, or recognizable surface features (Tuttle 1970). These will form an integral part in the subsequent discussions on archaeological site location and late Quaternary geologic history. The more prominent badland landforms include ridges, buttes, knobs, foothills, terraces, active floodplains, stream channels, gullies, and alluvial fans. The first three landform types occupy the higher elevations (above ca. 750 m asl) and are classified as uplands. They represent extant, but often segregated, portions of the Missouri Plateau, and are composed of Fort Union Group sediments generally overlain by a thin mantle of loess. Ridges are defined as long, narrow stretches of elevated ground (Goult 1991), with a length:width ratio of at least 2:1 (Kuehn 1990:24). Buttes are prominent, flat-topped remnants of elevated ground that generally exhibit a length:width ratio of less than 2:1 (Goult 1991; Kuehn 1990:24). Knobs are isolated, and generally small, rounded buttes or hills comprised primarily of eroded bedrock (Goult 1991). Terraces, floodplains, and stream channels are landforms of alluvial valleys that occupy the lower elevations (below ca. 730 m asl) and are classified as lowlands. Terraces have generally flat, horizontal surfaces (treads) that are frequently bounded at one end by a steep scarp, or riser. Floodplains are flat surfaces underlain by unconsolidated lateral and vertical accretion sediments. They are located adjacent to active stream channels and are subject to periodic flooding. Foothills have highly variable elevations, but are generally intermediate in elevation between the uplands and lowlands. They take the form of hills which lie at the base or foot of ridges, buttes, or other uplands, and usually remain articulated to the more elevated landforms.

In the study area, slopewash sediments can accumulate on sloping surfaces (sheet deposits), along valley margins, in the bottom of preformed valleys, and at the mouth of valleys (alluvial fans). Slopewash deposits are both incised and unincised by gullies. In valleys whose floors contain meandering streams, they frequently interfinger with fluvial sediments, especially terraces and active floodplains along the edges of the valley. Contrary to classic sedimentological models (cf. Boggs 1987; Bull 1972, 1977; Harvey 1989), alluvial fans in the badlands exhibit a wide variety of morphological characteristics, including traditional conical forms, but also forms that are more linear and amorphous. The latter are either bedrock controlled or are the result of post-depositional fluvial activity (i.e., aggradation and/or degradation). Fans are also located on the sides of valleys, where they can occur as either individual deposits, or as coalesced bajadas.
Gullies are ephemeral, entrenched channels with steep sides and steeply sloping or vertical head scarp’s (Brice 1966; Schumm and Hadley 1957). In the badlands, they are frequently incised into: (1) the slopewash fill of preformed valleys; (2) the Paleocene bedrock exposed on slopes or valley sides; and (3) the undissected upland surfaces where they encroach via the headward erosion of drainage systems (Campbell 1989).

As would be expected, the topography of the Missouri Plateau section adjacent to the badlands is generally much more subdued. For instance, the Missouri Slope Upland and McKenzie Upland regions consist primarily of rolling to hilly plains with gentle slopes comprising 50% to 80% of the area (Bluemle 1991:4). The Coteau Slope consists of rolling to hilly plains located east of the Missouri River, again with gentle slopes characterizing 50% to 80% of the region (Bluemle 1991:4).

All of the surrounding physiographic regions do contain isolated areas of high topographic relief in the form of buttes, and to a lesser extent, stream valleys, pediments, differential-erosion ridges, and blow-out depressions (Bluemle 1991). In the unglaciated Missouri Plateau subsection of southwestern North Dakota, buttes are far and away the most conspicuous topographic features. Numbering in the hundreds and varying greatly in height and surface area, the buttes all contain some form of erosion-resistant caprock (Bluemle 1991). The more prominent buttes visible from the THRO study area are Sentinel Butte, Square Butte, Bullion Butte, and Camels Hump Butte. The highest of these, Sentinel Butte, rises over 200 m above the surrounding Missouri Plateau surface.

Pre-Quaternary Geologic History

Most of the Missouri Plateau, including the study area, is situated in the Williston Basin, an intracratonic basin located southwest of the Canadian Shield. The basin, which also encompasses portions of northeastern Montana, northwestern South Dakota, southern Saskatchewan, and southwestern Manitoba, formed during the early Paleozoic (approximately 515 Ma) as the result of crustal weakening, subsidence, and subsequent marine flooding (Bluemle 1991; Carlson and Anderson 1965). The basin at its lowest point is over 4600 m deep and its floor is comprised of Precambrian rocks, some of which are dated by K-Ar and Rb-Sr methods at 1,550 to 1,740 Ma. These ages were obtained from well core samples taken within the badlands (Goldich et al. 1966, from Carlson 1983:8).

The Williston Basin was subsequently filled with marine, deltaic, fluvial, and lacustrine sediments that have been organized into six sedimentary sequences. These are, as illustrated in
Carlson (1983:7), the Sauk (Cambrian and early Ordovician), Tippecanoe (Ordovician and Silurian), Kaskaskia (Devonian and Mississippian), Absaroka (Pennsylvanian, Permian, and Triassic), Zuni (Jurassic, Cretaceous, and Tertiary) and Tejas (Tertiary). Each of the major sequences are separated by unconformities, some of which are temporally substantial (Bluemle 1991). In general terms, most of the sediment deposition in the Williston Basin occurred during repeated marine and lacustrine transgressions into the region, while fluvial processes contributed sediment and also initiated erosion during multiple periods of marine and lacustrine regression. Of the six principle sedimentary sequences, only a number of groups and formations associated with the Zuni and Tejas are visible on the surface of the Missouri Plateau in western North Dakota. These are the Pierre Formation, Fox Hills Formation, Hell Creek Formation, and Fort Union Group of the Zuni Sequence, and the White River Group of the Tejas Sequence (Carlson 1983).

The Pierre, Fox Hills, and Hell Creek Formations are visible only in the southern portion of the badlands, in Bowman and Slope Counties. The Pierre Formation (Late Cretaceous) has a maximum thickness of 670 m and consists of shale deposited during a transgression of the Western Interior Cretaceous sea. As the sea regressed around 70 Ma, the fine-grained clastics of the Fox Hills Formation were deposited. The Fox Hill has a maximum thickness of 120 m and consists primarily of sandstone, with lesser amounts of mudstone, siltstone, and silty shale. The Colgate Member at the top of the Fox Hill Formation is a sandstone unit representing the shoreline facies of the regressing sea (Carlson 1983:12; Cvanara 1976a).

The Late Cretaceous Hell Creek Formation is comprised of sandstone, siltstone, mudstone, and shales (carbonaceous and bentonitic). The Formation has a maximum thickness of 150 m, most of which was deposited by fluvial and deltaic processes in marginal-marine environments (Carlson 1983:13; Moore 1976). Dinosaur remains are common in the upper Hell Creek carbonaceous and bentonitic sediments (Carlson 1983:13).

By far the most widely exposed sediments in the Little Missouri Badlands are those associated with the Tertiary Fort Union Group. Formations comprising the Fort Union Group are (from oldest to youngest), the Ludlow, Cannonball, Slope, Bullion Creek, Sentinel Butte, and Golden Valley. The first three are associated with the only Tertiary marine transgression into the area, and represent both marine and marginal-marine depositional environments. As with the previously described Cretaceous formations, the Ludlow, Cannonball, and Slope Formations are visible only in the southern portions of the badlands. The Ludlow Formation has a maximum thickness of ca. 100 m and is comprised of siltsone, mudstone, sandstone, and lignite deposited in fluvial, lacustrine,
and paludal environments (Carlson 1983:17; Moore 1976). The Cannonball Formation represents a marine brackish-water facies which interfingers with the Ludlow and Slope Formations from the east (Cvancara 1976b). In the badland study area, only two Cannonball strata have been identified, both less than 6 m thick (Carlson 1983). The Slope Formation is comprised of sandstone, siltstone, clay, and lignite, with an upper "siliceous zone" (Carlson 1983; Clayton et al. 1977). It also represents deposition in fluvial and swampy coastal environments.

The Paleocene Bullion Creek and Sentinel Butte Formations comprise the bulk of exposed bedrock in the Little Missouri Badlands, especially in the central, northern, and eastern portions (Figure 2). Both formations share similar gross lithologies (clay, sandstone, siltstone, shale, lignite), depositional environments (marginai-marine marsh and fluvial), and approximate maximum thickness (160 m). There are nevertheless some noticeable differences. The Sentinel Butte Formation is sandier, darker and more gray in color, and contains less mollusk fossils but more petrified wood than the underlying Bullion Creek. These differences are credited to a gradual filling of the swamp and marsh environments by fluvial sands originating from the rising Rocky Mountains to the west (Blueme 1991; Carlson 1983; Jacob 1975). In terms of badland formation processes, these differences are significant, as demonstrated by Clayton and Tinker (1971) in a study of hillslope lowering in a small basin in the South Unit of THRO. There they found that slopewash on the Sentinel Butte Formation slopes was three times greater than slopewash on the Bullion Creek Formation due to significantly lower rates of infiltration and percolation in the Sentinel Butte sediments. These lower rates were attributed to a higher percentage of sand fraction and montmorillonite clay fraction in the Sentinel Butte Formation (Clayton and Tinker 1971).

By ca. 50 - 60 Ma, deposition in the Williston Basin was primarily fluvial and lacustrine in origin (Blueme 1991). The Late Paleocene-Early Eocene Golden Valley Formation, which unconformably overlies the Sentinel Butte, is comprised of up to 80 m of brightly colored clay, silty clay, and sand, although there are also erosion-resistant siliceous beds which form caprock in isolated upland areas (Carlson 1983; Hickey 1976).

Following the deposition of the Golden Valley Formation, lacustrine and fluvial deposition predominated in some portions of the Williston Basin, although other portions underwent substantial episodes of erosion (Blueme 1991). The Middle to Late Tertiary (Oligocene and Miocene) White River Group of the Tejas Sequence is comprised of conglomerates, siltstone, clay, freshwater limestone, tephra, and sandstone (Blueme 1991). The group contains two formations, the Chadron and Brule, each with a maximum thickness of about 30 m. The former is characterized by bentonic
silty clay and cobbly arkose, while the overlying Brule is characterized by poorly sorted clastics which range from mudstones to sandstones (Carlson 1983). The White River Group is noted for its high bentonitic clay/volcanic ash content and for the presence of indurated beds which form caprock on some of the more prominent buttes in the unglaciated portion of the Missouri Plateau. Additional Miocene and possibly Early Pliocene sediments which are somewhat lithologically dissimilar from the White River Group deposits have been recognized on the tops of some of the higher buttes in the region. Classification and the establishment of a chronology for these sediments has been rather inconsistent, however Denson and Gill (1965) assigned units of tuffaceous, coarse sandstone on some of the buttes to the Miocene Arikaree Formation, while Delimata (1975) assigned the ca. 120 m thick lacustrine unit on top of the Killdeer Mountains in Dunn County to the Killdeer Formation (Clayton et al. 1980).

By the end of the Tertiary period, ca. 5-3 Ma, the Pliocene Epoch saw the end of lacustrine deposition and the continuation of widespread erosion that, in most of the Missouri Plateau, began in Miocene time. This erosion has been attributed to planating stream action associated with major uplifting and the continued rise of the Rocky Mountains to the west (Bluemle 1991). Degradation associated with these broadly meandering streams was responsible for the removal of almost all of the Golden Valley and White River sediments from western North Dakota. The vertical extent of this erosion is evident today by the presence of Golden Valley and White River sediments on the top of isolated buttes that rise from 200 - 245 m above the present surface of the Missouri Plateau. With the exception of these buttes, by the start of the Quaternary Period (ca. 2.1 Ma), most of southwestern North Dakota was a broad rolling upland. Parts of this surface, termed the Missouri Slope by Moran et al. (1976), were covered with fluvial gravels analogous to the Flaxville Gravels of eastern Montana (Collier and Thom 1918). In the South Unit study area, this gravel-covered surface is extant on Petrified Forest Plateau, which is situated 150 m above the floor of the Little Missouri River. A second major period of erosion in southwestern North Dakota occurred after the Early Pleistocene Dunn, and perhaps Verone, Glaciations, when the land surface over much of the Missouri Slope between the Little Missouri Badlands and the Russian Springs Escarpment was lowered an additional 30 - 60 m (Moran et al. 1976). This extensive landscape is designated the Green River Stability Surface (Moran et al. 1976:146-147).
Modern Climate, Flora, and Fauna

Modern Climate

The modern climate in western North Dakota is semiarid continental, which is characterized by high daily, seasonal, and annual variability in precipitation and temperature (Jenson 1972). Precipitation is predominately seasonal, but often erratic. At Medora, North Dakota, on the south edge of the South Unit study area, the average annual precipitation is 15.2 inches (38.6 cm). Most precipitation (77%) occurs between May and September, which coincides with the average growing season of 124-130 days. Of this, 50% falls during short, often violent, thunderstorms between May and July (East et al. 1985). Drought hits the region about every 20 years, although dry years occur more often.

The area has an average annual snowfall of 30 inches, which is the lowest of any state where winter temperatures average below freezing (Wilkins and Wilkins 1977:14). The mean annual temperature in the study area (and North Dakota in general) is 40° F., which is the second lowest in the United States after Alaska (Wilkins and Wilkins 1977:8). Temperature extremes, often within a short time period, are common. For instance, record temperatures in the state of 121° F. and -60° F. were both recorded in 1936 (Wilkins and Wilkins 1977).

Wind is a persistent aspect of the modern climate that contributes to high evaporation rates in the summer and high wind-chill factors and drifting snow in the winters. The prevailing wind direction is west-northwest, although in the summer months southerly winds are common. Average daily wind speed is 10 mph but winds up to 70 mph have been reported and winds of 20 - 30 mph are not unusual (East et al. 1985).

Modern Flora

As part of the Missouri Plateau section of the Great Plains province, the predominant vegetation is mixed grass prairie. In southwestern North Dakota, Whitman (1979) has identified 10 major grassland types or associations on the basis of soil conditions, topography, and species composition. Nearly all of the type areas contain a variety of grass and sedge species, including blue grama (Bouteloua gracilis), western wheatgrass (Agropyron smithii), needle-and-thread (Stipa comata), little bluestem (Andropogon scoparius), and needleleaf sedge (Carex eglecharis). The topographic and elevational diversity of the Little Missouri Badlands has resulted in a diversity of natural ecosystems which are not found together elsewhere on the Missouri Plateau (U.S.D.A. Forest
Service 1974). In the badland portions of the Little Missouri National Grasslands, a ca. 2 million acre multiple-use area administered by the Custer National Forest (Figure 1), the Forest Service has identified nine separate ecosystems. These are characterized by distinctive assemblages of topographic, pedologic, and floral/faunal components. The nine badland ecosystems are river bottoms, hardwood draws, toe slopes, terraces, upland breaks, river breaks, upland grasslands, rolling grasslands, and hilly scoria (U.S.D.A. Forest Service 1974).

Upland landforms are generally covered with mixed grass and sedge species, while lowland landforms contain green ash (Fraxinus lanceolata), cottonwood (Populus deltoides), chokecherry (Prunus virginiana), wild plum (Prunus americana), buffaloberry (Shepherdia argentea), sagebrush (Artemisia sp.), and other woody shrubs. The ubiquitous badland hillslopes contain Rocky Mountain juniper (Juniperus sp.), creeping juniper (Juniperus horizontalis), soapweed yucca (Yucca glauca), and prickly pear cactus (Opuntia sp.). The density and composition of these species are greatly influenced by hillslope aspect. North-facing slopes receive less solar radiation and are therefore cooler and more moist, supporting dense stands of juniper and ash. South-facing slopes receive substantially more solar radiation, and are therefore warmer and dryer, often devoid of vegetation save occasional yucca and prickly pear (cf. Clayton and Tinker 1971).

Modern Fauna

The faunal composition of western North Dakota is typical of the northern Great Plains in general. For example, Stewart and Stewart (1973) have identified 252 species of wildlife in southwestern North Dakota, all of which can be found throughout the Great Plains province. The once dense populations of native fauna that populated the badlands regions are still present to a large extent in the protected environs of Theodore Roosevelt National Park. Common mammals include mule deer (Odocoileus hemionus), white-tailed deer (Odocoileus virginianus), pronghorn antelope (Antilocapra americana), coyote (Canus latrans), fox (Vulpes sp.), porcupine (Erethizon dorsatum), badger (Taxidea taxus), jackrabbit (Lepus sp.), cottontail rabbit (Sylvilagus sp.), and prairie dog (Cynomys ludovici). The major predator species, bear, wolves, and mountain lions, have been hunted to extinction in this region, with the exception of coyote, fox, and bobcat (Lynx sp.).

The South Unit study area also contains some indigenous, but reintroduced species, including bison (Bison bison), elk (Cervus canadensis), and bighorn sheep (Ovis sp.). The South Unit also contains several bands of wild horses (Equus sp.) which are remnants of feral herds dating from the 19th century.
Prehistoric Cultural History

The study area has a long and potentially complex history of human occupation, although the earlier time periods are poorly understood. In general terms, what is known suggests that the badlands were conducive to prehistoric utilization because of an abundance and diversity of natural resources. In addition, the existing archaeological data base indicates that chronologies previously established for the Northern Plains in general, and the Northwestern Plains in particular, are applicable to the Little Missouri Badlands region. For this reason, the following discussion will draw on the prehistoric cultural chronologies presented in Frison (1978), Lehner (1971), Mulloy (1958), Reeves (1970), and others, as summarized for central and western North Dakota by Gregg (1985a), and applied to the THRO study area by Kuehn (1990).

Paleoindian Tradition, ca. 11,500 - 7500 yr. B.P.

In THRO, little is known about the lifeway of Paleoindian groups because of a paucity of recovered material remains. In the lack of hard evidence, conclusions about the lifeways of these people are based on archaeological research conducted in surrounding areas. Nevertheless, some data from the badlands are available. Projectile point styles suggest that Paleoindian groups were in the Little Missouri Badlands proper by ca. 10,000 yr. B.P., as indicated by the recovery of two Agate Basin projectile points (Beckes and Keyser 1983:173; Kuehn 1990; Ludwickson 1981). Agate Basin points are diagnostic of the Hell Gap-Agate Basin Complex, which typically dates to ca. 10,500 - 10,000 B.P. (Agogino 1961; Gregg 1985a). More recently, even older Folsom projectiles (ca. 11,000 - 10,000 B.P.) have been reported from upland areas peripheral to the southern portion of the badlands (Blikre and Borchert 1992), and from the Missouri Plateau west of the badlands (Beckes and Keyser 1983:173).

Other paleoindian projectile points reported from the badlands include Scottsbluff/Cody complex (Hill 1988), and parallel oblique flaked (Beckes and Keyser 1983:173; Borchert and Loendorf 1986). The Cody complex dates from ca. 9950 - 8500 B.P. (Agenbroad 1978; Gregg 1985a:93), while the parallel obliqted flaked complex dates from ca. 8950 - 7500 B.P. (Gregg 1985a:96-98).

The long temporal extent of the Paleoindian tradition witnessed the gradual warming associated with the transition from the late Pleistocene glacial episodes to the Holocene interglacial (cf. Wendland 1978). This transition produced substantial changes in the floral and faunal make-up
of the Great Plains (cf. Baker 1984; Semken 1984). These changes are reflected in a basic subsistence shift among Paleoindian groups from hunting/gathering with an emphasis on hunting of now extinct megafauna, to hunting/gathering with increased emphasis on gathering and hunting of modern (i.e., non-extinct) species (cf. Agogino 1962; Haynes 1966). This basic shift, from the "classic" Paleoindian to an essentially Archaic lifeway, occurred slowly over a period of centuries if not millennia (Gregg 1985a:83).

Plains Archaic Tradition, ca. 7500 - 2000 yr. B.P.

The pattern of increased resource diversity established by late Paleoindian times continued throughout the early and middle Holocene and was concomitant with a basic climatic pattern of increased aridity and seasonality. With the final extinction of Pleistocene fauna, the Plains Archaic lifeway was firmly established by ca. 7500 B.P. This lifeway, characterized by hunting and gathering of modern flora and fauna within a Plains grassland biome (Gregg 1985a:100), persisted for many thousands of years. Indeed, in some portions of the Great Plains, particularly the Northern Plains, the Archaic lifeway remained essentially unchanged until Historic times (cf. Reeves 1969:37, from Gregg 1985a:100). This incredible stability in subsistence and settlement was made possible by the persistent but fluctuating presence of the American bison (Bison bison). So pervasive was bison hunting to the subsistence patterns of prehistoric groups in the Great Plains that "...significant technological and economic changes associated with transformations of the socio-ideological components of a social system..." (Michlovic 1986:209) are difficult to discern from archaeological evidence. Nevertheless, variation in projectile point form and settlement/subsistence patterns (often in conjunction with climatic changes) are sufficiently recognized to allow for the subdivision of the Plains Archaic tradition into several complexes. These are the Logan Creek/Mummy Cave complex, the McKean complex, and the Pelican Lake complex. These are temporally equivalent to the Early, Middle, and Late Archaic in other portions of North America (Gregg 1985a).

The Logan Creek/Mummy Cave complex in the Northwestern Plains dates to ca. 7500 - 5250 B.P. and is characterized by Simonsen side-notched projectile points (Gregg 1985a:101). While Logan Creek/Mummy Cave settlement and subsistence continued to be distinctively Archaic in nature, Greiser (1985:40-42) has hypothesized that the increased aridity and temperature of the middle Holocene Altithean created diminished resource availability and fluctuations in the once stable bison populations, which in turn, led to decreased task group size, and increases in territorial size and in the numbers of exploited plant and animal species (Greiser 1985:40; Reeves 1973).
These hypothetical changes in behavior have yet to be demonstrated in western North Dakota, perhaps due to the dearth of available archaeological data (cf. Gregg 1985a; Kuehn 1990; Root et al. 1985). Indeed, there are currently less than 10 sites in the region with documented Logan Creek/Mummy Cave components.

The number of extant archaeological sites in the Northwestern Plains post-dating 5000 - 4500 B.P. increase dramatically from those associated with earlier time periods. Consequently, the earliest well documented cultural group in the Little Missouri Badlands region is the McKean Complex. McKean Complex sites date from ca. 4500 - 3000 B.P. and are characterized by a great deal of variation in projectile points, with predominant forms including McKean lanceolate, Duncan, Yankee, Mallory, and Hanna (Gregg 1985a; Reeves 1970). While poorly documented in the badlands study region, Keyser and Davis (1984) and Syms (1969) propose that McKean settlement in the Northern Plains consisted of a centrally based circulating pattern which utilized base camps, field camps, locations, and stations. Greiser (1985:115) states that improved climatic conditions during McKean times led to increased carrying capacity, an increase in the number of desirable site locations, and a decrease in territorial size. In extreme western North Dakota, there have been at least 45 reported McKean sites (Hill 1988; Kuehn 1990).

The Late Archaic Pelican Lake Complex (ca. 3500 - 1700 B.P.) is characterized by corner-notched projectile points, and a continuation of the Plains Archaic lifeway. Hypotheses concerning Pelican Lake settlement and subsistence patterns are inconsistent, however Kuehn (1990) suggests that they may have remained essentially unchanged from the Middle Archaic.

Plains Woodland Tradition, ca. 2050 - 1200 yr. B.P.

The long-standing Archaic subsistence strategy continued beyond the suggested terminal date of the tradition, however technological and ceremonial changes are evident in the archaeological record of the Northwestern Plains after ca. 2050 B.P. These changes are associated with the Plains Woodland tradition, which has been characterized as essentially Plains Archaic with the advent of ceramic technology and burial mound ceremonialism (Johnson and Wood 1980:38 from Gregg 1985a:117). Of the six Plains Woodland complexes identified for central and western North Dakota, only the Besant Complex has been well documented in the badlands region.

Besant is a Plains group characterized by large side-notched dart points (Besant), small side-notched arrow points (Samantha), and the earliest known use of ceramics in the Northwestern Plains (Gregg 1985a:118; Johnson 1977). Besant settlement appears to have been similar to Plains Archaic
groups, with a centrally-based circulating pattern that included field camps often represented by tipi rings (Gregg 1985a; Reeves 1970; Schneider 1982). Besant sites are commonly found throughout the Little Missouri Badlands, however tipi rings, common elsewhere, are rare in this region (Kuehn 1990).

Plains Village Tradition, ca. 900 - 150 yr. B.P.

Sites attributed to the Plains Village tradition are concentrated along the valley of the Missouri River, where settlement centered on semipermanent summer and winter earthlodge villages (Lehner 1971). Plains Village subsistence was evenly divided between horticulture (corns, beans, squash) and hunting/gathering. While horticulture was practiced on the terraces and floodplain of the Missouri River, hunting and gathering forays, particularly hunting, were conducted along major tributaries such as the Knife, Heart, Cannonball, and Little Missouri Rivers (Bowers 1948; Hanson and Gregg 1983; Wilson 1928). Consequently, the Little Missouri Badlands are considered secondary or tertiary components of the Plains Village territory, according to Syms' (1977) "co-influence sphere" model (Kuehn 1990:31). Plains Village sites, distinctive due to the presence of simple stamped, check stamped, and cord impressed ceramics, and Plains side-notched arrow points, have been reported from the badlands by various researchers (cf. Kuehn 1990:31). The sites are, for the most part, field camps and resource procurement areas, although extended occupation is documented at the Jacobsen site (32DU1) near the eastern edge of the badlands (Wood 1980), and from two sites along the Little Missouri River south of Medora (Johnson 1983; Metcalf and Schweigert 1987).

As a distinctive lifeway, the Plains Village tradition persisted into the Historic period. At the time of first contact with Anglo-Americans (ca. 1737), the tradition was represented by the Mandan, Hidatsa, and Arikara tribes. The traditional Plains Village way of life, however, was severely disrupted by a series of smallpox epidemics in the late 18th and early to mid 19th centuries. These epidemics greatly reduced Plains Village populations, and together with subsequent warfare with the Sioux, led to a coalescence of the once separate tribes into a single village, Like-a-Fishhook, in A.D. 1861. This is the terminal date given to the tradition by Lovick and Ahler (1982:76-77).
CHAPTER III

METHODOLOGY

Archaeological Field and Laboratory Methods

The archaeological data to be discussed in Chapter IV were drawn from a multi-year archaeological survey of Theodore Roosevelt National Park conducted by the University of North Dakota under the direction of the author. The survey, undertaken between 1987 - 1989, involved the intensive examination of 49 individual tracts of land, 34 of which are located in the THRO South Unit (Figure 3). The latter survey areas, with a combined areal extent of 12,100 acres (4840 ha), encompass virtually every depositional environment, landform category, and ecosystem in the Little Missouri Badlands.

The survey was a Class III intensive cultural resource inventory involving the author and from two to five archaeological assistants. As described in Kueln (1990), several different survey techniques were utilized because of the high degree of topographic variability. In areas of rugged terrain, the survey was topographically oriented; that is, the individual investigators examined specific landforms rather than follow a rigid pattern of survey transects. Special attention was given to areas offering good surface visibility, such as buffalo wallows, cutbanks, deflated areas, trails, or eroding landform margins.

In areas of flatter terrain, such as ridgetops, a more standardized series of parallel transects were utilized. These transects were generally 15 m apart and were walked back and forth across the survey area. Ridgetops were surveyed by first walking a series of transects along the edge and sides of the ridge, with special attention paid to the identification of buried soils. Subsequent parallel transects were then walked across the flat areas away from the ridge edges.

In drainages, the survey team spread out and walked down the long axis of the stream. One or two surveyors would walk along the edge of the channel and modern floodplain, while the others would walk along the terraces or edges of the valley.

Sites were recorded by collecting information on site size, site morphology, surface artifact content (including a complete inventory of all surface visible artifacts), site flora, ground surface visibility, view, soils, slope, landform and depositional environment, and distance to nearest water.
Figure 3. Map of Theodore Roosevelt National Park, South Unit, showing 1987-1989 survey areas.
Data were also collected on flaking debris technology, core reduction and cobble-testing techniques, stages of bifacial manufacture, and lithic raw material type (cf. Kuehn 1989:10).

A scale map of each site was prepared using metric tapes and compass bearings. The maps illustrated site boundaries, cultural features, artifact concentrations, diagnostic artifact locations, and topographic landmarks.

Artifact collection was limited to temporally and/or culturally diagnostic specimens such as projectile points or ceramics. Laboratory work included the washing and labeling of all collected artifacts, and specific lithic and ceramic analyses. Projectile points were measured (length, width, thickness in mm), and assigned to one of the technological, functional, and use-phase categories described in Ahler (1975) and Ahler et al. (1977). Information was also collected on raw material type and degree of patination or carbonate encrustation.

All collected specimens were identified, if possible, according to accepted Northwestern Plains projectile point typologies (cf. Frison 1978; Mulloy 1954; Johnson 1970; Wettlaufer 1960). Ceramics were measured (thickness in mm), and were described according to form, color, (exterior, interior, and paste), temper, surface treatment, and decorative techniques. Again, sherds were identified as to specific ware or general cultural affiliation (cf. Johnson 1980; Lehmer 1971: Lovick and Ahler 1982).

Stratigraphic Field Methods

Basic Goals and Principles of the Field Research

A stratigraphic framework for the archaeological record was established by identifying, describing, and correlating major late Quaternary stratigraphic units in the THRO South Unit, and placing them within a temporal context through the procurement of radiocarbon age determinations. The stratigraphic investigations centered on eolian, fluvial, and slopewash/alluvial fan sediments because they are far and away the most prevalent deposits in the study area.

The geological component of the fieldwork had three primary tasks. The first was to determine the various agents of late Quaternary sediment deposition and their corresponding depositional environments. The second was to determine the sequence of sediment deposition, erosion, and landscape stability at each of 35 primary study sections. The third objective was to determine lithologic and temporal equivalency among the major stratigraphic units.
The first task was accomplished by examining the lithology and sedimentary structures evident at each of the study sections. Eolian loess deposits were identified on the basis of grain size (i.e., silt), moderately well to well sorted massive bedding, homogeneous overall appearance, and to a lesser extent, landform setting (cf. Pesci 1968; Ritter 1978:338-340). Fluvial deposits were identified on the basis of generally fining upward sequences of channel lag deposits (poorly sorted sand and gravels), point bar deposits (trough cross-beded sands grading upward to trough cross-beded and ripple cross-laminated sand, and planar bedded silt and clay) nearbank and overbank sediments (massive to planar bedded sand, silt, and clay) (Allen 1970; Brakenridge 1987; Walker and Cant 1984). The slopewash/alluvial fan deposits were more difficult to recognize, partly because they often interfinger with fluvial deposits, especially along valley margins. Although highly variable, the sediments in general resemble the relatively coarse-grained sheetflood deposits which commonly drape alluvial fans in the arid Southwest (cf. Harvey 1989). Individual slopewash beds that interfinger with fluvial sediments generally consist of massive, poorly sorted sand and gravels, while thick slopewash/alluvial fan deposits (i.e., valley bottom and valley mouth) are more complex, consisting of massive, poorly sorted, sand interbedded with planar medium to thin beds of poorly sorted sand, gravels, and redeposited Fort Union Group clay. Other fans are relatively fine grained throughout, with thick, massive mud beds. Contrary to the "classic fan" literature (Bull 1972, 1977; Harvey 1989), little difference is evident in the textural/structural composition of the proximal, medial, and distal portions of fans. An important distinction between the slopewash/fan and fluvial deposits is the tendency of the former to have sloping treads while the fluvial floodplain and terrace treads are generally flat.

Determining the sequence of deposition, stability, and erosion at each of the study sections was a matter of: (1) identifying major stratigraphic units as indicators of distinct periods of deposition; (2) studying the boundaries or contacts between units for evidence of erosion; and (3) identifying buried soils as evidence for periods of landscape stability (cf. Birkeland 1984; Waters 1992:60-61).

The third task, that of correlating the major stratigraphic units within the South Unit study area, proved extremely difficult because of the dynamic and spatially varied nature of badland geomorphic processes (cf. Bryan and Yair 1982). In the loess-mantled uplands for example, a late Pleistocene unit identified at one section may not be present at a nearby section, although the latter may contain two early to middle Holocene units that are not present at the former. Similarly, in the lowlands, one section may contain a total of six buried soils; a second location only four; and a third
none. These differences appear to be the result of differential post-depositional erosion on a massive scale, and to the widespread, but variable, deposition of slopewash deposits. Likewise, tread elevational criteria in the correlation of stream terraces proved difficult because some terraces along the valley margins are capped by slopewash sediments, while an equivalent terrace in the middle of the valley may be totally unaffected by slopewash deposition. The problem of demonstrating lithological equivalency was further complicated by the presence of fan deposits at the mouth of virtually every tributary stream. These deposits are extant at different heights above the modern stream channels, and they also overly, are inset by, and interfinger with, terrace deposits. In addition, the fan sediments are difficult to distinguish from fluvial sediments because of the homogeneous nature of the reworked parent material (i.e., generally pale gray Paleocene Fort Union Group clay, sandstone, and shale).

Despite these difficulties, correlation among eolian and fluvial stratigraphic units was possible on a broad temporal scale (i.e., chronocorrelation) because of basic stratigraphic similarities and the procurement of radiocarbon assays. In the uplands for instance, it was possible to correlate stratigraphic units with principal members of the eolian lithofacies of the Oahe Formation, as defined by Clayton et al. (1976) to include all late Quaternary eolian silt deposited on gently sloping upland surfaces in North Dakota. The Oahe Formation members are (from Clayton et al. 1976): the Mallard Island (ca. 14,000 - 13,000 yr B.P.), Aggie Brown (ca. 13,000 - 8500 yr B.P.), Pick City (ca. 8500 - 4500 yr B.P.), and Riverdale (ca. 4500 B.P. - present).

The correlation of individual stratigraphic units or specific buried soils from one section to the other was only marginally successful. In the fluvial lowlands, correlation was limited to the demonstration of temporal equivalency among individual terraces within and between drainage systems.

Correlations of any kind would have been difficult if not impossible were it not for the procurement of radiocarbon age determinations from a number of key study sections. These allowed for the establishment of a stratigraphic chronological sequence for the study area.

Specific Methods

As illustrated in Figure 4, the stratigraphic data were collected from 35 individual sections within the THRO South Unit. These were selected because they best represented the three principal late Quaternary depositional environments (i.e., eolian, fluvial, slopewash/alluvial fan). The specific methods involved in their selection and analysis are as follows.
The edge of every major upland landform in the South Unit was examined for the presence of temporally diagnostic buried soils, especially: (1) the late Pleistocene/early Holocene Leonard Paleosol, defined by Clayton et al. (1976) as the upper submember of the Aggie Brown Member of the Oahe Formation; and (2) middle Holocene soils distinctive because of heavy accumulations of calcium carbonate (so-called "Altithermal soils"). Detailed stratigraphic investigations were conducted at every location where these soils were identified (n = 6). There is little doubt that additional locations containing both soils exist in the study area, especially in areas away from the eroding upland margins, however these could not be identified by pedestrian survey, and the use of heavy equipment to expose them is both prohibited and impractical in a National Park.

A number of upland areas without diagnostic soils were also carefully examined in order to better understand the age of the eolian deposits and the extent of post-depositional erosion. At least one stratigraphic study section was examined at each of the major upland landforms in the South Unit (i.e., Big Plateau, Petrified Forest Plateau, Johnson Plateau, Radio Tower Plateau, Peck Hill, Boicourt Ridge, and the Little Missouri Escarpment).

Pedestrian surveys were conducted along the Little Missouri River and every major perennial, intermittent, and ephemeral tributary in the study area. These included Knutson Creek, Paddock Creek, Jones Creek, Sheep Creek, Jules Creek, Beef Corral Wash, and Petrified Forest Wash. Numerous other unnamed ephemeral drainages, usually in the form of gullies, were also surveyed. The investigations included: (1) determining the number of floodplain surfaces (active and abandoned) extant within each drainage valley; (2) searching the vertical scarp of each floodplain exposure for the presence of materials suitable for radiocarbon dating; (3) making detailed stratigraphic descriptions at a representative number of those locations containing datable materials. Exposures with datable materials that were in close proximity to recognizable lignite coal deposits, especially those lying beneath the lignite, were not selected for collection as ¹⁴C samples because of the likelihood of contamination from older carbon. Likewise, datable materials in a disturbed context, such as krotovina, pipes, slumped areas, or areas affected by human activity (e.g., road construction, culvert placement, etc.) were also not selected for sampling. Finally, although carefully walked, Jones Creek and Paddock Creek were not subjected to intensive stratigraphic investigation at large numbers of specific sections because both stream systems were the object of previous geomorphological research efforts. Jones Creek was investigated in 1967 by T.M. Hamilton of the University of North Dakota, and Paddock Creek was investigated in 1986 by M.A.
Gonzalez of the University of Wisconsin-Madison. Nevertheless, terrace correlation in Paddock Creek was undertaken, alluvial fans were identified and described, and three formal stratigraphic sections were examined and dated in an attempt to strengthen the local terrace chronology.

The slawash/fan deposits were not as systematically examined, in part because they are difficult to recognize, but more importantly, because they are ubiquitous throughout the study area. Nevertheless, alluvial fans and other slawash facies in the Knutsen Creek, Paddock Creek, and Little Missouri River valleys were identified and described. Several of the fans were subjected to more intensive examination, including radiocarbon dating. In all of the other major drainages, fans, and (to a lesser extent) other types of slawash deposits, were mapped and described in terms of facies relationships with fluvial terraces.

At each of the 35 study sections the investigations included: (1) the identification of major stratigraphic units; (2) textural description using the nomenclature outlined in Folk (1954); (3) determination of sediment and soil color, both wet and dry, using Munsell Color Charts; (4) description of primary sedimentary structures (cf. Boggs 1987:Table 6.1); (5) identification of depositional environments and facies relationships; (6) description of grain sorting, reaction with HCL, and inclusions; (7) measurement of unit thickness and thickness of the entire deposit; (8) determination of the height of floodplain, terrace and alluvial fan terraces above current channel levels; (9) description of unit contacts (conformable, unconformable, contact distinctiveness, and topography); (10) description of all buried and surface soils; and (11) collection of sediment samples from major stratigraphic units for subsequent laboratory analysis.

Soil descriptions in the field followed the conventions of the Soil Survey Staff (1951) in terms of texture, soil structure, consistency, color, mottling, reaction, and special features. Soil horizon nomenclature followed the Soil Survey Staff (1990), with master and subordinate horizons identified in the field, and described in terms of depth, thickness, and boundary characteristics. Diagnostic surface and subsurface soil horizons at the two most extensively studied eolian sections (Locality O--Petrified Forest Ridge and Locality A--Big Plateau) were also identified on the basis of laboratory analysis. Consequently, when possible, individual soils at these two locations were taxonomically identified as to soil order, suborder, great group, and subgroup (Soil Survey Staff 1990).
Sedimentary, Palynological, and Isotopic Laboratory Methods

Samples collected from Localities A and O underwent extensive laboratory analysis at the Texas A&M Soil Characterization and Stable Carbon Isotope Laboratories, as did samples taken from late Pleistocene/early Holocene soils identified at Localities L and C. These analyses were conducted in order to provide quantifiable data on late Quaternary paleoenvironmental conditions, geologic landscape history, and diagnostic soil horizons. In addition, the samples from Localities O, L, and C were also analyzed for fossil pollen at the Texas A&M Palynology Laboratory.

At the Soil Characterization Laboratory, the particle-size distribution (PSD), and total carbon content (both organic and inorganic) of 66 samples were analyzed. These consisted of 35 constant volume samples collected from Locality A at regular 10 cm intervals, 16 constant volume samples collected from Locality O at regular 20 cm intervals, 13 samples collected from each of the identified stratigraphic units at Locality O, and one sample from the buried A horizons encountered at Localities L and C. The constant volume samples from Localities A and O were taken from the surface to Paleocene bedrock along a straight column down the eastern side of the profiles. During the collection process, care was taken not to combine sediments from different stratigraphic units.

Particle-size distribution was determined by utilizing the pipette settling method for the clay and silt size fractions (i.e., less than 0.05 mm in diameter or greater than 4.25 Phi) following Stokes Law (cf. Boggs 1987:108-109; Folk 1980) and the Soil Survey Staff (1975). The distribution of sand-sized particles (i.e., those greater than 0.05mm in diameter) was determined by the sieving method, which involved drying, sieving, and weighing the fraction that remained after the pipette analysis of the silts and clays (cf. Carver 1971; Holliday and Stein 1989).

Total carbon content was established through dry combustion, a technique that utilizes a resistance furnace, or total carbon analyzer, which measures the decomposition of carbonate minerals and the amount of liberated CO₂ (Nelson and Sommers 1982; Holliday and Stein 1989). Total inorganic carbon, expressed as CaCO₃ equivalent, was determined through the use of the gaso-volumetric Chittick apparatus, which measures the volume of CO₂ gas released through the reaction of carbonates in the sample with HCL (Machette 1986, from Holliday and Stein 1989:349-350). The percentage of organic carbon in the sample was then calculated on the basis of the difference between total carbon and total inorganic carbon (Holliday and Stein 1989:349).

The particle size distribution data are presented both on a clay-free basis (with mean PSD expressed in Phi units and in mm), and in terms of total sand, silt, and clay percentages expressed
in mm and according to the Soil Survey Staff (1951) textural classes. Organic carbon is presented by percentage, as are carbonates, in the form of calcite, dolomite, and CaCO₃ equivalent.

At the Stable Carbon Isotope Laboratory, analyses were conducted on the 35 constant volume samples collected from Locality A, on samples collected from each of the stratigraphic units identified at Locality O, and from samples collected from the buried late Pleistocene/early Holocene soils at Localities L and C. The analyses included pH, percentage of total carbon (both percentage of total organic carbon and percentage of total inorganic carbon), percentage of total nitrogen, and stable carbon isotope (δ¹³C) composition. Total carbon and nitrogen were determined through the use of a Carlo-Erba NA-15000 elemental analyzer configured for total C and N. The samples were first weighed, then acidified with a 1 normal solution of HCL to remove carbonates. They were then dried, re-weighed, run through the elemental analyzer, and placed in a second HCL 1 normal solution to drive off the remaining inorganic carbon. After this, the samples were dried and the percentage of total carbon was determined by weighing the remaining fraction. Sediment pH was determined through the use of a Beckman Expandomatic IV, using a 50/50 mixture of sediment and distilled water which was allowed to stand for 10 minutes before the reading was taken.

The stable carbon isotopic composition of organic carbon in each of the 50 samples collected from Localities A, O, C, and L, was determined through dry combustion of soil organic matter into CO₂ at 850 degrees C in tightly sealed quartz tubes containing CuO and Cu. The isotopic composition of the resultant CO₂ was determined using a VG-903 triple collector isotope ratio mass spectrometer (Boutton et al. 1993:5).

The pollen analysis was performed on 13 samples collected from Locality O (one from each of the identifiable stratigraphic units), and from the two buried A horizons at Localities L and C. The processed samples consisted of 20 ml of loose eolian sediment, to which were added Lycopodium tracer spores (two tablets of approximately 11,300 spores each). The samples were then treated with 36% HCL to remove carbonates. After reaction ceased, the samples were screened through 200 micron mesh, swirled, and allowed to settle for 2.5 hours. Remaining silicates were then removed through treatment with 70% hydrofluoric acid. After the HF treatment, the samples were diluted with approximately 600 ml of H₂O and allowed to settle for 3 hours. Following aspiration, refilling, and repeated settling, the samples were centrifuged and mixed no less than three times. Extraneous plant material was then removed using the acetylation method, which utilized 99.3% glacial acetic acid (to dehydrate) and a 9:1 hot mixture of acetic anhydride and sulfuric acid. Following acetylation, the remaining silt fraction was removed by heavy-density separation utilizing
zinc bromide. After separation, the remaining light fraction was dehydrated with ethanol and the residue was placed in small glass vials along with a small amount of glycerin jelly. The residue was then mounted onto slides corresponding to each of the 15 samples. The slides were examined with the aid of a light microscope at 400x magnification. Subsequent identification was conducted with the aid of the Texas A&M pollen reference collection and appropriate reference literature.

Finally, previous investigators (i.e., Artz 1992; Gonzalez 1987) have emphasized the possibility of coal contamination in radiocarbon samples collected from western North Dakota, a cautionary approach that is well justified considering the lithology of the Fort Union Group parent material. Laboratory procedures proven the most effective in the recognition of coal contaminants include paleobotanical (i.e., fossil pollen and spore) analysis, and microscopic reflectance analysis (Tankersley et al. 1987). Regional attempts at controlling for possible lignite contamination have included dating only the NaOH-soluble humate fraction in bulk soil samples (Root et al. 1986) and employing flotation techniques in an attempt to separate coal from lighter organic material (Gonzalez 1987). Both procedures, however, are not infallible because laboratory experiments have demonstrated that contamination can result from the presence of water soluble organic compounds in coal (Tankerstein et al. 1987:326). For the THRO samples, possible contamination was addressed first by carefully examining potential stratigraphic sections for the presence of macroscopically-visible lignite. If any coal was observed in the section, particularly if it was seen to be overlying potential samples, the sample was not collected. Secondly, all recovered samples were examined under a 40x binocular microscope for the presence of visible coal particles. Samples of lignite from local Fort Union Group sediments and from reworked alluvial contexts, were collected and used as a control reference. Thirdly, a small portion of each bulk soil radiocarbon sample was ground with a mortar and pestle and mounted onto slides with a small amount of glycerin jelly. The same was done with the above mentioned coal samples and with samples of wood charcoal (juniper). Under examination with a 100x to 200x refracting microscope, the coal samples were found to contain translucent, angular particles which were light brown to amber in color. The charcoal samples, by contrast, were comprised of particles that were opaque, black, and non-angular. Armed with this comparative information, each of the bulk soil sample slides was then examined. Coal particles were identified on all but two of the 19 slides, although 14 contained fewer than 4 particles. The remaining three samples, A-6483, A-6480, and A-7138, contained 7, 13, and 16 particles respectively. Significantly, these samples, which are associated with radiocarbon ages of 17,640 ± 230, 22,290 ± 310, and 6925 ± 175 yr B.P., were already rejected on the basis of gross
chronostratigraphic incongruities. The remaining sample ages are all stratigraphically consistent and do not appear to be significantly affected by the presence of coal particulates, in spite of the presence of small numbers of microscopic grains. These admittedly tentative and hardly infallible results suggest that non thin-section comparative analysis under relatively low magnification may be an inexpensive and expedient method of identifying coal contamination. More importantly however, it illustrates the need to date multiple samples from badlands landscapes in order to recognize coal-induced temporal inconsistencies. While two of the rejected radiocarbon samples had ages that were obviously too old, the third (A-6483) would have been difficult to recognize as erroneous were it not for the fact that significant stratigraphic and chronologic data were already available.
CHAPTER IV

THE ARCHAEOLOGICAL DATA BASE

Introduction

In order to realize the principal research objective of identifying major noncultural site formation processes through the establishment of a stratigraphic and paleoenvironmental context for the archaeological record, the key elements of that record must first be identified and described. Rather than attempt to synthesize and unify the data from hundreds of disjointed and often incompatible archaeological research reports, it was argued in Chapter I that a feasible approach would be to focus attention on the archaeology of a small but representative study area. As mentioned, the area chosen for study was the South Unit of Theodore Roosevelt National Park, and the archaeological record there was investigated during a three-season archaeological survey conducted between 1987 and 1989. The 4840 ha survey resulted in the discovery and recording of 171 prehistoric sites. An additional seven were located during the course of the 1991-1993 stratigraphic investigations, bringing the total site assemblage to 178 (Figure 5). Together these reflect a broad range of prehistoric cultural activities conducted within highly variable spatial and temporal parameters. Because of natural site formation processes, the archaeological record described in this chapter is not considered to be an accurate reflection of all prehistoric cultural activity that took place within the study area. Nevertheless, the South Unit data are considered typical of the types of archaeological information that remain extant in the badlands as a whole (cf. Kuehn 1990:114-163). Therefore, the purpose of this chapter will be to elucidate three key elements of the THRO archaeological data base: site type, site age and/or cultural affiliation, and site setting.

Site Type

The material remains of prehistoric human behavior at the individual site level, and the broader implications of sites, or recognizable activity loci, to overall settlement patterns, are described under the rather simply titled rubric of "site type". The emphasis, therefore, is on sites as the principal analytical units rather than on artifacts. While by no means continuously distributed
Figure 5. Map of THRO South Unit showing the location of recorded prehistoric sites.
across the landscape, individual artifacts are selectively scattered throughout the badlands. Every so-called "isolate" encountered during the THRO survey was described in terms of artifact type, raw material, and site location. It is acknowledged that the study of these data at some future point could help facilitate the "congruence between theory and method", which is a valid, explicitly stated goal of off-site, or distributional, archaeology (Ebert and Kohler 1988:144-148). Given the stated goals of the present study, however, with its emphasis on the elucidation of noncultural site formation processes, the site is still considered to be the most appropriate unit of archaeological analysis.

The site type classification utilized in this analysis is drawn from Binford's (1980) paper on hunter-gatherer settlement systems, which as demonstrated by Gregg et al. (1986), Hanson (1983), Kuehn (1990), Root (1983), and others, is well suited to the study of prehistoric human occupation in the Northern Great Plains. In particular, Binford's five categories of collector sites: the residential base, field camp, location, station, and cache, have proven the most utilitarian. This could be related to the fact that ethnographic and archaeological evidence suggests that hunter-gatherer, and even hunter-gatherer-horticultural, groups in the region tended to procure specific resources through the organization of task groups which ventured out of residential bases of varying occupational intensity (cf. Bowers 1948; Hanson 1983; Keyser and Davis 1984; Wilson 1928). Elsewhere in North America, archaeologists have recognized the applicability of similar site type categories when discussing prehistoric hunter-gatherer land use, particularly collecting strategy (cf. Ebert and Kohler 1988:113). Bamforth (1991), for instance, describes three principal site type categories in the archaeological record of the Santa Ynez Valley in California: "seasonal residential base", "short-term camp", and "limited-use sites". Both Binford and Bamforth recognize the residential base, while Bamforth's "short-term camp" is analogous to Binford's "field camp", and his "limited-use site" and Binford's "locations" and "stations" reflect similar types of activities.

Definitions

Residential bases represent "...the hub of subsistence activities" from which hunting/gathering forays originated (Binford 1980:9). They are also the areas where the widest range of activities took place (Binford 1980:9). As defined for the North Dakota area by Hanson (1983:1404), residential bases were "the foci of subsistence activities for prehistoric hunter-gatherers" and were occupied by either nuclear family, stem family, or band-level groups. Field camps, on the other hand, are "...temporary operational centers for a task group...where a task group sleeps, eats, and otherwise maintains itself while away from the residential base" (Binford 1980:10).
In North Dakota, the camps were likely occupied by nuclear family, stem family, or communal "task
groups", (i.e., hunting/gathering teams) whose purpose was to procure needed resources (Hanson
1983:1399-1407). Field camp activities should reflect both procurement/extractive tasks and
maintenance activities such as tool resharpening and food processing (Binford and Binford 1969;
Greiser 1985:46-47). Locations are resource procurement areas or "...places where extractive tasks
are exclusively carried out" (Binford 1980:9). These extractive activities are conducted by task
groups which could have originated out of residential bases or field camps, but which are ultimately
tied to the bases (Binford 1980; Hanson 1983). Locations are frequently associated with the
procurement of floral and faunal resources and artifact raw materials such as chipped stone, ground
stone, clay, or wood. Stations are "sites where special-purpose task groups are localized when
engaged in information gathering, for instance the observation of game movement...or the
observation of other humans" (Binford 1980:12). Stations are, therefore, generally located in areas
which offer good views of the surrounding terrain. Finally, caches are "field" facilities designed to
temporarily store bulk resources en route to consumers (Binford 1980:12). Caches frequently take
the form of an excavated storage pit or other relatively small enclosed feature (Gregg 1985a:109).

Caveats to Classification

In spite of these rather straightforward definitions, the study of prehistoric hunter-gatherer
land use can be a highly complex matter, and the caveats associated with prehistoric site type
classifications are numerous. These difficulties are due to variations in local conditions, the
associated complexities of human behavioral response (Bamforth 1991), and also to the myriad of
site formation processes that can affect the archaeological record once sites are abandoned (Schiffer
1987). In the case of the South Unit study area, these problems are compounded by the fact that
archaeological data are limited primarily to surface visible artifacts and features. It would be
fallacious to assume that surface artifacts reflect the entire range of materials possible at any given
site, unless the site is totally eroded, or for whatever reason, was not subsequently buried after
occupation. If not buried, the mixing of different cultural components becomes a real possibility (cf.
Waters 1992:97). Similarly, in the absence of careful chronological control, intrasite contemporaneity among the visible material remains is not assured. As discussed by Dewar (1991)
and others, prehistoric settlement and occupation length are highly dynamic variables which are
easily overlooked by archaeologists who view only "static" material remains. For this reason, it
cannot be assumed that the artifacts observed on any given site are the product of a single cultural
component. Therefore, some sites may represent palimpsests of prehistoric activity where the various components cannot be identified or differentiated on the basis of surface evidence. Consequently, the variables of site size, artifact density, and artifact diversity at these sites may be the result of multiple, but indistinguishable, group activities, and therefore not amenable to site type classification (cf. Ebert and Kohler 1988:120). As suggested by Ebert and Kohler (1988:120), this possibility most acutely affects the differentiation between field camps and residential bases, the latter of which are characterized by the longest occupation length and the highest range of on-site activities of any of the site-type categories (Binford 1980; Hanson 1983; Kent 1992). In THRO, the criteria for base camp identification were: (1) high artifact diversity; (2) evidence of domestic activity and extended occupation (i.e., evidence of habitation structures, features, cooking activity, maintenance tasks); (3) dense artifact concentrations; and (4) proximity to resources necessary for extended occupation, such as water and fuel (Kuehn 1990:135). The possibility that some of these criteria, particularly high artifact diversity and density, could be the result of multiple group activity not necessarily associated with extended occupation, ruled out the identification of any residential bases in the THRO South Unit study area. Consequently, the site type classifications are rather conservative, and only three of Binford’s five categories were identified. It is acknowledged that this approach may result in the combination and/or omission of otherwise separate categories of activity and is therefore idealized. It can also be argued, however, that the classification scheme does have some utility to the study of badlands archaeology by providing for a consistent grouping of site activities, a grouping that could prove useful in the development of future settlement models. In addition, the usefulness of surficial data to archaeological research, particularly to the development and testing of models on the organization of human systems, has been effectively demonstrated by other researchers (cf. Ebert and Kohler 1988). The THRO archaeological sites were subsequently identified as either field camps, locations, or stations on the basis of site size, site structure, artifact density, and artifact diversity. Similar criteria were utilized by Bamforth (1991:227), who identifies site type categories on the basis of “internal spatial structure” and “range of general activity types”.

Of the different variables utilized in the THRO site type classifications, the most amenable to mathematical quantification is artifact diversity, which was evaluated through the use of the Index of Evenness statistic, otherwise called the diversity index (Pielou 1969). Diversity index measures the range of on-site activities represented by different artifact classes as they relate to site use and intensity of occupation (Wood 1978:260). Sites with high diversity indices indicate a wide variety of site activities, while a low index indicates a limited number of activities. Those sites with
evidence of numerous types of activities, such as manufacturing, processing, and maintenance, are more likely to reflect longer term occupations than do those sites with a limited range of activities (Binford 1980; Binford and Binford 1969). In the South Unit study area, diversity indices above ca. .4500 are considered high while indices less than ca. .1500 are considered low. As with the other caveats to site type classification, the use of diversity index as an indicator of occupation length is not without potential problems. Jones et al. (1983), among others, have demonstrated a relationship between artifact class richness and artifact sample size. That is, the apparent richness or diversity of a site assemblage may be heavily influenced by the size of the recovered sample. This relationship is most acute in sites with low artifact counts. Since many of the THRO sites fall into this category, the potential for sample-size bias must be considered a possibility. Nevertheless, many of the sites with low artifact counts were found to have high diversity indices, and these sites were invariably classified as field camps because they did not meet the criteria established for the other site type categories. Therefore, while acknowledging the potential relationship between apparent richness and sample size, it is argued that the site type classifications are still useful in providing some quantifiable measure of site activity and occupation length, especially as they relate to differences between extended occupation (i.e., residential bases), limited occupation (i.e., field camps), and resource procurement (i.e., locations).

Kent (1992) presents an ethnoarchaeological model of prehistoric mobility patterns based on a number of similar variables (i.e., artifact quantity and diversity). In her analysis, Kent recognizes the relationship between increased material density, diversity, and length of occupation: a relationship that is statistically tested through bivariate regression analysis (Kent 1992).

In THRO, the diversity index determinations were based on the formula presented in Wood (1978:260), where \( k \) = the number of possible artifact categories present in the data base assemblage. For the THRO sites, \( k = 29 \), which is the number of different artifact categories observed during the course of the field investigations (Table 1). Each category is assumed to represent a discreet on-site activity. Twelve of the categories reflect activities associated with stone tool production and maintenance (i.e., raw material testing, flake and biface production, core reduction, tool resharpening), while the remainder reflect domestic and resource procurement/processing activities (i.e., cutting, scraping, whittling, perforating, pounding, grinding, fire production, bone grease rendering). The resultant diversity indices from the THRO South Unit sites are presented in Table 2.
Table 1. Summary of Index of Evenness Artifact Categories (k).

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<td>1.</td>
<td>Hardhammer flakes and shatter</td>
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<td>2.</td>
<td>Bifacial thinning flakes</td>
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<td>3.</td>
<td>Pressure flakes</td>
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<td>Flake blades</td>
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<td>Stage 2 bifaces</td>
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<td>Stage 3 bifaces</td>
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<td>Stage 4 bifaces</td>
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<td>Stage 5 bifaces</td>
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<td>10.</td>
<td>Exotic materials and trade goods</td>
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<td>Projectile points</td>
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<td>Flake tools</td>
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<td>Endscrapers</td>
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<td>Sidescrapers</td>
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<td>Composite tools</td>
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<td>16.</td>
<td>Tested raw material and split cobbles</td>
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<td>Perforator</td>
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<td>18.</td>
<td>Fire cracked rock</td>
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<td>Bone and teeth, unmodified</td>
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<td>Charcoal</td>
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<td>Ceramics</td>
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<td>Bilateral cutting tools</td>
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Results

Field camps proved to be the largest site type category in terms of the number of sites recorded (n = 100). They account for 56.2% of the total site assemblage (Table 3), and are found in virtually every landform setting and depositional environment in the study area. The field camps consist primarily of lithic or cultural material scatters of varying size, density, and artifact diversity. Lithic scatters are concentrations of chipped stone flaking debris, cores, pieces of tested raw material, and stone tools in various stages of manufacture and life-cycle. Cultural material scatters generally contain lithics in combination with other material categories such as fire-cracked rock, ceramics, or bone.

The high percentage of field camps appears, to some extent, to be a reflection of the difficulties associated with the identification of site type. For instance, field camps often became the "catch-all" category for sites that did not exhibit evidence of specific resource procurement or that did not meet the criteria established for base camp inclusion. In the case of the former, some of the lithic scatters more than likely witnessed various types of resource procurement activities in addition to flintknapping, but the physical remains of those activities are no longer visible. In other words, the predominance of flaking debris at many of the sites may be more a reflection of the non-perishable nature of lithic materials, and not the predominance of flintknapping as an on-site activity. As an example, a ridgetop location at the head of a hardwood draw could have been selected for resource procurement activities, say chokecherry gathering. While some members of the group were busy procuring chokecherries, others (i.e., adult males) may have been passing the time resharpening stone tools. After the chokecherries were gathered and the group moved on, the only materials discarded at the site were flaking debris and a few chokecherry pits. Because the chokecherry remains quickly deteriorated in the elements, or were carried away by animals, all that survived to become part of the archaeological record were flakes. Consequently, a site such as this would have been classified as a field camp because no evidence of specific resource procurement was available.

Despite these difficulties, the field camp category is comprised of sites that by and large do fit the description of a temporary camp (cf. Binford 1980; Hanson 1983). They have higher overall diversity indices than do locations and stations (field camp mean = .2641, location mean = .2162, station mean = .1203), and are not solely associated with resource procurement activity. The South Unit field camps include a small sandstone rockshelter (32BI649), and 12 to 15 large, open-air cultural material scatters. The latter exhibit high diversity indices (i.e., between .450 and .550), evidence of fire hearths, dense artifact concentrations, faunal remains, and temporally diagnostic.
<table>
<thead>
<tr>
<th>Locations: Lithic Raw Material Procurement</th>
<th>Locations: Field Camps</th>
<th>Locations: Faunal Procurement Procurement</th>
<th>Stations</th>
</tr>
</thead>
</table>
artifacts (i.e., projectile points and ceramics). The more prominent of these include 32BI548, 32BI670, 32BI672, 32BI706, and 32BI614 (Table 3, Figure 5).

Structural remains associated with field camps in the North Unit of THRO and from other portions of the badlands include conical timbered lodges and stone circles (cf. Allen 1982; Beckes and Keyser 1983; Kuehn 1990; Loendorf 1978), however both types of features are relatively rare (in the actual badlands). Conical timbered lodges consist of a tipi-like arrangement of closely spaced juniper poles, often covered with brush. They are almost exclusively Late Prehistoric and Historic in age, and are found throughout the Northern Great Plains and Central Rocky Mountain regions (cf. Ewers 1968; Frison 1978). In western North Dakota, conical timbered lodges are reported primarily from well-sheltered locations such as hardwood draws or riparian forests. Stone circles are circular arrangements of rocks which are believed to have served as weights for hide tipi covers. They date from the Middle Plains Archaic through the Historic periods, and are ubiquitous in the Northern Great Plains and portions of the Rocky Mountains (cf. Davis 1983; Schneider 1982).

Locations are the second most prevalent site type in the study area, accounting for 43.3% of the recorded site aggregate (Table 3). The vast majority of the 77 locations (97.4%) are lithic raw material procurement sites associated with outcrops of channel lag gravels (both in situ and reworked). The gravels are associated with the Late Miocene/Pliocene Missouri Slope degradational surface (i.e., Flaxville surface) and to several Pleistocene terraces cut by the Little Missouri River (Collier and Thom 1918; Kuehn 1990; Laird 1950; Petter 1956). The former are concentrated on Petrified Forest and Radio Tower Plateaus (Kuehn 1990; Petter 1956), while the latter are extant on Big Plateau (Kuehn 1990; Laird 1950), Johnson Plateau, and Airport Plateau. While gravels from both sources have been redeposited onto lower-lying knobs and foothills, most of the locations are concentrated along the eroding edges of the plateaus and adjoining finger ridges.

The lithic raw material locations are characterized by evidence of hard hammer material testing, core reduction, small flake production, and limited, initial bifacial tool manufacturing (cf. Ahler and Christensen 1983; Callahan 1979; Kuehn 1990:141). The most common artifacts are cores, pieces of tested raw material, flakes, hammerstones, and occasional bifacial fragments. The sites have generally low diversity indices (mean of .2162), and none were associated with surface-visible diagnostic artifacts. The target resources are cobble-sized nodules of chalcedony, agate, petrified wood, Knife River flint, quartzite, porcellanite, chert, Tonque River silicified sediment, and Rainy Buttes silicified wood (Kuehn 1990:141).
The two remaining locations are faunal resource procurement areas. The first, 32B1549, is a bison kill site located at the base of the Little Missouri Escarpment (Figure 5). Numerous pieces of fractured bison bone (over 50 separate elements associated with at least three individuals) were observed eroding out of linear slump deposits. An 18 m high cliff is present a short distance south of the site, suggesting that the bison may have been procured by means of a jump. On the other hand, a natural spring is located immediately west of the site and much of the site area is waterlogged from spring discharge. This condition suggests that the animals may have been trapped or bogged-down in the wet muddy sediment. In any event, the site is associated with Plains Village ceramics and is located in an ideal setting for bison procurement activities. A well known bison kill site, the Geary Buffalo Jump, is situated in a similar Escarpment location approximately 20 km to the south (Beckes and Keyser 1983:197; Loendorf et al. 1982).

The final faunal procurement location, 32B1518, is a possible eagle trapping pit. The site consists of a 1.75 x 1.5 m oval depression, .65 m deep, located near the eastern edge of Boicourt Ridge (Figure 5). The feature is not associated with visible cultural materials, but is similar in morphology and setting to eagle-trapping pits recorded elsewhere in the western North Dakota. Eagle trapping activities in the badlands have been ethnographically associated with Plains Village groups such as the Mandan and Hidatsa (Allen 1982; Wilson 1928).

Stations, or observation and information gathering loci, are represented in the study area by a single site, 32B1662. This paucity is attributed to the problems associated with identifying observation activities solely on the basis of archaeological material evidence. While observation was probably a key activity at many of the upland sites (especially those with good views of the surrounding countryside), it is difficult to document. Superior view (in terms of distance and degree) has been identified as one of the key characteristics of upland sites in the THRO study area (Kuehn 1990:192). Although observation activity very seldom results in tangible material remains, the single THRO station was classified as such because it was located in an area with an exceptional view (i.e., greater than 8 km in distance and 360 degrees in aspect), and contained no evidence of on-site activity other than limited flintknapping. The site is situated on the northwestern edge of Big Plateau (Figure 5), and has a very low diversity index (.1203). The only discernable site activity was the reduction of a single quartzite cobble. These factors, together with the absence of subsurface artifact potential, suggest that observation was the predominant site function, in this case probably game spotting. It is hypothesized that the flintknapping was little more than a leisure activity while members of the group were waiting for game to appear in the surrounding lowlands. Target animals
may have included bison, elk, mule deer, white-tailed deer, bighorn sheep, and pronghorn antelope. In light of the fact that observation is very difficult to identify archaeologically, the identification of site 32Bl622 as a station is highly tentative.

It is important to note that all five of Binford's site type categories have been identified in the Little Missouri Badlands, although two of the types are quite rare. A single cache of Knife River flint biface preforms was located in the Sentinel Butte area (L. Loendorf, personal communication 1994). Likewise, only a handful of sites appear to meet the classification for residential bases; all of which are situated within the Little Missouri River valley. Three of the bases are Besant sites located in the North Unit of Theodore Roosevelt National Park (Kuehn 1990), and two are Plains Village sites located on private land south of Medora (Johnson 1983; Metcalf and Schweigert 1987). A possible fourth residential base (Besant) was recently investigated near the extreme southern edge of the badlands (Toom 1992a). The North Unit sites are cultural material scatters comprised of chipped stone tools, flaking debris, ground stone tools, ceramics, fire-cracked rock, and faunal remains. These sites have the highest diversity indices of any discovered during the THRO survey (mean of .5134). Even with the luxury of extensive subsurface data, previous investigators have had difficulties elucidating site type in the badlands region. When the question was addressed, terminology and interpretation have tended to be highly varied and inconclusive. With the exception of Kuehn (1985, 1987), Metcalf and Schweigert (1987), Toom (1992a), and Wermers et al. (1991) all of the sites subjected to large-scale subsurface excavations were classified in one way or another as field camps (cf. Avaizian 1981; Blikre and Borchert 1992; Borchert and Loendorf 1986; Campbell et al. 1983; East et al. 1985; Floodman et al. 1983; Fox 1981; Kuehn 1982, 1986; Kuehn et al. 1987; Metcalf 1984; Peterson and Foster 1991; Simon et al. 1982; Simon and Borchert 1981a, 1981b, 1981c; Simon and Keim 1983). All of these except the Sunday Sage site (Simon and Borchert 1981c), Davis Dam site (Peterson and Foster 1991), and Pretty Butte site (Borchert and Loendorf 1986), are located in upland settings. This pattern is discussed in some detail in Chapter VII.

Site Age and Cultural Affiliation

Introduction and Theoretical Context

The goals of contemporary archaeology as elucidated in Chapter I, are first and foremost dependent upon the establishment of a sound chronological framework for the archaeological record
(cf. Gregg 1985b:67). Indeed, temporal models have long been a traditional concern of archaeology, both in the Old and New Worlds (cf. Childe 1925; Kidder 1924; Willey and Phillips 1958). Because they are still necessary for the drawing of substantive behavioral inference, chronologies are an aspect of archaeology that cannot be outgrown (Trigger 1989:409).

The chronological model applied to the THRO site assemblage follows that developed by Gregg (1985a) for central and western North Dakota (Figure 6). The chronology is comprised of archaeological unit terms taken from Frison (1978), Lehmer (1971), Lovick and Ahler (1982), and Reeves (1970), among others. The three principal unit terms in Gregg's (1985a) chronology are tradition, complex, and phase. Each is formulated with respect to geographical distribution, cultural composition, and temporal framework, however traditions are concerned more with behavioral characteristics (i.e., lifeways), complexes with material remains, and phases with both material culture and temporal context (Gregg 1985b:69-75). Traditions are identifiable adaptive strategies and patterns of behavior that operated over large geographical areas for long periods of time, while phases are elements of a tradition expressed at the regional level (Gregg 1985b:71-72). Complexes are aggregates of distinctive material remains with demonstrable reoccurrence within a tradition (Gregg 1985b:73).

Like most chronologies, the one applied to western North Dakota is based on the arrangement of archaeological unit terms into a temporal framework established by relative and numerical dating techniques. Numerical dating methods, especially radiocarbon, have been widely utilized by archaeologists since the 1950's and 60's. However, many of the unit terms, particularly the larger units such as traditions and complexes, were identified and described before numerical dating techniques were available. Consequently, relative dating methods, especially artifact typological cross dating and stratigraphic position, have been the cornerstone of chronology-building in the Plains area.

The central element of typological cross dating is the artifact "type", which has been defined as "...the division of an assemblage of materials...into groupings based on the conscious recognition of dimensions of formal variation possessed by these phenomena" (Hill and Evans 1972:233). Types are groups of materials with a consistent patterning of attributes that are distinguishable from the attributes of other types. The attributes associated with each type are non-random in their distribution (Hill and Evans 1972; Spaulding 1953). In western North Dakota, the principles of cross dating can be applied to at least 12 different projectile point types and several different ceramic types.
Figure 6.  Chronological model for western and central North Dakota (from Gregg 1985a:80).
In addition to traditions, complexes, and phases, a fourth unit term, the period, is also important to the chronological framework in western North Dakota. Periods are defined as "horizontal time bands marked off on the chronological chart in years, centuries, or millennia" (Willey 1966:7, from Gregg 1985b:73). In the badlands study area, some sites have been assigned to periods, or general blocks of time, on the basis of typological cross-dating, however they cannot be identified with specific named cultural groups because of a lack of more culturally-sensitive artifacts. An example are sites that contain only Plains side-notched projectile points, which have a temporal range of ca. 650 - 250 yr B.P. (Kehoe 1966), but which are associated not only with the Plains Village tradition (Lehner 1971), but also with Northwestern Plains nomadic groups such as the Old Women's complex (Forbis 1962), the Mortlach aggregate (Wetlaufer 1955), and the Powder River tradition (Keyser and Davis 1981). In lieu of more specific information, sites such as these can only be identified as dating from the Late Prehistoric period. Periods are also useful for classifying sites that are associated with radiocarbon ages but not with diagnostic artifacts, and with sites containing non-descript ceramics such as smoothed, undecorated plainwares (Kuehn 1990).

With this introductory discussion as a theoretical context, we can now turn to the age and/or cultural affiliation of the THRO South Unit sites.

Results

Prehistoric cultural traditions identified in the South Unit study area include Paleoindian, Plains Archaic, Plains Village, and Plains Woodland. These archaeological units, however, are far from equally represented in the recorded site assemblage. Also, it is important to recognize that the vast majority of sites (n = 156 or 87.6%) could not be assigned to an archaeological unit or general time period because of a lack of surface-visible artifacts and/or datable materials. This paucity may reflect the temporary and expeditious nature of many badland sites, as indicated for instance, at the lithic raw material locations. Indeed, all but one of the sites that could be culturally and temporally classified are field camps, a factor that suggests a relationship between length of site use and the presence of diagnostic artifacts. The sites and/or cultural materials assignable to specific cultural units or temporal periods are as follows.

Paleoindian Tradition, Hell Gap-Agate Basin Complex. A single projectile point, identified by Ludwickson (1981) as Agate Basin, was recovered by James Sperry in 1969 at site 32B1648 on Petrified Forest Ridge (Figures 5 and 7). The Knife River flint specimen was apparently broken,
Figure 7. Photograph of Agate Basin projectile point collected from site 32BI648.

discarded, and subsequently retipped; a sequence of events suggested on the basis of morphology, flaking patterns, and the fact that the blade is somewhat less patinated than the haft element (Ludwickson 1981:40). Because of the possibility that the artifact was apparently reworked some time after its manufacture, a Paleoindian component at site 32BI648 cannot be conclusively demonstrated. While there is little doubt that the projectile point was originally associated with the Hell Gap-Agate Basin complex, ca. 10,500 - 9500 yr. B.P. (Frison 1978; Irwin-Williams et al. 1973, from Gregg 1985a:91), it could have been collected by members of a later group and then reworked, used, and discarded along with the other debris at the site. There is, however, the possibility that the artifact may have originally been discarded somewhere on the ridgetop, if not in the actual site area.

Site 32BI648, a field camp, covers a 380 m long portion of the western edge of Petrified Forest ridge and contains several distinct concentrations of flaking debris, chipped stone tools, fire-cracked rock, and faunal remains. It was re-recorded by the author in 1988, during which time a projectile point identified as Late Prehistoric (ca. 2000 - 150 yr. B.P.) was collected from the northern end of the site (Kuehn 1989:39). The presence of this artifact suggests that either: (1) the Agate Basin projectile point could have been reworked by members of an unidentified Late Prehistoric group, and that the site contains a single Late Prehistoric component, or (2) that the site has multiple, but mixed, components, one at least of which is Late Prehistoric and one at least of which is Paleoindian or some other group predating 2000 yr B.P. The first scenario is supported by the slight differential patination evident on the Agate Basin specimen and by the lack of other temporally diagnostic artifacts in the site area. The second scenario is supported by the large spatial extent of the site, the presence of multiple concentration areas, and by the fact that the site is highly
eroded (thereby facilitating artifact mixing). Although 32B1648 may or may not contain a Paleoindian component, the possibility that Paleoindian-aged materials are extant on the narrow ridgetop is suggested by the presence of late Pleistocene/early Holocene sediments 600 m from 32B1648. At stratigraphic Locality O, at the southern end of the ridge, 12 buried soils were encountered in a 3.4 m thick loess deposit (Figure 4). Two of the soils are associated with radiocarbon ages of 11,560, and 10,730 yr B.P. (Kuehn 1993:328). The presence of an Agate Basin projectile point and late Pleistocene/early Holocene-aged sediment on the same small ridgetop indicates that a Paleoindian component in the area is at least possible. Furthermore, the absence of recognizable equivalent paleosols at 32B1648 suggests that erosional processes stripped the older loess deposits from this portion of the ridgetop. If an intact Paleoindian component was present in the site area, the erosion could have resulted in a mixing of the Paleoindian materials with those of more recent cultural groups. In other words, later groups may have camped in an area where Paleoindian artifacts were lying in a deflated, surficial context. One of these artifacts could have been the broken Agate Basin projectile point, which was subsequently reworked, used (?), and discarded yet a second time.

Plains Archaic Tradition, Logan Creek-Mummy Cave Complex. Like the single Paleoindian projectile point, one Early Plains Archaic artifact was encountered during the course of the geoarchaeological research in the South Unit study area. A side-notched projectile point identified as Simonson was recovered during stratigraphic profile excavations at Locality O on Petrified Forest Ridge (Figure 4). As mentioned in Chapter II, Simonson points, dating from ca. 7500 - 5250 yr B.P., are the principal diagnostic artifact type of the middle Holocene Logan Creek-Mummy Cave complex. According to Gregg (1985a:101) the complex derives its name from two well documented sites in the Central and Northwestern Great Plains: the Logan Creek site in Nebraska and the Mummy Cave site in Wyoming. The specimen collected from Locality O, which also corresponds to the location of a field camp with at least one previously identified Late Prehistoric component (site 32B1703), consists of a moderately-patinated haft element of Knife River flint (Figure 8a). The point was recovered at a depth of 175 cm from stratigraphic Unit II, a pedogenically-altered loess deposit that produced a soil humate radiocarbon age of 6580 ± 160 yr B.P. (A-7109). The age correlates with the Pick City Member of the Oahe Formation, which was deposited under arid conditions between ca. 8500 - 4500 yr B.P. (Clayton et al. 1976; Kuehn 1993:328). Although the projectile point was the only artifact observed in association with the Pick City sediments, the
Figure 8. Drawings of key diagnostic projectile points, actual size: (a) Simonsen, 32B1703; (b) McKean lanceolate, 32B1522; (c) Pelican Lake, 32B1567; (d) unclassified Late Archaic, 32B1649; (e) Besant, Isolate 9E; (f) Plains side-notched, 32MZ915.
stratigraphic profile was only one m wide, and therefore does not constitute a thorough subsurface investigation of the site area. This factor, along with the well-defined stratigraphic context of the artifact, and the lack of visible evidence of tool re-working, suggests that there could be a Logan Creek-Mummy Cave component at 32B1703 (in addition to a more recent component of Late Prehistoric age).

**Plains Archaic Tradition, McKean Complex.** The earliest widely studied cultural unit in western North Dakota, the McKean complex, is represented in the South Unit by four recorded sites. The complex is well dated at between 4500 and 3000 yr B.P. and is characterized by five different projectile point types (Gregg 1985a; Syms 1969). These are McKean lanceolate, Mallory, Yonkee, Duncan, and Hanna. All but Duncan and Hanna were encountered in the South Unit study area, although a Hanna point was collected from a site in the THRO North Unit (Kuehn 1989:41). Likewise, Duncan points are commonly reported from sites throughout the Little Missouri Badlands (Beckes and Keyser 1983:177). In general terms, the McKean lanceolate (Figure 8b) appears to be the earliest point style in the complex, followed by Yonkee, Duncan, Mallory, and Hanna. There is, however, a great deal of overlap in both temporal and spatial distribution. The four South Unit McKean sites consist of three field camps located on the edge of the Little Missouri Escarpment (32B1522, 32B1548, 32B152), and one field camp, 32B1614, located on a low ridge within the Little Missouri River valley (Figure 5). In addition to McKean materials, site 32B1548 produced evidence of Besant and Late Prehistoric components, while 32B1614 produced evidence of Besant and Pelican Lake components. Both sites therefore appear to be palimpsests of short-term field camp activity, although stratigraphic separation between components is only evident at 32B1614. The lack of separation at 32B1548 is ascribed to post-occupational erosion, which appears to have resulted in a mixing of the cultural materials.

**Plains Archaic Tradition, Pelican Lake Complex.** The Late Archaic Pelican Lake complex, ca. 3500 - 1700 yr B.P., is also well represented in the Little Missouri Badlands; a factor reflected by the identification of five Pelican Lake sites in the South Unit study area. Similarities in geographical distribution and settlement and subsistence patterns, have led researchers to hypothesize that, in the Northern Plains, Pelican Lake developed out of the McKean complex (Reeves 1970, from Gregg 1985a:113). While Pelican Lake points are highly variable in size and overall morphology, there is evidence for a general trend toward decreased point size through time
(Gregg 1985a:112-113). Some excavated sites, however, also demonstrate a bimodal distribution in point sizes within a single assemblage (Gregg 1987:264-265,444-445).

The five South Unit sites with Pelican Lake components are all classified as field camps. Each produced a single Pelican Lake side-notched projectile point (Figure 8c). One site (32BI573) is located on the edge of the Little Missouri Escarpment, two (32BI562 and 32BI615) are located on foothills above ephemeral streams (Jones Creek and Boicourt wash), one (32BI567) is on a gentle colluvial slope, while the fifth (32BI614) is on a low ridgetop within the Little Missouri River Valley (Figure 5). Only one of the five sites (32BI614), contained evidence of multiple components (McKean, Pelican Lake, and Besant). Vertical separation between components is evident at this site.

**Unclassified Late Plains Archaic Period.** Two of the South Unit sites are assigned to the time period corresponding to the Late Plains Archaic tradition (ca. 3500 - 2000 yr B.P.). In both cases, the sites did not produce artifacts that could be definitively associated with specific Late Archaic cultural groups. The first, 32BI649, is a small sandstone rockshelter and field camp that, in addition to a sparse scatter of chipped stone cores and flaking debris, was also associated with a heavily patinated projectile point that does not fit existing point typologies for the region. The specimen contains shallow side-notches close to the slightly concave base (Figure 8d). The general morphology and placement of the notches is similar to points associated with the Plains Woodland Avonlea complex, ca. 1500 - 950 yr B.P. (Gregg 1985a:128), although the THRO specimen is larger and thicker than typical Avonlea points (Kehoe 1966). Other than the shallow side-notches, the point is most closely affiliated with some Pelican Lake varieties, especially with regard to the overall size and shape of the blade (Reeves 1970). Also, the presence of heavy patination is not common on Knife River flint artifacts younger than ca. 2000 yr B.P. (Ahler and Christensen 1983). This suggests that the specimen pre-dates Avonlea, and that the component at 32BI649 is most likely Late Plains Archaic in age. Similar enigmatic projectile points are commonly reported throughout the Northern Plains and reflect the high degree of stone tool variation associated with this time period. Such variation could be the result of relatively high population densities and a tendency toward increased regionalism (Gregg 1985a:117).

The second unclassified Late Plains Archaic site, 32BI629, contained two eroding cultural features exposed in the scarp of a terrace adjacent to the Little Missouri River (Figure 5). Feature #1, an amorphous hearth associated with charcoal and bone fragments, was exposed at a depth of 254 cm near the base of the scarp cutbank. Charcoal recovered from the excavated feature was
radiocarbon dated at 2190 ± 130 yr B.P. (Beta-27718). Feature #2, a concentration of bone fragments and fire-cracked rock exposed at a depth of 25 cm, was not associated with materials suitable for dating. In the absence of culturally diagnostic artifacts, the lower hearth is identified only as Late Plains Archaic on the basis of the radiocarbon age.

**Plains Woodland Tradition, Besant Complex.** Once thought to be poorly represented in the badlands region (Beckes and Keyser 1983:190), sites associated with the Besant complex, ca. 2050 - 1200 yr B.P., have since proven to be as frequent as any other named cultural group (Kuehn 1990:122). The South Unit research resulted in the identification of four Besant components. An additional seven were encountered in the THRO North Unit (Kuehn 1990). All are characterized by the presence of large side-notched Besant dart points, as defined by Johnson (1970). Smaller side-notched Samantha arrow points, reported from Besant components elsewhere in the Northern Plains (cf. Gregg 1985a, 1987; Reeves 1970), were not recovered from the THRO sites. The dart points collected from the four South Unit sites range in length from 30 - 46 mm and have an average width of 36.5 mm (Figure 8e).

Being the earliest named cultural unit in western North Dakota to produce ceramics (Gregg 1985a), Besant components are occasionally identified on the basis of diagnostic pottery. Although no Besant ceramics were encountered in the South Unit, they were recovered at five of the sites in the North Unit. Like Besant/Sonora ceramics from elsewhere in North Dakota (cf. Gregg 1985a), the THRO sherds exhibit cord roughened exterior surface treatment, straight rims with flat lips, crushed granite temper, and colors ranging from pale brown to reddish yellow on the exterior, and pale brown to pink on the interior (Kuehn 1990:123; 1989:Figure 37a-e).

The four South Unit sites with Besant components consist of two along the edge of the Little Missouri Escarpment (32Bl548 and 32Bl575), and two on loess-covered ridgetops (32Bl614 and 32Bl706). Three of the Besant sites in the North Unit have been classified as residential bases, indicating that the Little Missouri Badlands may have been part of the core territory of one or more Plains Woodland groups (Kuehn 1990:156-157; Syms 1977). A similar possibility has been suggested for the McKean complex (Hill 1988), although this hypothesis is neither supported nor refuted by the THRO data (Kuehn 1990:157).

**Plains Village Tradition.** The utilization of the badlands region by Plains Village groups, ca. 950 - 150 yr B.P., is ethnographically and archaeologically well documented (Bowers 1948;
Johnson 1983; Wilson 1928). The increasingly available archaeological evidence includes the discovery of six Plains Village sites in Theodore Roosevelt National Park, four of which are located in the South Unit study area.

All of the THRO Plains Village sites are characterized by the presence of culturally diagnostic ceramics. Two of the sites, including one in the South Unit (32BI568), also contained Plains side-notched projectile points (Figure 8f). As mentioned, Plains side-notched is the principal point type of not only the Plains Village tradition, but also of numerous Late Prehistoric nomadic groups.

Plains Village ceramic attributes include smoothed, simple stamped, and check stamped surface treatments; cord impressed, tool impressed, trailed, finger impressed, and stab and drag decorative techniques; and S-shaped, straight, and outcurved rim forms (Gregg 1985a:148; Johnson 1980; Lehmer 1971). The THRO ceramic assemblage includes sherds that are smoothed, simple stamped, check stamped, and cord impressed (Figure 9). Simple stamped ceramics represent lump modeled vessels finished with a grooved or thong wrapped wooden paddle, while check stamping results from the use of a cross-grooved wooden paddle (Calabrese 1972:18, from Gregg 1985a:148). Simple stamping as a surface finishing technique occurs from the Extended Middle Missouri through the Post-Contact Coalescent variants of Lehmer's (1971) chronology for Plains Village sites in the Middle Missouri subarea of North and South Dakota (i.e., from ca. 950 - 150 yr B.P.). This corresponds to the Clark's Creek through Knife River phases of Lovick and Ahler's (1982) chronology for sites in the Knife-Heart region of westcentral North Dakota (Figure 6). Check stamping in the Knife-Heart region was a common surface finishing technique during the Nailati Phase, ca. 800 - 600 yr B.P., and the Scattered Village complex, ca. 600 - 300 yr B.P. (Lovick and Ahler 1982:71-73, 178-182). Cord impressed decorative motifs were used occasionally from ca. 1100 - 600 yr B.P., and from ca. 445 - 320 yr B.P., but were most common in the Post-Contact Coalescent variant (Lehmer 1971), and the Heart River and Knife River phases (Lovick and Ahler 1982). These archaeological unit terms correspond to the period from ca. 300 to 150 yr B.P.

It should be noted that simple stamped, check stamped, and cord impressed ceramics are reported from sites associated with a number of Northern Great Plains groups other than Plains Village. Principal among these are the Powder River tradition, ca. 950 - 200 yr B.P. (Keyser and Davis 1981), and the Mortlach aggregate, ca. 500 - 220 yr B.P. (Gregg 1985a; Schneider and Kinney 1978; Syms 1977). No sites associated with either of these groups have been identified in the Little
Missouri Badlands, and it is argued (Kuehn 1990:126-127) that simple stamped, check stamped, and cord impressed ceramic-bearing sites in the badlands should remain classified as Plains Village until sufficient data to the contrary become available. This argument is based on the close geographic proximity of the Little Missouri Badlands to the Middle Missouri subarea, and to the ethnographic evidence that Plains Village groups conducted hunting and gathering forays into the Little Missouri region (Kuehn 1990:126).

The four Plains Village sites identified in the South Unit consist of a bison procurement location (32BI549), and three cultural material scatters (32BI626, 32BI568 and 32BI731). The last three are associated with landforms whose ages have been estimated on the basis of stratigraphic correlation with dated geologic sections (see Chapter V). The sites, therefore, are good examples of the utility of stratigraphic dating to chronology building in archaeology.

Unclassified Late Prehistoric Period. The largest number of sites in the THRO assemblage (n = 8) are synchronous with the Late Prehistoric period, ca. 2000 - 250 yr B.P. (Figure 6; Frison 1978). These either contain culturally ambiguous Plains side-notched arrow points or are associated with the general time period on the basis of stratigraphic evidence. In the Northern Great Plains, calendar ages associated with the end of the Prehistoric period are widely varied, depending upon the first documented arrival of Europeans and Anglo-Americans into a particular region, and on estimates as to the time of the first arrival of horses and trade goods (cf. Brown 1968; Byrne 1973; Frison 1978; Gregg 1985a; Lehmer 1971). It is generally assumed that horses and trade goods
preceded the arrival of significant numbers of Europeans and Anglo-Americans because of exchange among indigenous groups (Ewers 1968:23). Therefore, the time between the first arrival of horses and trade goods and the first significant numbers of Anglo explorers and settlers has been termed the Protohistoric period (cf. Frison 1978:72-77). The historic period in the Little Missouri Badlands is given a beginning date of A.D. 1742 (252 yr B.P.), which is the year of the Francois and Louis La Verendrye expedition up the Little Missouri River on their way to the Rocky Mountains (Kuehn 1990:129; Nelson 1946:40). Brown (1968:91-92) uses the year A.D. 1800 as the earliest appearance of European trade goods into the nearby Crow territory of southeastern Montana. This date may be applicable to the Little Missouri region in a practical sense, as trade goods may not have been widely available until this time. Also, the Little Missouri Badlands did not experience the arrival of significant numbers of Anglo Americans until the military campaigns of the 1860's and 1870's. Therefore, the time period between A.D. 1742 and ca. 1860 could technically be classified as the Protohistoric period. This fine temporal distinction, however, is not yet evident in the badlands archaeological record.

The eight unclassified Late Prehistoric components consist of five sites with Plains side-notched arrow points, one with radiocarbon-dated hearth features, and two with components stratigraphically situated above dated radiocarbon samples. The sites with Plains side-notched arrow points, ca. 650 - 250 yr B.P. (Figure 8f), are 32BI548, 32BI648, 32BI695, 32BI566, and 32BI91-1 (Figure 5). The sixth site, 32BI579, contained two shallow basin-shaped hearths located from 13 - 20 cm below the surface of the T3 terrace in Paddock Creek. The hearths, excavated in 1987, contained charcoal, flaking debris, fire-cracked rock, and bone fragments, but no diagnostic artifacts (Kuehn 1990). The charcoal from the hearths yielded radiocarbon ages of 1390 ± 100 yr B.P. (Beta-27442) and 1180 ± 90 yr B.P. (Beta-27745), which are just inside the suggested terminal age of the Besant complex (Gregg 1985a). Site 32BI703, as previously discussed, is a multiple component field camp (Figure 5). Non-diagnostic artifacts, consisting of flakes, bone fragments, and fire-cracked rock, were observed in situ from 5 - 20 cm below the surface (Kuehn 1990:128). A soil humate age of 2160 ± 70 yr B.P. (Beta-33062) was obtained from a paleosol situated below the cultural material at a depth of 31 - 46 cm (Kuehn 1993:329). Also previously discussed, the final site, 32BI629, contained two cultural components visible in the T2 terrace of the Little Missouri River (Figure 5). Since the lower component was dated at 2190 yr B.P., the vertical distance between the two components suggests that the uppermost is Late Prehistoric in age (i.e., less than 2000 years old).
Although only 12% of the South Unit sites contained components that could be assigned to named cultural groups or general time periods, these do span most of the known length of prehistoric occupation in the Northern Great Plains. The number of diagnostic materials recovered, however, increases markedly through time. While only single Paleoindian and Early Plains Archaic artifacts and/or components were identified, the number of components associated with the Middle Plains Archaic (i.e., McKean complex) increases significantly to four. Even more dramatic increases are evident in the number of Late Plains Archaic (i.e., Pelican Lake, Besant, and unclassified) and Late Prehistoric (i.e., Plains Village and unclassified) components (n = 11 for Late Archaic and n = 12 for Late Prehistoric). The causal factors behind this incongruous temporal distribution will be discussed at length in Chapter VII.

Site Setting

Introduction

This section addresses the question of site location; a topic in archaeology that generally includes settlement pattern analysis and/or predictive site modeling (cf. Judge and Sebastian 1988; Plog and Hill 1971). It is rather surprising that both the spatial analysis of archaeological data at the inter-site level, and the relationships between sites and the natural environment, remained in the background of archaeological research until the rise of cultural ecology in the 1930's - 50's (Steward 1955). Before that, when questions of human/environmental relationships were addressed, they tended to focus on the correlation between "culture types" and broad regional environments (Kohler 1988:26). Steward pioneered a new approach, by stressing the effects of local environmental conditions on specific aspects of culture and by identifying which cultural features were most closely linked to the environment via subsistence activities (i.e., the culture core). Steward and Setzler (1938) encouraged archaeologists to concentrate more on subsistence and settlement patterns rather than the traditional emphasis on artifact styles (Trigger 1989:279). This argument bore fruit after World War II with the rise of settlement archaeology; a field largely defined by Willey's (1953) research into settlement patterns in the Viru' Valley of Peru (Kohler 1988; Trigger 1989). This, and subsequent projects (cf. MacNeish 1978; Struver 1968), resulted in an increased emphasis on the study of site setting and its ecological relationship to human behavior (Trigger 1989). Settlement pattern studies became increasingly more sophisticated in the 1970's when researchers began to incorporate rigorous statistical methods into ecologically oriented site locational models (cf. Euler
and Gummerman 1978; Plog and Hill 1971). This emphasis on statistical analysis led to the "Era of Predictive Modeling" during the late 1970's and 1980's (Kohler 1988:33-35). Today predictive modeling is an integral component of archaeological research. The earlier models tended to focus primarily on predicting the location of past human activity on the basis of recognizable correlations between site location and various natural resources considered critical to human settlement (i.e., food, water, shelter, fuel). While a valid approach under ideal field conditions, most predictive models formulated in recent years have recognized the need to first take into account site formation processes: which as demonstrated by Schiffer (1987), Wood and Johnson (1978), and others, can seriously alter the integrity and visibility of archaeological sites (cf. Ebert and Kohler 1988:125; Thoms 1988:639). As a result, independent variables are expanded to include those that are relevant to questions of landscape deposition, erosion and stability. In addition, many contemporary modeling efforts are either preceded by, or are actually based on, geomorphological, pedological, and/or stratigraphic field research (cf. Artz 1985; Bettis and Benn 1984; Gardner and Donahue 1985). In a landscape as dynamic as the Little Missouri Badlands, geologically-oriented research is a unavoidable prerequisite to the building of any models related to the question of site setting. The results of such research in the THRO South Unit will be presented in Chapters V and VI. First, however, it is appropriate to simply describe where known archaeological sites are situated.

Definitions

The location of the South Unit sites is summarized in terms of their landform setting and association with sedimentary depositional environments. Other frequently studied natural variables such as distance to water, slope, elevation, and vegetation, have little utility in highly eroded landscapes because they reflect the modern configuration of the environment and not necessarily conditions as they existed in the distant prehistoric past. For instance, hillslope erosion can significantly alter the location of springs and seeps by diverting subsurface discharge; while piping, gullying, and incongruous slopewash deposition can alter surface runoff, subsurface flow, and stream base levels (cf. Bryan and Yair 1982; Campbell 1989; Schumm and Hadley 1957). Therefore, the proximity of an archaeological site to modern water sources may not accurately reflect the proximity to water when the site was occupied. The same possibility holds true for the location of plant communities, slope, land surface elevations, and view. On the other hand, extant landforms and sedimentary deposits are palpable features of the environment that can be classified, dated, and used to interpret the stratigraphic and paleoenvironmental history of a region (cf. Waters 1992:91).
As described in Chapter II, the principal landform categories in the badlands are ridges, buttes, knobs, foothills, terraces, active floodplains, stream channels, gullies, and alluvial fans. In addition to each of these, sites were also located on the Little Missouri Escarpment, hillslopes, toe slopes, and slopewash-filled valley bottoms. The Little Missouri Escarpment is the name given to the prominent slope marking the boundary between the Little Missouri Badlands and the uneroded Missouri Plateau (Jacob 1976). It runs along the eastern and southeastern portion of the THRO South Unit (Figure 5). All of the upland landforms in the study area such as buttes and ridges are remnants of the Missouri Plateau, therefore the top of the Escarpment is also considered part of the uplands. Hillslopes are inclined surfaces along the sides of landforms, particularly ridges, buttes, knobs, and foothills. They are frequently comprised of eroded Fort Union Group bedrock, although some have mantles of slopewash sediments and pre-badland fluvial gravels. For the South Unit study area, toe slopes are described as landforms located at the distal end of a slope. Sediments in toe slope settings are usually transported by mass movement over the slope surface (cf. Ritter 1978:156-158; Selby 1985:230-238), however deposition by subsurface water movement has also been reported (Dalrymple et al. 1968). As illustrated in Waters (1992:231), mass-movement, or colluvial, processes include slumps, flows, slides, creep, and falls. Finally, the term valley bottom refers to the floor of pre-existing valleys that currently lack active streams. Sediments in these landforms have been deposited primarily by slopewash, a process in which sediment is "...moved on the surface of slopes by ... overland flow (wash)" (Ritter 1978:158).

Although prehistoric sites are located on all of the above-mentioned landform categories, the sediments associated with the sites vary within landforms. For example, sites on ridgetops are associated with eolian loess deposits, eolian sand dunes, fluvial lag gravels deposited by pre-badland streams, and Fort Union Group bedrock. Likewise, sites on eroded foothills are associated with slopewash sediments, pre-badland gravels, and bedrock. For this reason, a simple description of the landform setting of the South Unit sites would tend to oversimplify or incorrectly identify the depositional environment associated with the site matrices. This, in turn, could hamper accurate identification of noncultural site formation processes. Therefore, the following discussion on site location includes a description of: (1) simple landform setting; (2) sediment type and depositional environment; (3) landform/sediment associations; and (4) the setting of sites assigned to named cultural groups or general time periods. Categories 2 and 3 are based on the observed association between site cultural materials and their primary sedimentary matrix.
Results

The landform setting and sedimentary associations of the South Unit sites are summarized in Table 4, while the setting of the culturally and temporally diagnostic sites are illustrated in Figures 10 and 11. An examination of these data reveals an archaeological record that is highly disparate in terms of the frequency of settings represented.

**Landform.** Ridgetops are by far the largest landform category, accounting for 68% (n = 121) of the recorded sites (Table 4). The other upland landforms (buttes, knobs, Little Missouri Escarpment) comprise an additional 9% of the site assemblage, bringing the number of upland sites to 137, or 77% of the site total (Table 4). Lowland landforms, on the other hand, account for only 14% of the site settings (n = 25). These landforms include fill terraces, cut terraces, alluvial fans, active floodplains, toe slopes, stream channels/gullies, and valley bottom (Table 4). The remaining landform categories, foothills and slopes, are intermediate in elevation between the uplands and lowlands and account for 9% of the site aggregate.

The unequal nature of landform setting was evident when sites from both the North and South Units were statistically analyzed (Kuehn 1990:181-190). The landforms associated with the recorded prehistoric sites (n = 214) and with a randomly generated assemblage of non-site loci (n = 106) were compared using Chi-square and discriminant function analysis. In both cases, the analyses demonstrated a non-random distribution in the landform setting of sites. Pooled-within group correlations identified landform as the variable which contributed most to significant discrimination between the sites and non-site loci (a pooled within-group correlation score of .66790 for landform, as compared to .64472 for soils, .49854 for view, .40271 for slope, and .24184 for distance to water). Likewise, chi-square tests comparing the landform setting of sites with those of the non-site loci produced significant results at the p = 0.0001 level (chi-square = 146.83 at 10 df) (Kuehn 1990:185-190).

**Sediment Type and Depositional Environment.** Of the 178 recorded South Unit sites, a total of 129 (72.5%) are associated with loess, or wind blown silt. This high percentage is not surprising considering the fact that loess deposits cap many of the upland landforms in the region (Kuehn 1993). A second form of eolian sediment, dune sand, is associated with 1.1% of the sites (n = 2, Table 4). The dunes are also located in the uplands (i.e., on ridgetops), where they originated from.
Table 4. Summary of Landform Setting and Sedimentary Association, All Prehistoric Sites, THRO South Unit.

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<th>Sedimentary Association</th>
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the weathering of exposed sandstone bedrock. Fluvially-deposited sediments are associated with only 12.4% of the sites (n = 22). These consist of pre-badland channel lag gravels (n = 6), modern channel lag gravels (n = 1), terraces (n = 12), and active floodplains (n = 3) (Table 4). The pre-badland lag gravels, of Miocene/Pliocene and Pleistocene age, are located primarily on upland landforms, but also occur on intermediate landforms such as slopes and foothills. The remaining fluvial deposits are concentrated in lowland settings. Colluvial sediments (i.e., slump deposits) are associated with only one recorded site (0.6%), while slopewash/alluvial fan deposits comprise 11.2% of the site aggregate (n = 20, Table 4). The final sediment category, poorly indurated bedrock, is associated with 2.2% of the sites (n = 4). This sediment, as discussed in Chapter II, is Paleocene in age and was deposited in marginal-marine and fluvial environments.

**Landform/Sediment Associations.** A total of 115 ridgetop sites are associated with loess deposits (64.6%), while three are associated with pre-badland channel lag gravels (1.7%), two with sand dune deposits (1.1%), and one with Paleocene bedrock (0.5%). Three sites (1.7%) are located on butte tops, all in association with loess deposits (Table 4). The other upland landforms include two sites on knobs associated with pre-badland channel lag gravels, and 11 sites (6.2%) on the loess-capped Little Missouri Escarpment (Table 4).

Terrace sites include five on T2 fill terraces (2.8%), five on T3 fill terraces (2.8%), and one on a T4 fill terrace (0.5%). All 42 are associated with overbank, or vertical accretion, sediments (Table 4). An additional site was located at the top of the T3 fill terrace, but was associated with slopewash sediments. A final site was located near the top of a cut, or strath, terrace tread in association with a thin cap of slopewash sediment. The cut terrace in this instance is comprised of a flat bench that was excavated into Paleocene bedrock by the Paddock Creek channel when it occupied the T4 base level. Other sites associated with fluvial landforms include one located in an ephemeral stream channel and associated with modern channel lag gravels, and three on the T1 terrace of the Little Missouri River (Table 4). All three of the Little Missouri sites are associated with overbank sediments.

Additional sites in lowland settings include six (3.4%) on alluvial fans (i.e., slopewash sediments), one in the bottom of a valley covered with slopewash sediments (i.e., valley bottom), and one in colluvial toe slope (i.e., slump) deposits at the base of the Little Missouri Escarpment (Table 4). Sites in intermediate settings include 11 on eroded foothills and five on slopes. Nine of the foothill sites are associated with slopewash sediments, one is in association with pre-badland
channel lag gravels, and one Paleocene bedrock (Table 4). The sites on slopes include three in slopewash sediments, and two resting directly on Paleocene bedrock (Table 4).

**Culturally and Temporally Diagnostic Site Setting.** The final aspect of site setting to be addressed is the landform and sedimentary context of those sites assigned to named cultural groups or temporal periods.

The oldest components (Paleoindian, Logan Creek/Mummy Cave, and McKeans), are all located on upland landforms in association with loess deposits (Figures 10 and 11). In contrast, the percentage of Pelican Lake components in similar settings decreases to 40%, while intermediate landforms appear for the first time, comprising 60% (Figures 10 and 11). Pelican Lake components also exhibit significant diversity, with loess and slopewash deposits each comprising 40% of the site total, and Paleocene bedrock 20% (Figure 10). The oldest recognized lowland component dates from the Late Archaic period. Late Archaic sites are evenly divided between upland and lowland landforms (and between loess and vertical accretion deposits). Somewhat surprisingly, Besant components are all located in upland loess settings (Figure 10). Plains Village components demonstrate a marked increase in lowland settings, which actually surpass all other landform categories for the first time. Likewise, vertical accretion deposits are the dominant sediment type at the Plains Village sites (Figure 10). The highest degree of diversity is evident in those components assigned to the Late Prehistoric period, which are located in all three landform categories and in association with four of the six principal sediment types (Figure 10).

In summary, the cultural/temporal data indicate an increase in the frequency of lowland settings through time, which in turn, is concomitant with an increase in the number of components associated with fluviatile (i.e., vertical accretion) and slopewash sediments. These locational patterns have been previously recognized by other archaeologists working in the Little Missouri Badlands (cf. Beckes and Keyser 1983; Hill 1988; Kuehn 1990, 1993; Loendorf et al. 1982; Simon 1982), and will form the basis for much of the discussion in Chapter VII.
Figure 10. Graph illustrating the landform/sedimentary association of sites assigned to specific cultural groups or temporal periods.

Figure 11. Graph illustrating the landform setting of sites assigned to specific cultural groups or temporal periods.
CHAPTER V
LATE QUATERNARY STRATIGRAPHY

Introduction

The previous chapter was concerned with identifying the functional, chronologic, and locational aspects of archaeological sites in the THRO South Unit. Attention now turns to the natural environmental context of the archaeological record, specifically to its geological context, in an attempt to identify significant noncultural site formation processes. As argued in Chapter 1, the elucidation of formation processes is a necessary prerequisite to the drawing of meaningful archaeological inference in the badlands region. It was further argued that the transformation of artifacts and ecofacts from systemic to archaeological contexts (Schiffer 1972, 1987) is a process that can best be studied through the systemic approach inherent in contextual archaeology (Butzer 1982; Schoenwetter 1981; Stein and Farrand 1985). Stratigraphy, the study of vertical and lateral relationships between sediments on the basis of lithologic, chronologic, and other characteristics (Boggs 1987:522; Krumbein and Sloss 1963), is a key geological element in this contextual approach. Through stratigraphy, important aspects of the earth's natural history, particularly geomorphology and paleoenvironmental conditions, are revealed (Reineck and Singh 1980:4-6). Sedimentary deposits envelop data of importance to a wide spectrum of scientific disciplines, including geomorphology (e.g. lithology, numerical dating samples, unit contacts); paleoclimatology (e.g. fossil pollen, macrofloral remains, faunal remains, stable carbon isotopes); and pedology (e.g. buried soil horizons, soil structure, soil texture). Stratigraphy provides the framework which makes the integration of these various data sets possible (Boggs 1987:13). Sedimentology and stratigraphy are particularly important to the study of archaeological site formation for several reasons: (1) they help establish a temporal framework for archaeological materials (Gifford and Rapp 1985; Gladfelter 1981; Hassan 1979); (2) they aid in the reconstruction of landscapes at the site or regional level (Albanese 1974; Bettis and Benn 1984; Waters 1988); and (3) they aid in the reconstruction of the geologic or landscape history of sedimentary deposits, including archaeological sediments (Stein 1985).

The late Quaternary stratigraphy of the South Unit study area was investigated by: (1) identifying the major depositional environments represented in the sedimentological record; (2)
determining the sequence of sediment deposition, erosion, and landscape stability at a number of representative study sections; (3) determining the age and chronology of depositional/non-depositional sequences; and (4) determining the lithological and temporal relationships between major stratigraphic units (cf. Reineck and Singh 1980:4-6; Waters 1992:61).

The discussion of depositional sedimentary environments within a given geomorphological landscape first requires an understanding of the relationship between process and response (Boggs 1987:306-307). Sedimentary environments, which include both static and dynamic natural elements (i.e., process elements), are responsible for the production, transportation, and deposition of identifiable bodies of sediment. These bodies of sediment, termed sedimentary facies, have distinctive sets of characteristics (Boggs 1987:306). In the South Unit study area there are three major categories of late Quaternary depositional environments: eolian, fluvial, and slopewash/alluvial fan. Lacustrine and colluvial environments, on the other hand, are relatively minor elements in the stratigraphic record. In terms of landform distribution, late Quaternary sediments in the uplands were deposited predominately by eolian processes, while sediments in the lowlands were deposited predominately by fluvial and slopewash processes. Lacustrine sediments were only encountered at one location (in the uplands) and will therefore be described along with the eolian deposits.

Eolian and Lacustrine Depositional Environments

The most common late Quaternary eolian facies in the South Unit study area are loess and dune sands. Of the two, loess is by far the most common form of wind transported sediment. Much less common are lacustrine (pond) facies, which were identified only at a single study section. The following discussion on eolian and pond depositional environments centers on the identification and description of a number of key stratigraphic sections, all of which are located in upland settings.

As used here, the term loess, or unconsolidated, wind-deposited silt, is not necessarily limited to sediment produced by glacial activity, although such an origin is possible for some of the older units. Significant accumulations of loess are extant on upland landforms, particularly buttes and ridges, where they mantle Fort Union Group bedrock or pre-badland channel lag gravels (which in turn overlie bedrock). On intermediate landforms such as slopes and eroded foothills, loess is virtually non-existent; a factor apparently related to high erosion rates, the predominance of slopewash processes, and topographic conditions. Thin loess deposits do appear to cap some fluvial
terraces, particularly those that are not associated with slopewash facies. One such example is Locality N in Paddock Creek (Figure 4), where a 40 cm-thick surficial silt unit overlies three superimposed buried soils formed in clayey silt overbank deposits. The silt unit was associated with two previously mentioned Late Prehistoric hearth features (32Bl579) dated at 1390 and 1180 yr B.P. An eolian origin for the unit is indicated by its texture, massive sedimentary structure, and stratigraphic position. A second example of a loess-capped terrace is evident at archaeological site 32Bl568, situated near the top of the T2 terrace in Paddock Creek (Figure 5). Here Plains Village ceramics were observed in a massive silt unit approximately 16 cm thick. A third example was identified on a T2 Little Missouri River terrace on North Dakota state lands in the southern portion of the badlands (Borchert and Wermers 1994). In spite of these examples, loess deposits in non-upland areas are rare. The causal factors behind this paucity are likely to include the dynamic nature of geomorphic processes in lowland areas, and the predominance of other types of sediment deposition (i.e., fluvial, and slopewash/alluvial fan). The flat-topped uplands, on the other hand, are geomorphologically stable and have not experienced significant amounts of fluvial and slopewash deposition. They also appear to have topographic characteristics that are favorable to the trapping of eolian sediment (i.e., flat to undulating surfaces with occasional small depositional basins).

All loess (and sand dune) deposits mantling the South Unit uplands are assignable to the eolian lithofacies of the late Quaternary Oahe Formation. As mentioned, the Oahe was first defined to include eolian silt deposited on gently sloping upland surfaces throughout North Dakota (Clayton et al. 1976). It was later redefined and expanded to include all late Quaternary sediment overlying the Coleharbor Group, regardless of depositional environment (Clayton and Moran 1979). The Coleharbor Group is the lithostratigraphic unit encompassing all Pleistocene glacial, glacio-fluvial, and glacio-lacustrine sediment in North Dakota (Bluemle 1977; Clayton et al. 1980; Moran et al. 1976).

The Oahe Formation is comprised of three "lithogenetic subdivisions", or lithofacies, reflecting sediment deposition in fluvial, pond, and eolian environments (Clayton and Moran 1979; Clayton et al. 1980:76-77). The Oahe is subdivided into four members on the basis of lithology, pedology, and stratigraphic setting (Clayton et al. 1976). Most of the physical, chemical, and biological components which characterize the individual members were identified during studies of the eolian lithofacies, which is our primary focus here. From oldest to youngest, the Oahe members are the Mallard Island, Aggie Brown, Pick City, and Riverdale. The Mallard Island Member generally overlies sediment of the Coleharbor Group, except in the unglaciated southwestern portion
of the state where its existence has yet to be demonstrated. Where present, it is the most sandy and dolomitic of the four units and contains generally low amounts of organic matter. The age of the Mallard Island Member is estimated at ca. 13,000 - 14,000 yr B.P. (Clayton et al. 1976:11).

The Aggie Brown Member overlies the Mallard Island and has two defined submembers. The lower has a distinctive reddish color, while the upper is very dark, and represents an A1 horizon of a buried soil named the Leonard Paleosol by Bickley (1972). The Aggie Brown sediments, while variable in texture, consist primarily of silt (72%), with lesser amounts of clay (19%) and sand (9%) (Clayton et al. 1976:5). The age of this member is estimated as late Pleistocene to early Holocene, or from 13,000 to 8500 yr B.P. (Clayton et al. 1976:11). This estimate has since been substantiated by radiocarbon ages on samples recovered from archaeological sites in the Knife River flint primary source area (Root et al. 1986), and at other locations in North Dakota (Artz 1992).

The Pick City Member overlies the Aggie Brown and is middle Holocene in age (i.e., 8500 - 5000 yr B.P.). It is the lightest-colored member of the Oahe, a factor which reflects its low organic matter content (Clayton et al. 1976: Clayton et al. 1980). Mean sample grain sizes from the Pick City at the Riverdale type section are silt (77%), clay (14%), and sand (7%) (Clayton et al. 1976:6). Paleosols have not been previously reported from the Pick City member (Artz 1992; Clayton et al. 1976; Root et al. 1986).

The Riverdale Member is late Holocene in age (ca. 5000 yr B.P. to present). The member was described as having three submembers (Clayton et al. 1976:6). The upper and lower submembers are soil A horizons, the latter having been named the Thompson Paleosol by Bickley (1972). At the type section, mean grain size in the Riverdale is similar to that of the underlying Pick City, with silt comprising 80%, clay 12%, and sand 8% (Clayton et al. 1976:6).

The classification and chronological framework of the Oahe Formation, particularly the eolian lithofacies, has proven to be of tremendous value to Quaternary geologists, geomorphologists, and archaeologists (Artz 1992). Nevertheless, information on the lithology, pedogenic composition, and spatial distribution of the unit is somewhat limited, having been derived almost exclusively from the Missouri River/Lake Sakakawea region in west central North Dakota (Bluemle 1971; Clayton et al. 1976; Reiten 1983), and from the adjacent Knife River flint primary source area in Dunn and Mercer Counties (Artz 1992; Artz and Ahler 1989; Root and Ahler 1987; Root et al. 1986). In southwestern North Dakota, previous research into the Oahe Formation was restricted to a handful of archaeological excavations conducted in upland settings (Jorstad et al. 1986; Metcalf et al. 1988; Wyckoff 1982). As a result, little is known about the temporal and spatial distribution, depositional
environment, and post-depositional alteration of Oahe sediments in the Little Missouri Badlands region. In spite of the limited research, some important stratigraphic and pedologic information has been made available. For example, Jorstad et al. (1986) documented the presence of between nine and 12 paleosols in unconsolidated eolian deposits on an extensive ridgetop in the north-central portion of the badlands (i.e., Cinnamon Creek Ridge). Most of the paleosols consist of thin accumulations of organic material with some evidence of carbonate leaching (Jorstad et al. 1986:165). Radiocarbon ages on soil humic acids range from 5390 ± 100 yr B.P. (SI-6160) to 530 ± 55 yr B.P. (SI-6163). These ages indicate that the soils are synchronous with the Upper and Lower Thompson Paleosols (Jorstad et al. 1986:179). Additional data on the lithology and pedogenesis of Oahe Formation sediments in the badlands are presented in Floodman et al. (1983), Kuehn et al. (1987), Metcalf (1984); Simon and Keim (1983), and Wyckoff (1982). These investigations have not encountered deposits older than ca. 5400 yr B.P. (Artz 1992; Kuehn 1992:176) and are therefore limited to the Riverdale Member. In each case, the Riverdale was found to directly overlie Fort Union Group bedrock. The paleoclimatic implications of this stratigraphic pattern will be addressed in Chapters VI and VII.

Description of Upland Study Sections

The stratigraphic investigation of eolian and lacustrine deposits in the South Unit centered on the examination of 15 study sections. At least one section was examined on every major upland landform in the study area. Six of the localities are associated with one or more radiocarbon ages (Table 5). The relative "completeness" of the sections vary widely. Some sections contain only one of the Oahe Formation members, some contain two, and others three. The thickness of individual members, their post-depositional history, and their stratigraphic relationships with other members is also extremely diverse. While the eolian deposits are therefore spatially and temporally inconsistent, a generally thin mantle of loess is present on most of the principal upland landforms in the study area. These include Big Plateau, Petrified Forest Plateau, Petrified Forest Ridge, Radio Tower Plateau, Johnson Plateau, Peck Hill, Boicourt Ridge, and the Little Missouri Escarpment (Figure 5). The following discussion of eolian and lacustrine depositional environments will focus on: (1) two sections which are the most well documented and complete in terms of sediment age, thickness, and pedogenic development; (2) three additional sections containing dated units associated with the Aggie Brown, Pick City, and Riverdale members; and (3) a number of largely undated sections where only the Pick City and/or Riverdale Members are present. Stratigraphic unit
Table 5. Summary of Radiocarbon Data.

<table>
<thead>
<tr>
<th>Locality</th>
<th>Stratigraphic Unit</th>
<th>Radiocarbon Age (yr. B.P.)</th>
<th>PDB (%)</th>
<th>Lab #</th>
<th>Material</th>
<th>Remarks</th>
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<tr>
<td>A</td>
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<td>10.480 ± 82</td>
<td>AA10538</td>
<td>Charcoal</td>
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<td>Sand Dune</td>
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</tr>
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<td>O</td>
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</tr>
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<td>O</td>
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<tr>
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**LACUSTRINE DEPOSITIONAL ENVIRONMENTS**

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<th>Material</th>
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**FLUVIAL DEPOSITIONAL ENVIRONMENTS**

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<td>A</td>
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<td>D</td>
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<td>*6115 ± 14500</td>
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<td>A-7113</td>
<td>Charcoal</td>
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<td>F</td>
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<tr>
<td>G</td>
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<td>150 ± 60</td>
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<td>H</td>
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<td>T1 Terrace (Hamilton 1967)</td>
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<td>Jones Creek</td>
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<td>185 ± 70</td>
<td>l-2325</td>
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* Δ 13C corrected
Table 5. Continued.

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<tr>
<td>I</td>
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<td>$930 \pm 90$</td>
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<td>-18.7</td>
<td>A-7111</td>
<td>Bulk Soil</td>
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* $\delta^{13}C$ corrected
designation at all of the upland sections reflects correlation with the three principal Oahe Formation members (i.e., Unit I = Aggie Brown Member, Unit II = Pick City Member, Unit III = Riverdale Member). Individual soils within units, however, do not generally correlate from locality to locality and are therefore simply designated S1, S2, S3, and so on, according to their location within the section.

**Locality O.** Locality O is located at the southwestern end of Petrified Forest Ridge (Figure 4). Here, a 3.4 m thick sequence of very fine to medium silt was laid down in a shallow basin near the apex of the ridge (Figure 12). The small basin covers an area of approximately 3200 square meters, and appears to have been formed by the action of wind deflation on soft sandstone bedrock. The locality contains thirteen distinct stratigraphic units, identifiable on the basis of bounding discontinuities (NACOSN 1983), in this case, the upper boundaries of buried soils (Figure 13). Unit 1a, the lowest at 320 - 340 cm below the surface, is a medium-thick bed of very fine silt unconformably overlying Paleocene-aged sandstone of the Sentinel Butte Formation (Carlson 1983:Plate 1). Onto the upper portion of this unit a thin dark A horizon (designated S1) developed. This soil was classified in the field as an Entisol because of its simple A horizon profile. The A horizon has a highly irregular boundary which suggests erosion after the brief period of pedogenesis (Figure 14).

Unit 1b, a thick bed of massive, very fine silt, overlies Unit 1a and is capped with a prominent soil (S2). Three disparate radiocarbon ages were obtained from the soil (Table 5). The first, 9330 ± 100 yr B.P. (Beta-27927), was obtained on bulk soil organics. This age is rejected because Unit 1b is overlain by paleosols that yield radiocarbon ages substantially older than 9300 years. As a result, the mean residence time soil-age appears to be too young. The second age, 13,070 ± 490 yr B.P. (Beta-33060) on charcoal, is suspect because the sample consisted of only 0.3 g of clean carbon after pretreatment, and was therefore given quadruple the normal counting time (Jerry Stipp, personal communication, 1989). The third age, 11,070 ± 280 yr B.P. (A-7112) on charcoal, is accepted because the amount of useable carbon after pretreatment was substantially larger than the Beta-33060 sample (1.0 g carbon equivalent). Also, the University of Arizona samples were corrected for carbon isotope fractionation, while the Beta samples were not. Therefore, Unit 1b appears to be late Pleistocene in age, and is equivalent to the Aggie Brown Member of the Oahe Formation in terms of temporal, lithostratigraphic, and pedostratigraphic composition (Clayton et al. 1976). The Unit 1b soil, which correlates with the Leonard Paleosol, has
Figure 12. Photograph of Locality O surface area to southeast.

Figure 13. Photograph of Locality O stratigraphic profile.
Figure 14. Stratigraphic profile of Locality O, showing the location of sediment samples and corresponding $\delta^{13}$C values, radiocarbon ages, and soils.
an A-AB-Btk genetic profile. The soil contains a mollic epipedon, as evidenced by dark color (10YR 3/2 moist, 10YR 5/3 dry), A horizon thickness (greater than 25 cm), base saturation (greater than 50%), and organic carbon content (1.58%) (Table 6). Although classifying buried soils can be tentative because of compaction, truncation, and ambiguous climatic regimes, the S2 soil appears to be either a Typic Cryoboroll or a Typic Haploboroll, if temperature and moisture estimates presented in Clayton et al. (1976:11) for this time period are accurate (i.e., cryic or frigid temperature regime and udic moisture regime). The S2 soil contains the highest percentage of organic carbon and lowest percentage of inorganic carbon of any of the soils at Locality O (Tables 6 and 7).

Units Ic through Ig are all medium-thick beds of very fine silt on which weakly developed soils (S3-S7) exhibiting simple A profiles (i.e., Entisols) developed. The A horizons (Figures 13 and 14) are less than 5 cm thick and are similar to the so-called "dark bands" common in loess deposits throughout the mid-continent (Ruhe 1976, 1983). Again, following climatic data provided by Clayton et al. (1976:11), the S3 through S7 soils are tentatively classified as either Typic Cryorthents of Typic Udorthents.

A slightly more well-developed soil with an A-Bw profile (S8), is evident at the top of Unit Il, a medium-thick bed of fine silt (Figure 14). The soil organics from the A horizon of soil 8 yielded radiocarbon ages of 11,560 ± 110 yr B.P. (Beta-33061) and 10,730 ± 460 yr B.P. (GX-15397). Given the large standard deviation associated with the 10,730 B.P. date, the two ages technically overlap, a factor which tends to confuse the chronologic sequence in the lower portion of Area O. This is not surprising, however, considering the inconsistencies and wide temporal scales associated with the dating of soil organics (cf. Holliday 1990). Nevertheless, the 10,730 ± 460 yr B.P. age from Unit Il is more or less chronostratigraphically consistent with the ages obtained from adjacent units. This suggests that Unit Il was deposited rapidly, and that the S8 period of pedogenesis occurred somewhere near the Pleistocene/Holocene boundary. The S8 soil is classified as an Inceptisol, possibly either a Typic Cryumbrept or a Typic Haplustrept.

Unit Il is a thick bed of very fine silt on which an A horizon soil developed. Although undated, its age is bracketed by the early Holocene/late Pleistocene date obtained from Unit Il and by a radiocarbon age on soil humates of 6580 ± 160 yr B.P. (A-7109) from overlying Unit II (Figure 14).
Table 6. Locality O - Summary of Data from Texas A&M Soil Characterization Laboratory.

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Unit II is a thick bed of very fine silt, and is significant for a number of stratigraphic, pedologic, and paleoenvironmental reasons. The unit is dominated by a soil exhibiting a genetic Bk profile (Figure 14). The lack of an A horizon, and the irregular (i.e., concave) upper boundary of the Bk horizon indicates erosional truncation after the pedogenic episode (cf. Boggs 1987:527). The horizon is carbonate-impregnated almost to the point of being indurated (many small nodules) and in general meets the criteria for stage III carbonate accumulation (cf. Gile et al. 1966). The horizon is very pale brown (10YR 7/3 dry) but with many, fine, prominent white mottles (10YR 8/2 dry). Laboratory analysis indicates that the horizon has the lowest percentage of organic carbon (0.38% and 0.44%) and the highest percentage of inorganic carbon (1.67% and 1.5%--CaCO₃ equivalent of 12.5) of the entire section (Tables 6 and 7). The horizon also exhibits a significant decrease in mean particle size compared to the overlying units (.0057 mm in Unit II and .0170 in Unit IIIa)(Table 6). The soil appears adequately drained, therefore the abundant carbonates suggest that pedogenesis occurred under conditions of reduced precipitation (cf. Leopold and Miller 1954; Reider 1980, 1990). The absence of an A horizon prevents legitimate classification, although the mesic temperature and aridic or torric moisture regime hypothesized by Clayton et al. (1976) for this time period suggest that it could be a Typic Haplumbrept, Typic Eutrochrept, or Typic Camborthid (Soil Survey Staff 1990).

A radiocarbon age of 6580 ± 160 yr B.P. (A-7109) was obtained on soil organics from the Unit II soil. This, together with stratigraphic position and the heavy accumulation of carbonates in the Bk horizon, suggests that Unit II was deposited prior to ca. 6600 yr B.P. and is assignable to the Pick City Member of the Oahe Formation. While acknowledging past problems associated with such designations (cf. Holliday 1989a:200), the S10 soil does appear to represent one of the so-called "Altithermal soils" so commonly reported in the west and southwestern portions of the United States (Albanese 1986; Haynes 1968; Leopold and Miller 1954; Reider 1980, 1990). The S10 soil, hereby informally designated as the "Medora" soil, is among the first Altithermal soils to be identified in North Dakota.

The remaining units at Locality O are late Holocene in age, as evident by bulk humate radiocarbon ages of 4070 ± 130 yr B.P. (A-7108) and 2160 ± 70 yr B.P. (Beta-33062) on samples collected from Units IIIa and IIIb. Both units have buried soils exhibiting A-Bk profiles (S11 and S12), and both are tentatively classified as Typic Haplumbrepts. The S11 soil appears equivalent to the lower Thompson Paleosol of the Riverdale Member, while the S12 soil remains unclassified (Clayton et al. 1976). It is important to note that pedogenic features such as visible structure and
carbonate accumulation are not present at the bottom of Unit IIIa, which appears to be unaltered parent material. The S12 soil, however, is imprinted onto the top of Unit IIIa (Figure 14).

The youngest unit, Unit IIIc, is late Holocene in age. The modern surface soil, with a weakly developed A-Bw profile, has developed on this unit. The soil is designated in the old soil survey manual as Bainville Loam, hilly phase series (Edwards and Ableiter 1944). It is classified as a Typic Haplumbrept. As previously discussed, the upper two units at Locality O contain several Late Prehistoric cultural components, while Unit II contains a possible Early Archaic, Logan Creek/Mummy Cave component (site 32BI703).

Locality A. A second significant and temporally extensive sequence was investigated at Locality A on Big Plateau (Figure 4). The section contains eight eolian and two lacustrine stratigraphic units (with a combined thickness of 3.5 m). The base of the section is composed of gravels that were deposited by a channel of the Little Missouri River. These gravels unconformably overlie Paleocene clay of the Fort Union Group (Figure 15). The fluvial gravels yielded wood fragments that were radiocarbon dated to 24,230 ± 1510/1270 yr B.P. (A-7114). Because Locality A lies 90 m above the present floodplain of the Little Missouri River, this age suggests that the Little Missouri River has downcut over 90 vertical m within the last 24,000 years. The geomorphological and climatological implications of this important date will be discussed in the section on fluvial stratigraphy and in Chapter VI.

The Pleistocene gravels are unconformably overlain by Unit Ia, a thick bed of very fine silt (Figure 16). This unit is massively bedded, but highly disturbed by rodent activity. The central portion of the unit contains a thin, but broken and irregular, dark band (Figure 15). While reminiscent of the thin A horizons so common in mid-continent loess deposits (Ruhe 1976, 1983), the band in Unit Ia was not designated as a buried soil because of its discontinuous morphology. Nevertheless, it could indicate a brief pause in deposition. The upper boundary of the unit is marked by a distinctive erosional disconformity (Figure 15). The depositional environment associated with Unit Ia has yet to be determined, although texture and sedimentary structure tend to indicate an eolian source (i.e., massive silt).
Figure 15. Stratigraphic profile of Locality A, showing the location of sediment samples, corresponding δ¹³C values, radiocarbon ages, and soils.
Figure 16. Vertical sequence of stratigraphic units at Locality A showing mean phi values.
Unit 1a is overlain by a 22 to 35 cm thick unit of fine silt, very fine silt, and clay (Unit 1b) that appears to represent deposition in a pond/lacustrine environment (Table 8; Figure 15). As illustrated in Figure 16, the basal portion of Unit 1b is a coarse clay which yielded a charcoal radiocarbon age of 11,830 +340/-330 yr B.P. (A-7115). A lacustrine rather than eolian origin for the deposit is suggested by the dramatic decrease in grain size in the central portion of the unit. While the sample collected from the unit base (C8001) is 8.68 mean Phi, the sample from the unit center (C8000) is 11.14 mean Phi with 83.4% clay (Table 8; Figure 16). Grain size then increases dramatically up the section to 6.88 Phi in sample C7999 and 7.46 Phi in sample C7998 (Table 8; Figure 16). This fining and then coarsening upward sequence is not associated with visible stratigraphic contacts. The sudden decrease in grain size in the center of the unit, however, is representative of a lithic discontinuity (Soil Survey Staff 1975). As a result, the unit was divided into two subunits (1b1 and 1b2)(Figures 15 and 16). The lower subunit (1b1) corresponds to the upward fining portion of the sequence which is unaltered parent material, while the upper subunit (1b2) corresponds to the coarsening upward portion of the sequence which has been affected by pedogenesis (Figure 16; Table 8). The soil in subunit 1b2 (S1) exhibits a simple Ak profile (Entisol). Both the A horizon and the unaltered parent material contain abundant carbonates, visible as coalesced streaks and few to common small nodules (i.e., stage II). Classification of the soil is difficult and highly tentative because of the change in lithology and uncertainty over paleoclimatic conditions. The extremely high clay content and abundance of carbonates throughout the unit are indicative of a poorly drained, ponded environment (cf. Kelts and Hsu 1978; Langbein 1961). The high carbonate levels suggest saturation with carbonate-rich water and precipitation and/or capillary rise concomitant with evaporation (Soil Survey Staff 1975), while the high clay content (given the paleotopographic setting) suggests the settling of fine suspended sediment (eolian derived?) in a lacustrine environment (cf. Picard and High 1972, 1981). The soil, therefore, appears to be a poorly drained Entisol (i.e., Aquent), either an Typic Cryaquept or Typic Haplaquept, depending upon paleotemperature regime (Soil Survey Staff 1990).

Additional evidence in support of a lacustrine origin for Unit 1b includes the coarsening upward sequence evident at the top of the unit, and the concurrent increase in the percentage of inorganic carbon (from 0.96 to 1.47%) (Figure 17, Appendix C). Both factors could be indicative of regressive pond filling, with the coarser upper sediments reflecting the encroachment of pond-shore sediments onto finer pond-interior sediments, and the increase in inorganic carbon reflecting

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Figure 17. Locality A - Vertical distribution of organic and inorganic carbon.
the precipitation of soluble carbonates during evaporation (Boggs 1987:378; Fouch and Dean 1982; Picard and High 1981). Likewise, the radiocarbon age from Unit Ib is substantially older than any currently reliable ages on loess deposits in the region. In other words, the deposition of Unit Ib does not appear synchronous with significant deposition of Oahe Formation loess, which, as argued by Clayton et al. (1976) and Jorstad et al. (1986), is thought to have occurred during periods of dryer climate. It is therefore possible that Unit Ib was deposited during a wetter, (i.e., pluvial) climatic episode that preceded widespread eolian deposition. Although not evident on the surface, Locality A is situated in a small bedrock depression (i.e., less than 25 x 100 m surface area) that is visible along the eroding edge of the plateau. Because the locality is located on a Little Missouri cut terrace surface, the depression could have originated as a pool in the Little Missouri channel. A second possibility is that the depression formed as the result of wind deflation of the Unit Ia sediments. In any event, sometime after ca. 11,800 yr B.P. the pond dried and the landscape underwent a period of brief stability and pedogenesis. This episode was followed in turn by an episode of loess deposition and subsequent stability, as indicated by Unit Ic, the oldest stratum to be identified as eolian.

Unit Ic consists of a 20 cm thick deposit of medium silt that has been pedogenically altered into an A-Bw profile (Figures 15 and 16). It is late Pleistocene in age, as indicated by bracketing radiocarbon ages from Units Ib and Id (Figure 15). The S2 soil is tentatively classified as a Typic Haplumbrept or Typic Cryumbrept (Soil Survey Staff 1990:271).

Unit Id is a fining upward deposit of coarse silt to clay (Table 8; Figure 16). A weakly developed, but distinctive, soil horizon is present at the top of this unit. The dark-colored Ak horizon (10YR 3/2 wet) is comprised of 41% clay (mean Phi of 8.05) (Table 8; Figure 16). A charcoal sample from near the top of the A horizon produced an AMS radiocarbon age of 10,480 ± 80 yr B.P. (AA-10538). Both the A horizon and the unaltered parent material contain moderate to abundant carbonates visible as filaments and small nodules (i.e., stage II carbonate accumulation). The parent material exhibits a sudden downward increase in grain size to coarse silt (Table 8; Figure 16). Like Unit Ib, this textural change is abrupt enough to suggest a lithic discontinuity, although it does not appear to represent a significant interruption in sediment deposition (neither did the discontinuity in Unit Ib). The unit was therefore divided into two submembers (Id1 and Id2) rather than into two separate units (Figure 15). Pedogenesis is evident only in the upper subunit, but unlike Unit Ib, Unit Id exhibits a fining upward sequence, not fining and then coarsening. The soils in both units have been tentatively classified as Typic Cryaquents or Typic Haplusterts. The high clay
content in subunit Id2 (the S3 A horizon) and the presence of carbonates in both subunits, indicates deposition and evaporation/evapotranspiration in a second ponded environment. The simple fining upward sequence in grain size suggests that the pond may have been smaller, and the pluvial episode of shorter duration, than the one evident in Unit Ib. The sediment in subunit Id1, however, is sufficiently coarse (but yet well sorted) to suggest an eolian source prior to ponding. This possibility is supported by the lack of evidence for stream transport or slopewash deposition (i.e., a lack of coarse-grained sediment or diagnostic sedimentary structure). It is therefore hypothesized that
subunit Id1 represents a period of loess deposition that occurred after the S2 period of stability. As precipitation increased, loess deposition waned and a pond formed in the still extant (but less pronounced) paleodepression. Unit Id2 was then deposited sometime prior to 10,500 yr B.P. as dust dropped onto the surface of the pond settled to the pond floor. The supply of dust to the surface most likely increased along with the increase in rainfall, because rain is known to be an important agent in the deposition of air-born dust (Bagnold 1941; Greeley and Iversen 1985; Selby 1985:331). Shortly thereafter, the pond began to dry and carbonates in the water were allowed to precipitate out as the water evaporated. The drying of the pond was followed by a short-lived period of stability
and S3 soil formation, which as indicated by the radiocarbon age on charcoal recovered from the surface horizon, occurred around and after 10,500 B.P.. This was followed by at least five cycles of eolian deposition, pedogenesis, and erosion, and one cycle of deposition and stability. These events are evident in Units Ie through III (Figure 15).

Unit Ie is a medium bed of very fine to medium silt, while Unit If is a thick bed of very fine to coarse silt (Figure 15; Table 8). Both units are identified as loess deposits on the basis of particle size and massive sedimentary structure. Pedogenesis has altered the upper portion of Unit Ie into a Btk-Bk soil profile, while in the upper portion of Unit If, a Btk soil horizon is present (Figure 15).

The age of Units Ie and If have not been firmly established, however they are bracketed by the 10,480 B.P. age in Unit Id and by an age of 7255 ± 170 yr B.P. (A-7107) on soil humates from Unit Ilc (Figure 15). The presence of Btk horizons in both units indicate fairly substantial periods of pedogenesis after deposition, and precipitation sufficient to permit clay translocation. This suggests pedogenesis occurred under relatively mesic climatic conditions. In addition, stratigraphic position indicates that the units and soils are early Holocene in age (ca. 10,400 - 8,500 yr B.P.). They therefore appear temporally equivalent to the upper portion of the Aggie Brown Member (Clayton et al. 1976).
Units IIa and IIb are medium beds of very fine to medium silt, and medium silt, respectively (Figure 15; Table 8). Both are interpreted as loess deposits (based on the above mentioned criteria). The more or less identical units are characterized by light coloration (2.5Y 5/2), and the presence of truncated Bk soil horizons (Figure 15). They also contain the lowest percentage of organic carbon of the entire section (Figure 17; Appendix C). Unlike the soils in the two underlying units, the soils in Units IIa and IIb do not contain evidence of clay translocation, but instead, have abundant stage III carbonates, which are indicative of pedogenesis under conditions of reduced precipitation. These characteristics, together with their stratigraphic position between units that have yielded radiocarbon dates, suggest that Units IIa and IIb are middle Holocene in age and are assignable to the Pick City Member of the Oahe Formation (Clayton et al. 1976).

Unit IIc is a thick bed of medium to very fine silt. Like Units IIa and IIb, it is lightly colored, low in organic carbon, and high in inorganic carbon (Figure 17; Appendix C). The upper portion of the unit contains a Bk soil horizon (S8), conspicuous by the presence of stage III carbonates (visible as nearly indurated small nodules). Erosional truncation of the unit is evident by the absence of an A horizon, and by the upper boundary of the Bk, which is abrupt and irregular (Figure 15). As mentioned, humates from the S8 soil yielded a radiocarbon age of 7255 ± 170 yr B.P. (A-7107). All of these factors suggest that the S8 soil is another example of an Altithermal soil, and that Unit IIc correlates with the Pick City Member of the Oahe Formation (Clayton et al. 1976).

The uppermost unit in the section, Unit III, is a thick bed of fine to medium silt (Figure 15; Table 8). Again, the unit appears eolian in origin on the basis of particle size, although sedimentary structure could not be used as a criteria because pedogenesis has affected the entire deposit. The resultant soil (S9) contains an A-Bw-Bk profile and is classified as a weakly developed Inceptisol (cf. Soil Survey Staff 1990). It is described on the old soil survey maps as being Farland silt loam, high terrace phase (Edwards and Ableter 1944), and has the characteristics of a Typic Haplumbrept (Soil Survey Staff 1990). The Bk horizon in the S9 soil has noticeably fewer carbonates than do similar horizons in the Unit II soils (i.e., class I vs class III). This, together with its stratigraphic position (above an erosional disconformity at the top of the section), suggests that Unit III is temporally and stratigraphically equivalent to the Riverdale Member of the Oahe Formation (Clayton et al. 1976).

The loess deposits, like most of the units in the section, demonstrate a good association between unit boundaries and changes in grain size (Table 8). In general, grain size tends to coarsen downward (Figure 16), although it is important to note that all of the sequences have been affected
by either erosion, pedogenesis, or both (Figure 15). In those units where some parent material remains unaltered by pedogenesis, the coarsest sediment corresponds with the unaltered parent material at the base of the unit (Figure 16). Grain size then decreases upward into truncated Btk (Units Ie and II) or Bk (Units IIA, IIb, IIC) horizons. The upward decrease in grain size in those units with Btk horizons could be the result of pedogenic clay translocation, but the fact that units without Bt horizons exhibit the same general trend suggests that an upward decrease in grain size may be the "normal" pattern of loess deposition in this area. If so, this pattern could be the result of climatic conditions, particularly changes in precipitation. As mentioned, Clayton et al. (1976) and Jorstad et al. (1986) have suggested that episodes of widespread loess deposition in North Dakota correspond with periods of dryer climate. If this is the case, it stands to reason that the same episodes of loess deposition wane as climatic conditions become more mesic. With the increase in rainfall, the loess may become more fine-grained as coarser terrestrial sources of sediment are gradually replaced by finer atmospheric dust deposited onto the surface by rain (cf. Greeley and Iversen 1985). The effects of pedogenesis on grain size were no doubt greatest in those units with Bt horizons, however the overlying (i.e., A and E) horizons associated with these soils were eroded away and grain size data are incomplete. Therefore, pedogenic influences can only be addressed in Unit III, the single loess unit that does not appear truncated (also the most recent unit in the section). The grain-size sequence in Unit III is slightly S-shaped, grading from fine silt at the bottom, to medium and fine silt in the middle, to medium silt at the top (Table 8; Figure 16). The decrease in grain size near the surface does occur in the B horizon and theoretically could reflect some illuvial clay accumulation. Unfortunately, the soil is rather weakly developed and contains no good evidence of a Bt horizon (cf. Soil Survey Staff 1990). As a result, the question of pedogenic influence on grain size remains unresolved.

Locality A is a truly significant section for a number of reasons. First, it yielded a radiocarbon date from Pleistocene-aged Little Missouri River gravels that has important implications concerning the timing and geomorphic history of badlands formation. Second, it contains evidence of at least two periods of clastic deposition in a closed lacustrine or playa-like environment. Similar pluvial episodes have not previously been identified in the badlands region. Third, in addition to five probable Aggie Brown units, it contains as many as three units associated with the Pick City Member. The deposition of each of these units was followed by a period of pedogenesis, a factor which indicates that there may have been multiple episodes of stability during the middle Holocene. Finally, the more recent Riverdale Member is only represented by a single stratigraphic unit; one that
is separated from the underlying Pick City units by a prominent disconformity. Considering the age obtained from Unit IIc, and the age of uppermost Riverdale units elsewhere (i.e., Area O this report; Jorstad et al. 1986), the erosional unconformity could represent a lacuna of 5000 years or more (cf. Krumbein and Sloss 1963; Wheeler 1958).

Localities C, K, and L. While Localities A and O are temporally and paleoenvironmentally significant sections, other loess deposits in the THRO South Unit contain less complete, but nonetheless important, sequences. Among these are Localities C, K, and L, which share similar stratigraphic and pedogenic characteristics.

Locality C is situated at the head of a hardwood draw on the edge of the Little Missouri Escarpment (Figure 4). It contains four late Quaternary stratigraphic units with a combined thickness of 2.3 m (Figure 18). The units overly a weathering profile at the top of Fort Union Group clay. The soil contains a dark (10YR 4/2 dry, 10YR 3/2 wet) mollie epipedon which yielded an age on bulk organics of 10,740 ± 120 yr B.P. (Beta-32478). Like synchronous soils at Localities O and A, the S1 soil at Locality C has a high percentage of organic carbon (i.e., 1.433%--Table 7).

The weathered bedrock is overlain by two thin beds of fine silt (Units 1a and 1b) which appear to be the oldest loess deposits at this locality (Figure 18). Weakly developed Entisols (S2 and S3) have formed on these units, which are pedologically similar to the thin A-C soils present in the bottom portion of Locality O.

Locality C contains one middle Holocene Pick City unit, Unit II, which is a very thick bed of coarse to medium silt. It is capped by a thin ochric epipedon and a weakly developed Bk horizon with stage I carbonate development. The soil (S4) is not characterized by the high levels of carbonate impregnation that are evident in the Pick City soils at Localities A and O. Therefore, it is likely that the S4 period of pedogenesis is post-middle Holocene in age, or synchronous with the lower Thompson Paleosol of the Riverdale Member (i.e., 4000 - 5000 yr B.P.).

The single Riverdale-aged unit at Locality C, Unit III, is a thick bed of medium silt. Pedogenic weathering is evident throughout the entire unit. The resultant soil exhibits an A-Bw-Bk profile (i.e., an Inceptisol), which is a common surface soil in the South Unit uplands. Flaking debris, chipped stone tools, and cores (site 32BI733) are buried within Unit III.

Locality K is a 1.32 m thick deposit of fine to medium silt located along the western edge of a ridge in the Petrified Forest area (Figure 4). It contains four Holocene eolian units (Figure 19). The lowest (Unit I) is a medium bed of fine silt which is capped by a weakly developed Inceptisol
Figure 19. Stratigraphic profile of Locality K.

(S1). The soil is characterized by a dark, thin A horizon and a Bk horizon exhibiting stage I carbonate accumulation. The A horizon yielded a radiocarbon age on bulk organics of 7520 ± 100 yr B.P. (Beta-33044). Unit I is stratigraphically and pedologically similar to the upper Aggie Brown units at Localities A and O (Units 1e and If at Locality A and Unit II and Locality O), all of which appear to have been deposited between ca. 10,000 - 8500 yr B.P. (cf. Clayton et al. 1976). For this reason, the 7520 B.P. radiocarbon age from the S1 soil is somewhat problematic, as it falls outside of the temporal parameters proposed for deposition of the Aggie Brown Member by Clayton et al. (1976) (i.e., 13,000 - 8500 yr B.P.). Given the fact that the Aggie Brown age-estimates have been substantiated by radiocarbon dates recovered from several archaeological excavations (Artz and Ahler 1989; Root and Ahler 1987; Root et al. 1986; Artz 1992), the radiocarbon age from the S1 soil at Locality K appears to be too young. If this is the case, the error may be the result of contamination of the S1 epipedon by younger carbon leached from overlying soil horizons. Indeed,
the S1 soil is overlain by two units (Units IIa and IIb) of medium silt that have been pedogenically altered into Bk horizons with distinctive stage III carbonate accumulations. The A horizons of the soils are absent and the upper boundaries of the Bk horizons are irregular and undulatory. This suggests erosional truncation after pedogenesis (under arid conditions). Consequently, the soils in Units IIa and IIb appear to be middle Holocene (i.e., Pick City) in age. The uppermost unit at Locality K (Unit III) is a thick bed of fine to medium silt (Figure 19). The top portion of the unit contains the modern surface soil, which exhibits an A-Bw profile (i.e., Inceptisol). This unit appears stratigraphically analogous to the Riverdale Member of the Oahe Formation.

Locality L is located approximately 150 m to the southwest of Locality K on the same ridgetop (Figure 4). The two localities are stratigraphically similar, although stage III carbonate accumulation is not apparent in the Pick City sediments at Locality L. Instead, the section has a lower Aggie Brown unit of very fine silt (Unit I) which overlies Fort Union Group shale at a depth of 1.26 - 1.60 m (Figure 18). Again, the unit is capped with a buried Inceptisol (A-Btk profile) which is characteristically high in organic carbon (1.205%--Table 7). Bulk organics from the dark A horizon produced a radiocarbon age of 9200 ± 70 yr B.P. (Beta-33059). This in turn is overlain by a thick Pick City unit of medium silt capped with a truncated Bk horizon (stage I to II carbonates). The late Holocene Riverdale Member is represented by Unit III, which is topped by the modern surface soil (A-Bk profile). This locality is associated with prehistoric artifacts that are eroding out of the Unit III sediments (site 32B1728).

Other Localities. Research at the above mentioned sections made it possible to identify specific Oahe Formation members at additional locations that are not associated with numerical-age data, because the principal members are recognizable on the basis of diagnostic stratigraphic and pedologic criteria. The prominent Aggie Brown units are often at the bottom of depositional sequences overlying Paleocene bedrock or Pleistocene gravels. The more well developed of the Aggie Brown soils are typically the darkest and most organic rich of any of the soils in the Oahe Formation (Clayton et al. 1976:5). The Pick City Member overlies Aggie Brown sediments in the more complete sections. In less complete sections, it is often the lowest member in the sequence. The Pick City sediments are characterized by light coloration, low organic carbon content, and the presence of truncated soils with B horizons. These are often distinctive Bk horizons with stage III carbonate development. Finally, the Riverdale Member is typically represented by one to three units situated at the top of the depositional sequence. Its sediments are darker than the underlying Pick
City and higher in organic matter (Clayton et al. 1976:6-7). Buried Riverdale soils tend to be weakly developed Inceptisols or occasional Mollisols, typically exhibiting A-Bk or A-Bw profiles. The surface soils, on the other hand, are often very youthful Entisols.

With these diagnostic criteria to work with, additional eolian sections comprised of Riverdale and/or Pick City sediments were identified in the South Unit uplands. These exhibit substantial variability in terms of post-depositional preservation and weathering.

Pick City/Riverdale sequences are evident at three undated localities. Locality V is situated on Johnson Plateau where it overlies Pleistocene gravels associated with a cut terrace of the Little Missouri River (Figure 4). The Holocene section consists of a 0.5 m thick deposit of medium to coarse silt separated into two units by a buried soil (Figure 20). The soil is another example of a middle Holocene "Altithermal Soil" with stage III carbonate development. (Figure 20). It is unconformably overlain by a Riverdale-aged silt unit topped with an Inceptisol (Figure 20).

A second Pick City/Riverdale sequence is evident at Locality S, located along the edge of Radio Tower Plateau near the southwestern corner of the study area (Figure 4). The Holocene deposits here consist of 1.75 m of fine sand to medium silt overlying two units of Miocene/Pliocene laminated and cross-bedded sand, and a unit of massive sand whose age and depositional origin are unknown (Figure 20). The Pick City member is represented by a thick bed of light-colored massive silt (i.e., loess), the top of which again contains a distinctive Bk horizon (Figure 20). The section contains two late Holocene/Riverdale loess units (Units IIIa and IIIb). The first is capped with a well-developed Thompson Paleosol, while the second is capped with a very weakly developed Entisol.

A third section with Pick City sediments, Locality Z, is located near the edge of the Little Missouri Escarpment, a short distance north of Locality C (Figure 4). The section contains a thick bed of light yellowish silt (Unit II), capped with a truncated Bk horizon of silty clay (Figure 20). The S1 soil is a composite, or polygenetic, soil (cf. Yallon 1983:248) characterized by stage II carbonate accumulation and strong, medium prismatic structure. Unit II is identified as Pick City on the basis of light coloration of the parent material and the truncated Bk horizon. It is unconformably overlain by a late Holocene Riverdale unit comprised of silt which has been weathered into an A-Bk profile. The Unit III soil is imprinted onto the Unit II soil, indicating that pedogenic alteration extended into the S1 soil during the period when Soil 2 formed. The S2 surface soil is characterized by a mollie epipedon, suggesting that late Holocene landscape stability on the
Figure 20. Stratigraphic profiles of Localities S (left), V (center), and Z (right).
Missouri Plateau portion of the study area may have been more substantial than pedogenic episodes in the badlands during this same period.

Four eolian sections are representative of those localities where only the Riverdale member is extant. The first of these, Locality B, is a rather interesting combination of incipient weathered sandstone and overlying sand dune sediments. It is located on Boicourt Ridge near the northeastern corner of the South Unit (Figure 4). At this locality, the in situ weathering of Paleocene sandstone produced a medium bed of white sand, while more substantial subsequent weathering resulted in the formation of a prominent mollic epipedon at the top of the unit (Figure 21). The soil (possibly an Entic Haplustoll), yielded a radiocarbon age on bulk organics of 1480 ± 80 yr B.P. (Beta-31544). The paleosol is overlain by a unit of fine sand (Unit III) which represents part of a large blowout dune (cf. Ahlbrandt and Fryberger 1982:14) that caps the ridgetop in this area. At present the dune is stabilized by mixed prairie grasses, although in some areas small deflation hollows are present. Unit III is topped with a very weak Entisol, classifiable as a soil only on the basis of its ability to support plant growth, and by the obliteration of primary sedimentary structure by dense root activity.

Additional examples of Riverdale eolian sand deposits are present at Localities R near the Wind Canyon Overlook and Locality Y on Peck Hill (Figure 4). Stratigraphic profiles of these localities are illustrated in Figure 21.

The last Riverdale-aged section to be discussed is Locality U on Johnson Plateau. It mantles the highest (i.e., Pt4) Pleistocene cut terrace surface in the study area (Figure 4). Similar nearby sequences (Localities T, W, X) are evident on the three lower adjacent terrace surfaces (Figure 4). Locality U contains three silt to fine sand units (total thickness 1.8 m). All of the units have been significantly altered by pedogenic weathering, and each contains an A-Bk horizon profile (Figure 22). The entire section appears to be Riverdale in age, as indicated by color, morphology, and soil genesis. The section is of interest because of the degree to which the overlying soils have been welded onto the lower soils (cf. Waters 1992:58). In particular, the A horizons of both buried soils are also part of the Bk horizons of overlying soils (Figure 22). While this welding is fairly common among upland paleosols in the study area, rarely are they polygenetic to the extent evident at Locality U (cf. Bronger and Catt 1989).

Chronocorrelation of Upland Stratigraphic Units

Before leaving the section on eolian and pond stratigraphy, a discussion of unit and member correlation is in order. At the local level, direct correlation between stratigraphic units and paleosols
Figure 21. Stratigraphic profiles of localities R (left), B (upper right), and Y (lower right).
at sections (cf. NACOSN 1983; Boggs 1987:550-551) is hampered by limited visibility and the high degree of stratigraphic variation evident between individual localities. This variation is most apparent in terms of post-depositional sediment preservation and weathering. Nevertheless, radiocarbon ages and visual diagnostic criteria do make possible some basic indirect chronocorrelations among the loess sequences based on gross similarities outlined before. These correlations are illustrated in Figure 23. They represent demonstrated temporal equivalency among upland stratigraphic units, as evidenced by radiocarbon ages, stratigraphic position, and diagnostic lithologic or pedologic characteristics. The previously described criteria for recognition of the Aggie Brown, Pick City, and Riverdale Members are an important element in the correlations because each member is distinctive, and because the ages of each are fairly well established (Artz 1992; Clayton et al. 1976; Reiten 1983; Root and Ahler 1987; Wyckoff 1982). Therefore, while the correlations between individual units or groups of units indicate temporal equivalency, they also infer a certain
Figure 23. Chronocorrelation of stratigraphic units at principal eolian localities, THRO South Unit.
amount of lithologic equivalency because of their association with specific Oahe Formation members, which were originally defined as lithostratigraphic units (Clayton et al. 1976). In spite of the diagnostic Oahe Formation criteria, the correlations would have been more or less arbitrary (cf. Shaw 1982:Figure 2) had it not been for the procurement of radiocarbon ages from a number of key sections. Using radiocarbon dating to establish temporal equivalency, however, requires accurate knowledge of the context of the dated sample and a clear "...delineation of the nature of the event...for which temporal placement is being sought" (Taylor 1987:108). The objective was to establish the temporal parameters of individual episodes of sediment deposition, erosion, and landscape stability in upland settings throughout the South Unit study area. Emphasis therefore, was placed on the "event age" rather than the "sample age" represented by the radiocarbon assays (Dean 1978). Determining the timing of depositional and non-depositional "events" was accomplished by recording the composition and superposition of stratigraphic units, unit boundaries, and buried soils at a number of representative sections: sections associated with strategically located radiocarbon samples (cf. Waters 1992:77). The resultant radiocarbon sample ages therefore represent both periods of sediment deposition and periods of organic accumulation during pedogenesis. In those instances where disconformable contacts are bracketed by radiocarbon ages, they also help establish the temporal parameters of erosional episodes (cf. Waters 1992). Since the geomorphic events associated with the radiocarbon dates are identified in Table 5, the subsequent discussion on the correlation of stratigraphic units will center on the association between stratigraphic units and the three principal members of the Oahe Formation.

In the South Unit uplands, stratigraphic units temporally analogous to the Aggie Brown Member have been identified at Localities A, O, C, L, and K (Figure 23). The number of individual units include six at Locality A (Units Ia-IIf), nine at Locality O (Units Ia-Ib), two at Locality C (Units Ia and Ib), and one each (i.e., Unit I) at Localities L and K (Figure 23). Again, the Aggie Brown units are at or near the bottom of each sequence and associated with one or more Leonard Paleosols. The six radiocarbon ages associated with these units date episodes of sediment deposition (i.e., eolian and lacustrine) and soil formation.

Upland units temporally equivalent to the Pick City Member have been identified at Localities A (Units IIa, IIb, IIc), O (Unit II), C (Unit II), L (Unit II), K (Units IIa and IIb), V (Unit II), S (Unit II), and Z (Unit II)(Figure 23). The two radiocarbon ages directly associated with Pick City sediments date episodes of stability and soil formation. The remaining units have all been identified on the basis of pedologic/lithologic characteristics and stratigraphic position. Again, Pick
City units are characterized by light-colored parent material, low organic matter content, and significant carbonate accumulation.

Sediments temporally equivalent to the Riverdale Member were identified at every upland locality in the South Unit (Figures 4 and 23). While the number of units at any given locality ranges from one to three, they are always at the top of the sequence and are frequently separated from underlying Pick City sediments by relatively dark and well developed buried soils (i.e., the upper and lower Thompson Paleosols). All three radiocarbon ages recovered from the Riverdale sediments date these periods of pedogenesis.

This examination of temporal correlation serves to illustrate the well-known fact that rarely (if ever) does a single stratigraphic section contain a complete sediment record (cf. Boggs 1987:551; Waters 1992:97-98). Consequently, the stratigraphic record of any one depositional environment within any given geochronologic unit is usually a composite of numerous isolated depositional sequences (Gladfelter 1985; Krumbein and Sloss 1963; Shaw 1982; Waters 1992). This point is indeed well illustrated in the South Unit uplands.

At the inter-regional level, Clayton et al. (1976) hypothesized that the eolian lithofacies of the Oahe Formation may partially correlate with the Sanborn Group loess in Nebraska and Kansas, particularly the Peoria and Bignell Formations (Frye and Leonard 1952; Hibbard 1958). More specifically, the Oahe Mallard Island Member may be equivalent to the upper portions of the Peoria Formation, while the Pick City and Riverdale Members may be equivalent to the Bignell Formation (Frye and Leonard 1952, from Clayton et al. 1976:8). Also, the Leonard Paleosol of the Aggie Brown Member may correlate with the Brady soil, which separates the Peoria and Bignell Formations (Frye and Leonard 1952). In Illinois, the Peoria loess is more or less synchronous with the Woodfordian substage, dated at ca. 22,000 -12,500 yr. B.P. (Willman and Frye 1970), while the Bignell loess, which has also been reported from South Dakota (Flint 1955), is apparently Holocene in age (cf. Ruhe 1976, 1983). Although not explicitly identified, the Bignell may also be equivalent to portions of the so-called "lip-loess" which overlies the post Late Yarmouthian Red Dog Loess in the White River Badlands of South Dakota (Harksen 1968:Figure 2). Of particular significance to the present study is Harksen's (1968:11-15) hypothesis that much of the Red Dog Loess sediment originated locally in the White River Badlands.
Summary of Late Quaternary Upland Stratigraphy

Stratigraphic investigations in the South Unit uplands reveal a late Quaternary geologic history dominated by episodes of widespread eolian and localized lacustrine sediment deposition interrupted by periods of erosion and landscape stability.

The most common late Quaternary eolian facies are loess, and to a much lesser extent, dune sands. Significant loess accumulations mantle Fort Union Group bedrock or pre-badland channel lag gravels on virtually every major upland landform in the study area. Sand dune deposits, on the other hand, are extant only on several isolated buttes and ridges. Lacustrine (i.e., shallow pond) facies appear to be much less common, and were identified only at a single study section.

All loess and sand dune deposits mantling the South Unit uplands are assignable to the eolian lithofacies of the late Quaternary Oahe Formation (Clayton et al. 1976; Clayton and Moran 1979). Likewise, the lacustrine sediments, which represent deposition in a closed playa-like environment, are assignable to the "pond" lithofacies of the same formation (Clayton and Moran 1979; Clayton et al. 1980). Of the four originally defined Oahe Formation members, three (the Aggie Brown, Pick City, and Riverdale) are identified in the South Unit uplands. Each of these are identifiable on the basis of lithology, soil characteristics, and stratigraphic position.

The relative "completeness" of the upland study sections varies widely. Some sections contain only one of the Oahe Formation members, some contain two, and others three (Figure 23). In addition, the number and thickness of stratigraphic units associated with the individual members is also quite variable. In general, the most complete sections are located in small upland basins, or bedrock depressions, which have served to trap and preserve sediments.

Although the stratigraphic record is highly fragmentary, temporal and physical correlations enable construction of a generalized stratigraphic section that illustrates the sequence of Late Quaternary sediment deposition in the South Unit uplands. This section (Figure 24) highlights the vertical and temporal arrangement of the three principal members of the Oahe Formation, as well as their stratigraphic and pedologic composition.

Sediments and soils associated with the Aggie Brown Member (Unit 1 in the generalized section) are often situated at the bottom of late Quaternary depositional sequences, overlying Paleocene bedrock or Pleistocene gravels. The latter represent channel lag sediments deposited by the Little Missouri River prior to a series of rapid downcutting episodes that occurred after ca. 24,000 yr B.P. (Figure 24). The downcutting created a number of cut terrace treads that range in height from 60 to 90 m above the elevation of the present Little Missouri channel.
Figure 24. Generalized stratigraphic section illustrating the sequence of late Quaternary sediment deposition in the South Unit uplands.
The number of stratigraphic units identified as Aggie Brown at individual sections range from one to nine. These include both eolian (loess) and lacustrine (pond) facies (Figure 24). The loess units typically consist of massive, very fine silt arranged in thick, medium, and thin beds. In the more complete sections, the thick beds tend to be located at the base of the sequence, the thin beds in the center, and the medium beds at the top. The lacustrine units are apparently limited to closed upland depressions and are characterized by massive, thick mud beds and precipitated carbonates.

The Aggie Brown units are frequently capped with soils, the most well-developed of which are typically the darkest and most organic-rich of any soils in the uplands. In the original description of the Aggie Brown Member, Clayton et al. (1976) recognized only one such soil, named the Leonard Paleosol by Bickley (1972), however research in the South Unit study area indicates that as many as nine soils may be extant in the Aggie Brown sediments. The most distinctive of these are Mollisols, however Inceptisols and Entisols are generally more common.

A total of ten radiocarbon age determinations have been recovered from the Aggie Brown units. These range from 13,070 ± 490 yr B.P. to 7520 ± 100 yr B.P. (Figure 24).

Sediments and soils associated with the Pick City Member (Unit II in the generalized section) overlie Aggie Brown sediments in the more complete sections. In less complete sections, they are often the lowest unit in the sequence. The Pick City is typically the lightest-colored member of the Oahe Formation, a factor which reflects both low organic matter content and pedogenic carbonate accumulation. The number of Pick City units identified at any one locality range from one to three. They are comprised of coarse to very fine silt arranged in medium to very thick beds. All of the units are interpreted as loess deposits on the basis of grain size and massive sedimentary structure.

A distinctive characteristic of the Pick City Member is the presence of paleosols exhibiting stage III calcium carbonate development. This high level of pedogenic carbonate accumulation is not evident in any of the other Oahe Formation members. The soils, which are primarily Inceptisols, consist of truncated Bk horizons with abrupt irregular or undulatory upper boundaries. Humates from two of the Pick City soils yielded radiocarbon ages of 7255 ± 170 yr B.P. and 6580 ± 160 yr B.P. (Figure 24).

Sediments and soils associated with the Riverdale Member (Unit III in the generalized section) are generally located at the top of the depositional sequence (Figure 24). In the more complete sections, they are separated from the underlying Pick City sediments by erosional
unconformities. The Riverdale sediments tend to be darker than the Pick City and higher in organic matter content. At individual localities, the number of stratigraphic units identified as Riverdale range from one to three (Figure 24). These include both loess and sand dune facies. The loess units, which are extant throughout the Riverdale sequence, typically consist of medium to fine silt arranged in medium to thick beds. The sand dune units, on the other hand, are generally found at the top of the Riverdale sequence and typically consist of thin to medium beds of fine sand.

In their original description of the Riverdale Member, Clayton et al. (1976) identified two associated soils, termed the upper and lower Thompson Paleosols (Clayton et al. 1976). In the South Unit study area, soils are present in the upper portion of every Riverdale unit (Figure 23). Therefore, in those sections with three Riverdale units, there are two buried soils and a surface soil, while in those sections with two Riverdale units, there is only one buried soil. In most cases, the buried soils are more well-developed than those at the surface (i.e., buried Inceptisols and Mollisols as compared to surface Entisols and Inceptisols).

Three radiocarbon age determinations are associated with the Riverdale Member units. These ages are 4070 ± 130 yr B.P., 2160 ± 70 yr B.P., and 1480 ± 80 yr B.P. (Figure 24).

The stratigraphic and radiocarbon data indicate that deposition of the Aggie Brown Member began prior to ca. 11,800 yr B.P. and ended prior to ca. 9200 yr B.P. The timing and spatial extent of the earliest episodes of eolian deposition are poorly understood, however the oldest identified loess units are characterized by abrupt and irregular upper unit contacts, indicating that deposition was followed by one or more periods of erosion. During the late Pleistocene, sometime prior to ca. 11,800 yr B.P., ponding and lacustrine deposition occurred in isolated, poorly drained upland basins. At the same time, more well-drained areas underwent landscape stability and pedogenesis. This ponding and synchronous soil formation are indicative of pluvial climatic conditions.

At least one significant episode of loess deposition, possibly resulting from a decrease in effective precipitation, occurred in the uplands during the period from ca. 11,800 to 11,000 yr B.P. This deposition was followed by widespread landscape stability and soil formation, as indicated by the presence of buried mollic epipedons at several localities. These soils are temporally and pedogenically similar to the Leonard Paleosol as described by Clayton et al. (1976), Root et al. (1986), and others.

Loess deposition, interrupted by brief periods of stability, continued throughout the uplands until approximately 10,700 - 10,500 yr B.P., when pluvial conditions returned and lacustrine sediments were again deposited in upland basins. There is some evidence to suggest that the ponds
may have become more shallow by this time as the result of earlier lacustrine and eolian sedimentation. In any event, the last episode of ponding ended prior to ca. 10,400 yr B.P.

The most recent Aggie Brown eolian units were deposited prior to ca. 9200 yr B.P. (Figure 24). These units are again capped with soils, indicating soil formation subsequent to deposition. The pedogenic events were terminated by one or more episodes of upland erosion. This erosion signals the beginning of the generally warm and dry middle Holocene.

The deposition of units associated with the Pick City Member began prior to ca. 7300 yr B.P. and ended sometime before ca. 4100 yr B.P. (Figure 24). These units represent at least two, and possibly three, individual episodes of loess deposition. Each of the depositional events was followed by a period of soil formation which occurred under conditions of reduced precipitation. The associated soils are truncated and separated from overlying stratigraphic units by prominent disconformities. These characteristics indicate that erosional events followed each of the soil forming intervals. At some localities, the amount of time represented by erosional unconformities could exceed 5000 years. It is clear, therefore, that significant upland erosion occurred during the middle Holocene. The agent of this erosion was most likely wind deflation.

Loess accumulation in the uplands resumed sometime prior to ca. 4100 yr B.P., resulting in the deposition of the lowest units associated with the late Holocene Riverdale Member. The deposition was followed by one or more episodes of pedogenesis which resulted in the formation of soils temporally analogous to the lower Thompson Paleosol (Clayton et al. 1976). These soils have dark colored epipedons and relatively high percentages of organic carbon; factors which suggest a return to more mesic climatic conditions.

The lower Thompson episodes of soil formation were apparently widespread throughout the uplands, and were terminated by one or more equally widespread periods of loess deposition prior to ca. 2150 yr B.P. Again, the deposition was followed by at least one episode of upland stability and pedogenesis. The resultant soil appears temporally analogous to the upper Thompson paleosol (Clayton et al. 1976).

Eolian deposition continued throughout the last ca. 2000 years, producing late Riverdale units comprised of loess and dune sands. The sand dune facies represent blowout dunes that mantle small portions of some upland surfaces. The sand originates from weathered exposures of Fort Union Group sandstone that generally outcrop short distances windward of the dunes. Most of the dunes are presently stabilized by mixed prairie grasses, although wind deflation is active in some areas.
While sand dune deposits are present, temporally equivalent loess facies are the most common form of late Holocene surface deposits in the South Unit uplands. These, and the dune sediments, are typically capped with weakly developed surface soils (i.e., Entisols and Inceptisols). The paucity of more well developed soils (i.e., Mollisols) indicates that the South Unit uplands continue to be geomorphologically dynamic landscapes, subject to synchronous, but localized, eolian deposition and erosion.

**Fluvial Depositional Environments**

**Introduction**

Naturally occurring badlands are formed in large part by the erosive action of running water on poorly indurated and sparsely vegetated sediments (Bryan and Yair 1982; Warren 1984). In terms of the scale and timing of geomorphic processes, they are also considered among the world's most dynamic landscapes (cf. Campbell 1989). It therefore stands to reason that fluvial processes in badland environments are equally dynamic. The active and highly variable characteristics of badland fluvial systems have important implications concerning the preservation and distribution of archaeological materials, and hence to our goal of elucidating non-cultural site formation processes. The following discussion on late Quaternary fluvial depositional environments includes a summary of previous geomorphological work in the Little Missouri badlands, a description of key fluvial sections in the THRO South Unit, and a look at the distribution of temporally equivalent floodplain surfaces.

Like the eolian and lacustrine sediments previously described, fluvial deposits in the South Unit are technically assignable to one of the "lithogenetic subdivisions" of the Oahe Formation; in this case the fluvial, or river, lithofacies (Clayton and Moran 1979; Clayton et al. 1980). The Oahe fluvial lithofacies consist of "...as much as 10 meters of channel and overbank sediment", with the overbank deposits comprised of "poorly sorted, obscurely bedded clay and silt with some thin layers of sand, weak paleosols, scattered mammal bones and teeth, terrestrial small snail shells, and fragments of wood" (Clayton et al. 1980:77). The channel sediment is predominately poorly exposed, cross-bedded sand with occasional planar-bedded gravel in alluvial fans at the base of steep scarps (Clayton et al. 1980:77).

The fluvial lithogenetic subdivision of the Oahe Formation is poorly studied and only briefly described. This is the result of a general lack of field research and the fact that most exposures are
limited to the late Holocene Riverdale Member (Clayton et al. 1980). Consequently, as presently
defined, the subdivision has little utilitarian value as a formal lithostratigraphic unit. This is
especially true of the individual members, which are not defined to the point that they can be
unequivocally recognized (NACOSN 1983:854). For this reason, Oahe member designations will
not be applied to the late Quaternary fluvial sediments in the South Unit. The deposits, however,
are subdivided into three major temporal and morphological categories: (1) lag deposits associated
with Pleistocene-age channels of the Little Missouri River (i.e., cut terrace veneers); (2) Holocene-
age meandering stream deposits (i.e., channel bed, bank, and overbank); and (3) Holocene-age gully
deposits (i.e., predominately channel bed). While stratigraphic data are available from each of these
categories, the bulk of available information comes from meandering stream deposits. Unlike
upland eolian sediments, correlation of fluvial deposits will not follow Oahe Formation member
equivalency, but will simply be based on the distribution of synchronous terrace fills. This
distribution is illustrated in Appendix A.

Previous Investigations

The first substantive research into badland fluvial environments was conducted by former
North Dakota State Geologist Wilson Laird in the THRO South Unit (Laird 1950). He identified
four terrace levels associated with the Little Missouri River and made the observation that terraces
in the upper ends of tributary streams are often cut on bedrock and are not associated with
substantive deposits of alluvium (Laird 1950:10). The four Little Missouri terraces were described
as: (1) terrace No. 4, a cut terrace located some 65 m above the present stream level; (2) terrace No.
3, located approximately 12 m above the modern channel, and taking the form of a fill terrace within
the present Little Missouri valley and a cut terrace away from the main stream; (3) terrace No. 2, a
fill terrace located about 4 m above the modern channel; and (4) terrace No. 1, the present stream
level (Laird 1950). Although he did not place these terraces solidly within a temporal framework,
Laird hypothesized that terrace No. 4, evident on Johnson and Big Plateaus, represents the
Pleistocene level of the Little Missouri prior to glacial advances and incipient badland formation
(Laird 1950:8-9).

Laird's terrace definitions were utilized, but somewhat redefined, by two graduate students
from the University of North Dakota who conducted masters thesis research on the geological
history of the Little Missouri River (Schmitz 1955; Petter 1956). Schmitz made the argument that
the eastern-flowing segment of the Little Missouri (i.e., that portion east of "The Bend" in
southwestern McKenzie County) was established during the early Pleistocene by a combination of two episodes of stream capture, and an episode of deep trenching initiated by increased precipitation in advance of the Kansan-Illinoian ice sheet (Schmitz 1955). He also hypothesized a period of lacustrine aggradation initiated by blockage of the Yellowstone, Missouri, and pre-diversion Little Missouri Rivers by the Kansan-Illinoian ice sheet, and a subsequent period ofplanation which created terrace No. 3 (Schmitz 1955). Both Schmitz and Petter attribute the formation of terrace No. 2 to re-entrenchment initiated by isostatic rebound following the final retreat of Wisconsin ice (Schmitz 1955:36; Petter 1956:25). Although substantive chronostratigraphic and numerical age data were not available, Schmitz (1955) identified the No. 4 terrace as Pliocene/Pleistocene in age, the No. 3 terrace as Pleistocene, the No. 2 terrace as "post-Wisconsin", and the No. 1 terrace as modern. It is interesting to note that Petter (1956:13-16), relying on lithological characteristics and heavy mineral analysis, identified the fluvial gravels on Petrified Forest Plateau (Missouri Slope surface) as being temporally equivalent to the Flaxville gravels of Miocene/Pliocene age (Collier and Thom 1918).

More recently, Ben Everitt (1968) investigated the age and history of the modern Little Missouri River floodplain by examining the relationship between floodplain and channel sediments and the growth of cottonwoods. Utilizing age estimates derived from the counting of tree rings he developed a model describing channel migration and sediment transport along a 2.5 km section of the valley in the North Unit of THRO. The resultant data indicated that the modern floodplain sediments were deposited within the last ca. 300 years. He also rejected the earlier No. 1 and No. 2 terrace designations of Laird (1950), Schmitz (1955), and Petter (1956), and argued that both surfaces are components of the modern floodplain. Citing recent flooding of the No. 2 surface (in 1947 and 1950) and lack of evidence for significant downcutting, Everitt further suggested that the ca. 5 m difference in elevation between the No. 1 and No. 2 surfaces was the result of the shifting of floodplain sediment during a significant hydrographic event (Everitt 1968:436-437). He therefore classified the floodplain as split-level, comprised of a "high" and a "low" floodplain (Everitt 1968:436). Everitt also identified alluvial fans at the mouths of tributary streams and along the valley margins; fans constructed by slopewash derived from "the barren slopes of the badlands" (Everitt 1968:435). Although fan surfaces as high as 15 m or more above the low water level of the Little Missouri were observed, Everitt concluded that all of the fans were likely to have been deposited within the last 1000 years (Everitt 1968:435).
In 1966, Thomas Hamilton conducted a geological investigation into late-recent alluvium in the Little Missouri Badlands, concentrating on Jones Creek, an ephemeral stream in the South Unit (Figure 4). Hamilton identified five lithostratigraphic and pedostratigraphic units comprising the lowest valley fill in the region (Hamilton 1967). These units are (from the oldest to the youngest) clayey sand up to 1.5 m thick, a buried soil .3 -.78 m thick, clayey silt 2.5 - 4.7 m thick, a second buried soil .16 -.62 m thick, and sandy silt .31 - 3.1 m thick (Hamilton 1967:152-153). A radiocarbon age on wood of 185 yr B.P. (I-2325) was recovered from the oldest soil, indicating that the overlying units were deposited since A.D. 1765 (Hamilton 1967:153). Hamilton tentatively correlated the late-recent valley fill with the upper part of the Lightning Formation in Wyoming (Leopold and Miller 1954).

Another University of North Dakota graduate student, Jonathan Hagmaier, undertook thesis research in 1967 on the surface configuration of a small drainage basin near Jones Creek (Hagmaier 1967). A study of surface geometry, areal relationships, relief, and surface morphology indicated that the small basin "conformed to prescribed laws of drainage configuration" (Hagmaier 1967:55). Bifurcation and elongation ratios (Horton 1945; Schumm 1956a; Strahler 1964) from the drainage basin were found to be similar to those in other badland areas (Hagmaier 1967).

Most recently, Mark Gonzalez (1987) conducted M.S. thesis investigations into the fluvial geomorphology of Paddock Creek, a perennial meandering stream located less than 5 km south of Jones Creek (Figure 4). Gonzalez identified five fluvial surfaces within the Paddock Creek Valley (four fill terraces and the modern floodplain), situated between 1.5 m and 10 m above the current channel (Gonzalez 1987). The three highest terrace fills were described as having similar thicknesses (3.3 to 6.5 m) and lithology (coarse sand and gravel arranged in large-scale trough cross-beds and muddy vertical accretion sediments arranged in massive planar and ripple cross laminated beds). The lowest terrace and modern floodplain were observed as having similar basal units (i.e., channel lag and point bar) with thin vertical accretion sediments (i.e., less than 1 m). As many as eight buried soils were identified in the three oldest terraces (Qt1, Qt2, Qt3). Five of these contained simple A-C horizon profiles, while three (one in each terrace) were identified as having mollic epipedons (Mollic Ustifluent and Entic or Typic Haplustolls)(Gonzalez 1987:26-31). A total of five radiocarbon ages on bulk organics were recovered from buried soils in three of the terraces (Gonzalez 1987:56). These ranged from 3860 ± 70 yr B.P. (WIS-1907) to 2280 ± 70 yr B.P. (WIS-1910). From these ages, and from tree ring counts on cottonwoods rooted on the modern floodplain, Gonzalez reconstructed the late Holocene fluvial history of Paddock Creek; a reconstruction that
indicated an absence of fluvial sediment older than ca. 5000 yr B.P. in the Paddock Basin (Gonzalez 1987:58). In addition, Gonzalez attributed the dichotomy in sediment age between fluvial deposits in Paddock Creek (i.e., over 3800 years) and nearby Jones Creek (i.e., less than ca. 300 - 400 years) to differences in drainage basin size and sediment storage capacity (Gonzalez 1987:85-87). He also stated that the oldest Paddock Creek terrace (Q1) appeared to grade into Laird's (1950) No. 2 terrace at the Paddock Creek mouth (Gonzalez 1987:74-75). It is significant to note that no mention was given to slopewash deposition, either as interfingering with fluvial sediments in facies relationships, or as alluvial fans at the mouths of ephemeral tributaries.

Pleistocene Channel Lag Deposits

Broad terrace surfaces located high above the modern valley floor are prominent features along much of the Little Missouri River (Figure 25). The terraces have long been recognized in the geological literature as pre-badland, or Late Pliocene to Pleistocene, in age (Laird 1950; Leonard 1916; Petter 1956; Schmitz 1955). As stated previously, Laird (1950) designated one of these surfaces as Little Missouri terrace No. 4, and identified specific remnants of it in the THRO South Unit (i.e., Big Plateau and Johnson Plateau). Although described as an erosional, or cut terrace, Laird observed and photographed sediments associated with the terrace (gravel and silt) on Airport Plateau near Medora (Laird 1950:Figure 4). The deposits are clearly evident on the Geologic Map of North Dakota (Figure 2), where they have been designated as map unit QTu:

Quaternary and Upper Tertiary sediment undivided. Largely river sediment; includes Upper Quaternary terrace, fan, and pediment gravel; gravel composed of rounded pebbles and cobbles of quartzite and porphyry derived from the Black Hills and Rocky Mountains; pebbles and cobbles of locally derived material such as sandstone, silicified wood, and concretions; and Pliocene (?) to Middle Quaternary (?) clay, silt, and sand [Clayton 1980].

The mapped QTu sediments are concentrated along the north-trending portion of the Little Missouri valley, including the abandoned valley northeast of "The Bend", occupied today by Red Wing, Cherry, and Tobacco Garden Creeks (Figure 2). They are virtually absent along the east-trending portion of the modern valley, a factor cited by Leonard (1916:303) as evidence that this arm of the Little Missouri is "postglacial" in age. From "The Bend" east, a significant portion of the river
follows the southern-most margin of Pleistocene glaciation (indicated on Figure 2 by the dashed blue line). This, and the distinct abandoned valley, certainly substantiate the long-held observation that the course of the Little Missouri was diverted to the east by a glacial advance during the Pleistocene (Bluemle 1972, 1977, 1991; Laird 1950; Leonard 1916). The presence of the QTu terrace sediments along the non-diverted and abandoned portions of the valley, but not the diverted portion, indicate either that the QTu sediments pre-date the diversion episode or that they have been largely eroded away in the diverted portion. In any event, the traditional theory holds that it was the diversion which led to the creation of the badlands, by forcing the Little Missouri to flow over a steeper and shorter route, thereby increasing the stream gradient and initiating rapid downcutting (Bluemle 1977:11, 1991:16).

Research into Pleistocene channel lag sediments in the South Unit centered on the examination of six stratigraphic sections. These include previously described Locality A on Big Plateau, and five localities on Johnson Plateau (Figure 4). The latter are associated with at least four different cut terrace surfaces, separated by three clearly visible risers (Figure 26). Equivalent surfaces are present on Big Plateau, but the risers separating them are not as distinct. Thus, based primarily on the Johnson Plateau evidence, a minimum of four Pleistocene Little Missouri cut terraces appear to be extant in the South Unit. These are designated Pt1 (lowest) to Pt4 (highest).
Figure 26. Plan view map of Johnson Plateau showing the location of Pleistocene cut terrace surfaces.
Figure 27. Generalized cross-section showing the vertical relationships between Pleistocene cut terrace surfaces and the modern Little Missouri River floodplain.

Brief descriptions of the stratigraphic sections associated with these terrace surfaces are as follows.

The highest Pleistocene terrace, Pt4, is represented by two stratigraphic sections. These are Locality A on Big Plateau, and Locality U on Johnson Plateau (Figure 4). This surface is situated at an elevation of 2520 ft (768 m) asl, or 90 m above the average low water level of the modern Little Missouri channel, and 55 m below the Missouri Slope surface represented by Petrified Forest Plateau (Figure 27). On Johnson Plateau, it is separated from the Pt3 terrace by a 7.3 m high scarp.

The Pt4 sediments at Locality A unconformably overlie Fort Union Group clays (Figure 15). They have a maximum thickness of .33 m and are comprised of ca. 50% poorly sorted, rounded to subrounded gravels (granule to cobble-sized), and ca. 50% medium sand. The gravels appear to be both local and non-local in origin, although locally-derived clasts (i.e., sandstone, shale, scoria, iron concretions) are the most prevalent. As previously described, the gravels were associated with a radiocarbon dated wood sample (unidentifiable as to species) which yielded a radiocarbon age of 24,230 ± 1510 yr B.P. (A-7114). This important date indicates that the Little Missouri channel occupied the Pt4 surface until at least 24,000 B.P. and has since downcut some 90 vertical meters.
At Locality U on Johnson Plateau, the Pt4 unit has a maximum thickness of 1.5 m and again unconformably overlies Paleocene clay. It consists of poorly sorted sand and gravel, with ca. 80% rounded to subangular granule to cobble clasts and ca. 20% medium to very coarse sand. The Locality U gravels are overlain by three loess units of apparent Riverdale age which are separated from the Pt4 sediments by an erosional disconformity (Figure 22).

The Pt3 terrace sediments were investigated at Locality T, located along the edge of the northern arm of Johnson Plateau (Figure 26). The Pt3 and lower Pt2 surfaces in this area are separated by a 2 - 6 m high scarp. The Pt3 sediments at Locality T (Figure 28) are comprised of a lower medium-thick bed of poorly sorted, medium to coarse sand, and an overlying bed of moderately sorted, subangular to rounded gravels. The gravel clasts are predominately local (i.e., sandstone, scoria, and concretions). The upper gravel bed is overlain by a .7 m thick deposit of Riverdale member loess (Figure 28).

Locality V, also located along the northern arm of Johnson Plateau, is the representative section for the Pt2 sediments. The Pt2 deposits here overly Paleocene sandstone and are comprised of a 1.7 m thick bed of loose, poorly sorted, medium to coarse sand, which coarsens upward to ca. 85% gravel (subangular granules to cobbles) and 15% medium sand. Some of the larger clasts are partially encrusted with carbonates. The unit is overlain by previously described Pick City and Riverdale loess units (Figure 20).

The lowest terrace surface on Johnson Plateau is Pt1, located at the extreme northwestern end of the Plateau at an elevation of 2420 ft (738 m) asl, or 60 m above the current Little Missouri average low water level. The Pt1 and Pt2 surfaces are separated by a distinctive scarp that is 15.7 m in height (Figures 26 and 27).

Sediments associated with the Pt1 terrace were investigated at Locality X (Figure 26). The Pt1 sediments here consist of a 1.4 m thick unit of medium to coarse sand and poorly sorted subangular gravels (Figure 29). Like the lowest unit at Locality V, the sediments exhibit a coarsening upward sequence, with predominately sand at the bottom and gravel at the top (Figure 29). The section is capped with a .5 m thick deposit of Riverdale loess.

The research into Pleistocene cut terraces points to a late Quaternary geologic history that is both more complex and recent than previously recognized. The lithostratigraphic composition and spatial distribution of the terrace sediments leaves little doubt that they are channel lag deposits...
Figure 28. Stratigraphic profile of Locality T.
Figure 29. Stratigraphic profile of Locality X.
associated with a laterally migrating and downcutting stream (cf. Brakenridge 1987; Cant 1982; Walker and Cant 1984). The broad distribution of the terrace surfaces does tend to support the earlier observations (i.e., Laird 1950; Schmitz 1955) that the pre-badland Little Missouri River flowed through a valley that was significantly wider than the modern one. The composition, morphology, and apparent origin of the channel deposits appear generally consistent with the description given to the QTu unit by Clayton (1980). Bluemle (1991:14), however, makes the argument that the Little Missouri River terraces (among others) contain gravels that originated both locally and in the Beartooth Mountains in Montana, rather than the Black Hills and Rocky Mountains. This observation is neither supported nor refuted by the recently acquired data. In any case, surface morphology strongly indicates that the Little Missouri in this area, occupied, migrated away from, and then abandoned by incision, at least four different surfaces within the last ca. 24,000 years. The resultant downcutting covered a distance of at least 90 vertical meters; a process of denudation that certainly contributed a great deal to, if not actually created, the present geomorphological configuration of the badlands.

Holocene Meandering Stream and Gully Deposits

Stratigraphic investigations were conducted along every perennial and major intermittent stream in the South Unit (n = 9), and along scores of minor ephemeral streams and gullies. As a result, five Holocene surfaces were identified in the study area which could be correlated from drainage to drainage. These consist of four terraces, designated (from lowest to highest) T1 through T4, and the modern floodplain (TO). These surfaces correspond in a general sense to those identified by Gonzalez (1987) in Paddock Creek, although there is a much higher degree of variability in age, elevation, distribution, pedogenic, and facies composition than that described by Gonzalez. Within a single drainage like Paddock Creek for instance, this variability is at least partially due to spatial differences in post-depositional erosion (i.e., stream incision, channel migration, high energy flood events), and to the ubiquitous but selective distribution of slopewash sediments. The greatest variation, however, occurs in the degree to which sediment is stored within the different stream valleys; a factor that was clearly recognized by Gonzalez (1987:85-87). These differences (and the resultant number of extant fluvial surfaces) are the result of hydrological factors related to drainage basin size, channel and valley morphology, and discharge (Gonzalez 1987; Horton 1945; Schumm 1977; Strahler 1952). In general, the smaller drainage basins have the lowest capacity for storage, while the larger basins have the highest capacity (Gonzalez 1987:85-87). This
pattern is generally consistent with known geomorphological models (cf. Hadley and Schumm 1961; Schumm 1977). The one exception, however, is the trunk stream itself, the Little Missouri River, whose valley has very minimal sediment storage (Kuehn 1993:321). This could be due to the narrow and deep morphology of the valley which tends to increase flow velocity by restricting the area available to receive discharge.

The research into Holocene stream deposits involved the investigation of 15 stratigraphic sections, each corresponding to one of the four terrace fills (Figure 4). For this reason, the subsequent locality descriptions are organized according to the terrace designations rather than to individual stream systems. Individual stratigraphic unit designations indicate the order of sediment deposition within each section, not correlations between sections. Therefore, Unit 1 at Locality D, for instance, may not necessary be equivalent to Unit 1 at Locality M. The general location of fluvial sediments within the study area and individual terrace correlations are presented in Appendices A and B.

Fluvial systems within the study area exhibit some rather complex morphological characteristics that deserve explanation prior to discussion of the four principal terrace deposits. At the head of larger drainage basins, the oldest terraces often take the form of cut surfaces capped with thin veneers of slopewash (Appendix A, Block K). Associated terrace fills in these areas are largely eroded away by high velocity discharge events. Channels here take the form of steep walled, ephemeral gullies incised into pre-existing slopewash valley fill. These gullies are sometimes inset by younger terrace fills, which in turn are separated from still younger fills by knickpoints, as is the case in the southern branch of Paddock Creek (Appendix A, Block J). Some valleys in the drainage basin are covered with slopewash and are not incised at all (Appendix A, Block K). Further downstream from the drainage basin, in the area roughly equivalent to Schumm's (1977) Zone 2, the slopewash valley fill gives way (via interfinger ing) to bank and overbank meandering stream deposits. Slopewash sediments are still present, but become more confined to the tributary valleys and their mouths (i.e., fans) and to the valley margins. The low-sinuosity gullies, downslope from their intersection point (cf. Patton and Schumm 1975; Schumm and Hadley 1957; Waters 1992:145), give way to meandering stream channels. In the gullied portion of the streams, most sediment is transported during periods of high discharge, in suspension and as bedload within the deep gully channels. Overbank deposition is rare, and when it does occur, the sediments interfinger with slopewash deposits. During periods of low to nonexistent flow, gully bottoms contain very shallow, slightly sinuous channels which gently meander around bedload lag deposits (sand to boulder-sized
gravel). Shallow pools of standing water accumulate in bedrock depressions and at the base of knickpoints. Below the intersection point, sediment is transported and deposited by traditional meandering stream processes (cf. Brakenridge 1987; Walker and Cant 1984). In the meandering portion of the system, older terrace fills (i.e., T3 and T4) are rare and limited primarily to the valley margins, where they almost invariably interfinger with, and/or are capped by, slopewash deposits. The fluvial deposits, whether actively flooded or abandoned, frequently exhibit lateral and/or vertical facies relationships with alluvial fans. Cut-and-fill relationships (cf. Brakenridge 1987:152) are present at the unconformable contacts between different terrace fills, but these contacts are often obscured by erosion, vegetation, or thin veneers of sheetwash.

Smaller drainage basins are often completely unincised or contain only short, relatively simple gully networks (Appendix A, Block G). In these instances, the gully floors are similar to those described for the upper reaches of meandering stream systems (i.e., shallow channels, scattered channel lag, and occasional pools). Gully-mouth fans (Patton and Schumm 1975; Waters 1991, 1992) are present below the intersection points of gullies generally confined to hillslopes. Here the fan deposits interfinger with slopewash and/or fluvial sediments. At the mouths of larger gullies, sediments generally debouch directly into the Little Missouri River channel. In these cases, the unincised slopewash/alluvial fan deposits adjacent to the gully channel interfinger with Little Missouri floodplain sediments. A good example of this type of gully-slopewash/fan system is Beef Corral Wash (Appendix A, Block G).

This, then, is a brief and somewhat idealized summary of meandering stream and gully systems in the South Unit study area. It should be noted that sediment yield to these systems is influenced by a myriad of highly variable factors, including vegetation, bedrock lithology, slope aspect, piping, and mass movement (cf. Campbell 1989). There are, for instance, significant differences in surface runoff and sediment yield between slopes comprised of Bullion Creek and Sentinel Butte Formation materials, because of differential sand and clay composition, which in turn affects infiltration and percolation rates (Clayton and Tinker 1971; Schumm 1956a, 1956b). Likewise, south-facing slopes yield more sediment to the drainage basin (via surface runoff) than north-facing slopes because of greater amounts of effective solar radiation and resultant decreases in vegetative cover (Clayton and Tinker 1971). In addition, subsurface piping allows for the transport of large amounts of sediment during snowmelt or periods of heavy rainfall. Piping is also responsible for the creation of new gully systems as the result of roof collapse (Harvey 1982). Therefore, badland fluvial systems, especially the smaller drainage basins and the upper ends
of larger basins, are constantly undergoing readjustment, often in the form of complex response, as intrinsic thresholds are exceeded by these myriad fluctuations in sediment yield (Schumm 1973, 1977).

Although complex, fluvial systems in the study area are amenable to stratigraphic investigation and subsequent diachronic reconstruction because of the presence of temporally equivalent terraces and floodplains throughout the study area. The presence of these equivalent surfaces in different drainage basins suggests that all of the major streams in the area responded simultaneously (but progressively upstream) to changes in the base level of the Little Missouri River. Likewise, the equivalent surfaces also tentatively indicate that, although intrinsic complex response is a common process in some settings, the driving force behind area-wide episodes of deposition, stability, and erosion in the major streams has been extrinsic in nature, probably fluctuating climatic conditions (cf. Brakenridge 1987; Knox 1984a; Schumm and Brakenridge 1987). Since hydrologic factors related to basin size and morphology have apparently been responsible for the differential preservation of sediment, the next step in the discussion of fluvial depositional environments is to describe those equivalent terrace fills and/or surfaces that have survived denudation. For quick reference, the following Holocene terrace and floodplain configurations are identified in each of the principal South Unit streams:

Little Missouri River—modern floodplain; T1 terrace; T2 terrace (very isolated remnants).

Knutson Creek—modern floodplain; T1 through T4 terraces.

Paddock Creek—modern floodplain; T1 through T4 terraces. At head of drainage basin, terrace fills are inset into gullies and T3 and T4 occur as cut terrace surfaces.

Sheep Creek—modern floodplain; T1 and T2 terraces; T3 and T4 cut terrace surfaces at head of drainage basin.

Jules Creek—modern floodplain; T1 and T2 terraces.

Jones Creek—modern floodplain; T1 terrace.

Beef Corral Wash—modern gully in slopewash valley bottom.

Boicourt Wash—modern floodplain; T1 through T3 terraces; T4 cut terrace surfaces.

Petrified Forest Wash—modern floodplain; T1 through T3 terraces; T4 cut terrace surfaces.
**T4 Terrace.** The distribution of T4 terrace fill in the South Unit is limited to 10 small remnants, concentrated primarily along the margins of larger stream valleys and in the upper reaches of large drainage basins (Appendix A). In the latter, however, equivalent erosional terraces, with little or no associated fluvial sediments, are usually more prevalent. The fill terrace surface generally lies from 10 to 12 m above the modern channel floors, although actual elevations were found to be highly variable. Along the valley margins, the terrace tread is generally higher because of slopewash mantles on the tread. Away from the valley margins, the tread is sometimes lower because of erosion and a lack of slopewash. Although not readily visible during the present investigations, a cut terrace surface approximately 12 m above the modern channel has been reported in the Little Missouri River valley by earlier investigators (i.e., the No. 3 terrace of Laird 1950; Schmitz 1955; Petter 1956). The elevation of this reported surface is analogous to that of the T4 terrace.

The terrace fill was intensively investigated at three stratigraphic sections in Knutson Creek (Localities D, E, and M), however all visible exposures in the study area were carefully examined (Figure 4). The three key sections vary significantly in tread height, facies relationships, and number of buried soils; factors directly related to the intensity of, and proximity to, slopewash deposition.

Locality D, being situated in the center of the Knutson Creek valley away from active slopewash sedimentation, is the only T4 terrace section whose fill appears completely fluvial in origin. The terrace, however, is truncated, as evident by a rounded, sloping tread surface, and by an apparent disconformity near the top of the section. As a result, the highest portion of the fill surface is only 7.1 m above the current channel. Unconformable cut-and-fill contacts with the younger T2 terrace fill are evident along the east and west edges of the T4 remnant. The basal portion of the section consists of approximately 2.2 m of gravel and coarse sand which were not exposed, but were identified in soil probes. These sediments are interpreted as being channel lag and lower point bar deposits. The lowest exposed stratigraphic unit (Unit 1, thickness 1.6 m), is comprised of small scale trough cross-bedded fine sand, massive fine sand, and interbedded silt, fine to coarse sand, and granules (Figure 30). This unit is interpreted as representing the upper portion of a point bar (cf. Allen 1970; Reineck and Singh 1980; Walker and Cant 1984). Unit 2 consists of 2.2 m of ripple cross-laminated silty clay topped with a medium thick bed of massive silt and a weakly developed soil, S1 (Entisol). A charcoal sample from near the top of the unit (near the bottom of the massive silt) yielded a radiocarbon age of 6115 + 145/-140 yr B.P. (A-7113)(Figure 30). Unit 2, interpreted
Figure 30. Stratigraphic profile of Locality D.
as floodplain vertical accretion deposits (Reineck and Singh 1980), is overlain by four predominately massive silt units, each capped with a soil (Figure 30). The S2 and S3 soils are Entisols with very thin A horizons, while the S4 soil is more strongly developed, having a dark (10YR 3/2 moist) mollic epipedon, the lower part of which has noticeable calcium carbonate accumulation (stage 1). These characteristics suggest that the S4 soil developed during a substantially longer period of formation than either S2 or S3. The S4 soil is tentatively classified as a Typic Haplustoll (Soil Survey Staff 1990). The upper contact of Unit 5 is irregular, indicative of erosional truncation. The A horizon of the S4 soil, however, does not appear eroded; a factor that suggests the S4 period of stability followed, rather than preceded, the erosional event that truncated Unit 5. The section is capped with Unit 6, a medium-thick bed of loose silt containing a very weakly developed Entisol. This unit appears to be a thin loess mantle that post-dates the deposition of the T4 fill.

Locality E, located near the mouth of Knutson Creek, is a good example of Terrace 4 in a valley-margin position; one in which fluvial sediments both interfinger with, and are topped by, slopewash sediments (Figure 31). This section is located on the concave side of a meander loop at the distal end of a bajada-like series of slopewash fan deposits, which overlie the T4 terrace fill sediments. The terrace exhibits cut-and-fill contacts with T2 terrace fills along its east and west margins. Unit 1, at the base of the section, consists of 1.6 m of imbricated gravels, large scale trough cross-beds of coarse to medium sand, interbedded thin beds of sand/granules and mud, and massive to ripple laminated mud (Figure 31). This unit is interpreted as lateral accretion sediment. Unit 2 consists of 3.3 m of ripple cross-laminated sand and occasional parallel laminated mud layers, interpreted as a natural levee deposit (cf. Reineck and Singh 1980). A charcoal sample collected from near the bottom of Unit 2 (Figure 31) produced a radiocarbon age of 6660 ± 45 yr B.P. (A-6481). Unit 3 consists of 3.4 m of interfingerling thin to medium beds of silt (vertical accretion) and sand and gravel (slopewash). The slopewash beds are comprised of coarse sand and angular, poorly sorted, granule to cobble-sized clasts of locally derived gravels. The locality is capped with 1.8 m of slopewash (Unit 4), which is comprised primarily of massive sand and thin beds of sand and gravel. The only soil evident at Locality E is the weakly developed surface soil (an Entisol). The paucity of buried soils is common within the T4 sequence when it is situated along the valley-margin. This is attributed to a lack of stability resulting from frequent slopewash deposition.
Figure 31. Stratigraphic profile of Locality E.
While Locality D is an example of a mid-valley T4 terrace with no appreciable slopewash, and Locality E an example of the more common valley-margin terraces with abundant slopewash, the third T4 section to be described is more or less intermediate. At Locality M, some slopewash is present, but not in sufficient quantity to have prevented multiple pedogenic episodes. The locality is located near the northern edge of the valley but not flush against the valley margin (Figure 4). It is separated from active slopes by a large exposure of sandstone bedrock; a factor that may have slowed the rate of slopewash deposition in this portion of the valley. The T4 fill at Locality M consists of 2.5 m of large scale trough cross-bedded sand and interbedded fine sand and mud (Unit 1), overlain by 5 m of ripple cross-laminated fine sand and silt with two thin plane beds of slopewash (sand and gravel). Unit 2 in turn is overlain by six thin to medium silt beds (total thickness 1.5 m), each topped by soil horizon development (Figure 32). The most well developed soil is the lowest (S1), characterized by a 10 cm thick ochric epipedon and a Bk horizon with Stage 1 carbonate accumulation (Entic Haplumbrept?). The remaining soils are simple Entisols (thin A horizons) with the exception of S6 (Figure 32). It is characterized by a A-Bt profile, with a noticeable clay increase in the B horizon. It also has strong medium prismatic structure. The locality has a thin (7 cm) veneer of loose silty fine sand that has barely weathered into a soil (Entisol). It is interpreted as a mantle of loess or slopewash from the adjacent exposed sandstone. These recently deposited veneers are a common feature of terrace treads throughout the study area, and are responsible for the youthful surface soils that characterize most terrace surfaces. This pattern is contrary to the data provided by Gonzalez (1987:35-38) who reports a predominance of Mollisols at the surface of the oldest terraces in Paddock Creek (i.e., Typic Argiustolls and Typic Haplustolls).

The charcoal dates recovered from near the bottom (6660 ± 45 yr B.P.) and top (6115 ± 145/ -140 yr B.P.) of T4 in Knutson Creek indicate that the terrace fill was deposited between ca. 6000 - 7000 years ago. This places the period of aggradation within the relatively warm and dry middle Holocene. It also corresponds to central portion of the "Wolf Creek Unstable Episode", identified by Bluemle and Clayton (1982:13-14) as a period when xeric "semi-desert?" conditions prevailed in southwestern North Dakota. The well developed soils at the top of Localities D and M most likely formed during a relatively lengthy period of stability that post-dates the middle Holocene (i.e., synchronous with the lower Thompson Paleosol of the Riverdale Member). Based on the stratigraphic evidence from Locality D, this period of stability appears to have followed a significant erosional episode.
Figure 32. Stratigraphic profile of Locality M.
T3 Terrace. Although more prevalent than the older T4 terrace fill, T3 sediments are by no means common anywhere in the study area. Like T4, the fill is primarily extant along valley margins, where it interfingers with, or is capped by, slopewash. Terrace 3 also occasionally occurs as eroded remnants: mid-valley "islands" (Appendix A, Block L). In the upper reaches of the drainage basins there are a fair number of corresponding cut terrace surfaces with thin mantles of slopewash. In general terms, these often follow T4 surfaces in a progression downslope from the upper ends of the basins (i.e., T4 cut surfaces dominate the upper reaches while T3 surfaces dominate the lower reaches). There are exceptions, however, and some upper basins are completely dominated by T3 erosional surfaces (Appendix A, Block K). The elevation of the T3 tread generally ranges from 7.5 to 8.5 m above current channel bottoms, but again variation is common. Three stratigraphic sections are offered as representative examples of the T3 fill sequence. 

Locality N is situated near the southern edge of the Paddock Creek valley at a point where the two major branches of Paddock Creek converge (Figure 4). The T3 and lower T2 terrace fills in this area exhibit clear unconformable cut-and-fill contacts. The terrace fill at Locality N consists of a thick bed of poorly sorted coarse sand and angular gravel overlain by a unit of ripple cross-laminated fine sand and silt, two units of mud, and a thick mantle of loose silt (Figure 33). The ripple laminated unit is capped with a fairly well developed soil with a A-Btk profile. The 4 cm thick A horizon yielded a bulk organic radiocarbon age of 5000 +130/-125 yr B.P. (A-7137). The Btk horizon contains a noticeable clay increase and stage 1 calcium carbonate accumulation. The overlying mud units have also been weathered into soils. In Unit 3 the soil exhibits an A-Bw profile, while the Unit 4 soil is characterized by a thin A horizon. The silt unit at the top of the section has a simple A-Bw soil profile. As mentioned earlier, this unit was associated with two Late Prehistoric fire hearths located at a depth of 13 - 20 cm below the surface. The charcoal from these hearth features yielded radiocarbon ages of 1390 ± 100 yr B.P. and 1180 ± 90 yr B.P.. The disparity between these ages and the 5000 +130/-125 B.P. age obtained from the soil developed on Unit 2, indicates that the deposition of Unit 5 significantly post-dates the deposition of the T3 terrace fill. Given this disparity, and the lithology of the Unit 5 sediments (i.e., loose silt), Unit 5 is interpreted as a post-fluvial mantle of wind-blown loess. The former terrace surface appears to be represented by the soil developed on Unit 4. Although previously suspected on the basis of noticeable lithological differences and a general lack of terrace soil chronosequences, the recovery of radiocarbon ages from the two hearth features is the first hard evidence of post-fluvial terrace veneers in this region (cf. Artz 1992). This is in keeping with Hassan's (1985a:59) observation that
Figure 33. Stratigraphic profile of Locality N.
so-called "desert loess" is a common facies in arid and semi-arid fluvial environments. In other areas, veneers also occur in the form of younger alluvium, which can hamper terrace correlations based solely on surface soil profiles (Brakenridge 1987:152).

Locality BB is situated along Boicourt Wash in the northeastern corner of the South Unit (Figure 4). While there are occasional T4 and T3 cut terrace surfaces in the vicinity, the locality is a good example of the T3 terrace fill. It consists of a basal unit of planar bedded coarse sand and gravel (grauels to pebbles), a unit of small scale trough cross-bedded sand, a unit of ripple cross-laminated fine sand and silt, and an upper unit of silty fine sand (Figure 34). An A horizon soil is present at the top of Unit 3, while a more substantial A-Bw soil profile has formed onto the Unit 4 alluvium (Figure 34).

Locality AA is located near the eastern end of the northern branch of Paddock Creek, or in the upper portion of the drainage basin (Figure 4). The section consists of a basal unit of channel lag overlain by a ripple laminated fine sandy silt unit, and an upper unit of mud (Figure 35). The ripple laminated unit is topped with a weakly developed soil (Entisol) which is associated with a prehistoric fire hearth excavated into the former surface (Figure 35). Like many T3 terraces in the region, the surface soil, situated 8.2 m above the current Paddock Creek channel, is a simple Entisol.

The basal lag deposits at Localities N, AA, and BB are all situated approximately 5 m above the floor of the adjacent modern channels, a distance that represents the amount of incision that has occurred in the study area since the T3 period of aggradation. This aggradation is estimated to have occurred between ca. 5500 and 4500 yr B.P., based on the 5000 +130/-125 yr B.P. age from the S1 soil at Locality N and on the basis of two additional ages which appear to bracket the aggradational episode. A dark buried A horizon 1 m below the surface of an alluvial fan in Knutson Creek (Fan 93B) yielded a bulk organic age of 4300 ± 125 yr B.P. (A-7111). This age suggests that the lowlands may have experienced landscape stability by ca. 4300 years ago. Likewise, a buried soil on a second alluvial fan (Fan 93 F) produced an age of 5900 ± 90 yr B.P. (Beta-31872). This indicates that another stable episode may have followed the earlier T4 period of aggradation. If this is the case, the T3 episode of fluvial deposition must have occurred after ca. 5900 B.P. but before ca. 4300 years B.P. Although this temporal framework is admittedly sketchy, the estimated time of the T3 aggradation does fall within the middle Holocene "Wolf Creek Unstable Episode" (Bluemle and Clayton 1982:13), a period characterized by reduced effective precipitation, increased sediment yield, and resultant valley aggradation.
Figure 34. Stratigraphic profile of Locality BB.
Figure 35. Stratigraphic profile of Locality AA.
**T2 Terrace.** The tread of this terrace ranges in height from 7 to 5.5 m above the floor of modern channels. It is widespread in Knutson and Paddock Creeks, where it is just slightly less prevalent that the younger T1 fill (Appendix A, Blocks H and L). It is also common along Boicourt Wash, the lower reaches of Jules Creek, and along portions of Petrified Forest and Sheep Creeks (Appendix A). It is conspicuously absent from all remaining low-order tributaries and from the Little Missouri River valley, with the exception of two small remnants. In Paddock Creek, which is the only perennial stream that is entirely located within the South Unit study area (i.e., from drainage basin to mouth), T2 fill tends to increase in frequency downstream (Appendix A, Blocks E, G, H, I, J). Near the head of the drainage basin, it gives way to T3 and T4 (both in the form of isolated fill remnants and as cut terrace surfaces). Downstream from a prominent knickpoint in the major southern branch of the creek, the T3 fill becomes entrenched. The T2 fill is inset against T3 along this segment. At a second knickpoint further downstream, the T2 fill becomes entrenched. The T1 fill is inset against T2 along this segment. Finally, at the lowest knickpoint, the T1 fill becomes entrenched and the former gully gives way to the meandering Paddock Creek channel (Appendix A, Block J).

Like the other terraces, T2 also exhibits a high degree of variability because of varying contributions of slopewash. As a result, three rather disparate, but representative, examples of the T2 fill are described.

Locality P is one of only two T2 remnants to be identified within the Little Missouri River valley (Figure 4). It consists of interbedded medium to fine sand topped with a thin bed of mud (Unit 1), interpreted as the upper portion of a point bar. A prehistoric fire hearth near the top of Unit 1 produced a charcoal radiocarbon age of 2190 ± 130 yr B.P. (Beta-27718)(Kuehn 1989:20-23). Unit 2, vertical accretion sediment, is comprised of ripple cross-laminated fine sand topped with an Entisol (Figure 36). The vertical accretion deposits are overlain by a unit of interfingering slopewash and overbank sediment (interbedded thin beds and massive thick beds of medium sand, fine sand, and mud) (Figure 36). The two upper units (Units 4 and 5) consist of slopewash in the form of fine to medium sand with very thin beds of coarse sand and granules. The upper portion of Unit 4 has been altered into an A horizon soil. A bulk soil sample from the A horizon yielded an age of 1210 ± 60 yr B.P. (Beta-32316).

While at Locality P most of the stream-deposited sediments are at the bottom of the section, a reversed situation is evident at Locality CC in Knutson Creek (Figure 37). Here stream sediments
Figure 36. Stratigraphic profile of Locality P.
Figure 37. Stratigraphic profile of Locality CC.
overlie an alluvial fan facies (Figure 4). The fan sediments consist of 3.5 m of interbedded massive sand (coarse to fine) and gravel (angular granule to cobble) (Figure 37). The coarsest sediment is at the bottom of the section, but arranged in horizontally bounded beds with no visible internal sedimentary structure. The number of gravel beds decreases near the top of the unit, while the thickness of the sand beds increase (as does sand grain size). The alluvial fan sediment was truncated by a stream channel, the bottom of which is visible as an erosional disconformity (Figure 37). The channel is filled with ripple cross-laminated mud (Unit 2) overlain by a series of thin, wavy laminated clay beds (Figure 37). The top of the Unit 2 channel fill has been altered into an A horizon soil. Charcoal from the A horizon yielded a radiocarbon age of 2475 ± 60 yr B.P. (A-7110). Unit 2 is overlain by a thick bed of clay, Unit 3, most of which has been weathered into the modern surface soil (A-Bw profile). Locality CC therefore reflects initial alluvial fan deposition that was subsequently interrupted by channel erosion apparently resulting from lateral migration of the Knutson Creek channel.

Channel migration during the T2 period of aggradation is also reflected in the stratigraphic composition of Locality DD, located near the mouth of Knutson Creek (Figure 4). The bottom of the section consists of 1.3 m of interbedded medium sand and clay grading upward to massive fine sand (Unit 1). The unit was truncated by a channel cut, visible as a prominent concave disconformity (Figure 38). This is overlain by Unit 2, 0.6 m of thin planar bedded medium sand, the top of which is truncated by a second channel cut (Figure 38). Unit 2 is overlain by 1.8 m of massive medium sand which fines upward to fine sand. The Unit is topped with a medium bed of laminated clay that is partially weathered into a buried soil (simple Entisol). The uppermost unit (Unit 4) consists of 2.4 m of interbedded thin beds of fine sand, granules, and clay drapes, in turn overlain by 1.0 m of massive silt (Figure 38). Charcoal recovered from near the bottom of the silt bed (at 90 cm below the surface) produced a radiocarbon age of 880 ± 25 yr B.P. (A-6482). The silt is capped with a 60 cm thick surface soil, exhibiting an A-Bw profile. The stratigraphy at Locality DD is interpreted as upper point bar sediments which were truncated by a migrating channel, partially filled with relatively coarse bedload material, and then truncated a second time by rapid migration of the channel back into the same portion of the valley. The second channel was then filled with fining-upward sand and topped with a bed of clay. After a brief period of stability, deposition resumed in a natural levee environment.
Figure 38. Stratigraphic profile of Locality DD.
A second example of T2 fill along the Little Missouri River was investigated at the location of archaeological site 32SL208 in the southern portion of the badlands (Figure 1). A buried epipedon at 52 cm below the tread surface (at the top of unit 3, wavy laminated sandy mud) yielded a bulk organic age of 2400 ± 60 yr B.P. (Beta-38736)(Figure 39). Charred corn from a prehistoric fire hearth in an overlying loess veneer was dated at 275 ± 50 yr B.P. (AA-1318).

![Diagram of stratigraphic profile](image)

**Figure 39.** Stratigraphic profile of vertical accretion sediments and loess veneer at site 32SL208, North Dakota State Lands.

The four apparently reliable radiocarbon ages reported from fluvial sediments at Localities P, CC, DD, and 32SL208 cumulatively indicate that the T2 period of aggradation occurred between ca. 2700 and 800 yr B.P. This relatively long time interval includes one, if not several, periods of landscape stability and soil formation (as indicated by the presence of buried soils at every locality). The estimated period of T2 aggradation corresponds to the xeric late Holocene "unstable episode" as described by Bluemle and Clayton (1982:13).
**T1 Terrace.** This is the youngest and most widespread terrace fill in the study area. In Paddock and Knutson Creeks it is only slightly more common than the somewhat older T2 fill, while in the low order ephemeral tributaries it is the only terrace fill present (Appendix A). It is equivalent to Everitt's (1968) "high flood plain" of the Little Missouri River, although that surface is rarely flooded (last recorded floodings in 1947 and 1950). Following the definition provided by Wolman and Leopold (1957:105), a floodplain becomes a terrace when the surface is no longer inundated by the annual flood event (which is less than once every two years). That being the case, Everitt's "high flood plain", whose tread stands 5 m above the current low water level, technically classifies as a terrace (T1). In addition, Everitt's "low flood plain", which is comprised of active and inactive (ridge and swale) point bar surfaces and meander scrolls, is regularly flooded, and therefore is the actual TO surface, or modern floodplain. Everitt (1968:42) also specifically uses the term floodplain to denote those portions of the valley floor that are topographically molded by overbank deposition.

In addition to being the youngest and most widespread, the T1 terrace is also the most extensively studied and well-dated terrace in the region. As previously summarized, is was the object of at least three extensive geomorphological investigations (Everitt 1968; Hamilton 1967; and Gonzalez 1987). All of these produced similar results as far as the age of the fill is concerned. Because of the significant amount of information already available, the examination of the lithology and age of the terrace will center on the description of one key stratigraphic section, and on a brief discussion of the timing and sequence of events which led to its abandonment. Radiocarbon and limited lithostratigraphic data from four additional localities are also provided.

Locality G is an exposure of the T1 terrace along the Little Missouri River near the Mike Auney Bottoms (Figure 4). As illustrated in Figure 40, the locality contains classic floodplain stratigraphy in the form of fining upward channel bed (i.e., bottom stratum) deposits, overlain by overbank (i.e., top stratum) sediments. The channel bed deposits (Units 1 and 2) consist of coarse-grained sand and gravel with interbedded clay arranged in large scale troughs (channel lag) and trough cross-bedded sand fining upward to thin horizontal beds of fine sand, silt, and clay (point bar). The overbank deposits consist of a thick unit (Unit 3) of predominately massive fine sand with occasional clay drapes (Figure 41). Two charcoal samples collected from one of the clay drapes at a depth of 2.2 m produced radiocarbon age determinations of 380 ± 120 yr B.P. (Beta-32261) and 150 ± 60 yr B.P. (Beta-31786). The only soil evident at Locality G is the modern surface soil, a youthful Entisol. Of additional interest is the fact that the basal units at the locality were found to unconformably overlie a lignite coal unit of the Fort Union Group. The same unit was observed at
Figure 40. Stratigraphic profile of Locality G.
the bottom of the current river channel immediately south of the profile during a period of very low discharge. This indicates that the river in this area flows over bedrock, thereby precluding the presence of deeply buried alluvium (Kuehn 1993:322). Finally, the presence of a thick overbank unit at the top of the section is a common feature of rivers that experience rapid channel migration (Reineck and Singh 1980:251-252). This is in agreement with Everitt's (1968:434-435) research that documented rapid channel migration through measurements of average rates of bank erosion (1.8 million cubic feet per year per mile of valley). Rapid channel migration could certainly be a dominant factor in the river's poor capacity for sediment storage (Gladfelter 1985:44-48), and hence the lack of older fluvial sediment.

Additional stratigraphic and chronologic information on the T1 terrace was collected from Locality F on the Little Missouri, Locality H on Jules Creek, Locality Q on Sheep Creek, and Locality EE on Knutson Creek (Figure 4). The stratigraphic profile associated with Locality EE is illustrated in Figure 42. Together the sections yielded three additional radiocarbon ages (Table 5):
Locality F -- 290 ± 80 yr B.P. (Beta-31937); Locality H -- 100 ± 0.8% modern (Beta-32164); and Locality Q -- 300 ± 65 yr B.P. (Beta-31477).

The five radiocarbon dates recovered during the course of the current study, coupled with the radiocarbon and tree-ring data reported by Hamilton (1967), Everitt (1968), and Gonzalez (1987), clearly indicate that the T1 period of aggradation occurred between ca. 400 and 150 yr B.P. This corresponds with the "Mandan Stable Episode" of Bluemle and Clayton (1982:13); a time characterized by essentially modern climatic conditions, but with a short-lived period of decreased precipitation and increased slopewash erosion from approximately 1000 to 500 years ago. In terms of broader regional patterns, the T1 period of aggradation correlates temporally with the deposition of the Lightning Formation in Wyoming (ca. 750 - 110 yr B.P.) according to Leopold and Miller (1954) and Albanese and Wilson (1974).

As to the abandonment of the T1 surface from regular inundation, evidence from Locality G suggests that the Little Missouri has not been associated with significant bedrock incision since sometime prior to the onset of the T1 episode of deposition, which occurred about 400 yr B.P. This
observation is based on the fact that the basal units of the T1 terrace at Locality G overlie the same Paleocene coal unit that currently forms the bottom of the modern channel. On the other hand, Gonzalez (1987:70), counting tree rings from cottonwoods rooted on the modern floodplain, argues that the incision responsible for the abandonment of T1 and formation of the modern floodplain in Paddock Creek, must have occurred sometime between ca. 200 and 115 yrs B.P. This temporal estimate is supported by Everitt's dendrochronological data, and by the radiocarbon data gathered during the current study. So, if the evidence suggests that the Little Missouri has not downcut significantly into bedrock since the deposition of T1, then what was responsible for the abandonment of T1 and the creation of the modern floodplain? A possibility is that the channel incised to its current level prior to the aggradation of T1, but then filled to a level ca. 1.5 to 3 m higher than at present (due to a significant decrease in discharge). Sometime between ca. 200 - 115 years ago, one or more significant flood events flushed the channel fill from the valley, causing the Little Missouri base level to lower to its former cut surface (i.e., 1.5 to 3 m). This lowering, which occurred without significant bedrock incision, for all practical purposes left the T1 surface abandoned, first in the trunk stream, and then in the tributaries (as their base levels adjusted). While only a working hypothesis, this scenario is plausible considering the lack of evidence for channel bedrock incision, and in light of arguments by Everitt (1968:436-438) that the Little Missouri witnessed the shifting, or significant rearrangement, of floodplain sediment during an "extraordinary hydrographic event" prior to 1880.

Summary of Late Quaternary Fluvial Stratigraphy

Late Quaternary fluvial sediments in the THRO South Unit are subdivided into three broad temporal and morphological categories: (1) Pleistocene-age meandering stream sediments deposited by the Little Missouri River (i.e., channel lag); (2) Holocene-age meandering stream deposits (i.e., channel lag, point bar, and overbank); and (3) Holocene-age gully deposits (i.e., predominately channel lag). These sediments are arranged in a number of cut terrace veneers, depositional terrace fills, floodplains, and active channels. The surfaces of these landforms have a broad vertical distribution, as illustrated in Figure 43, a generalized cross section of late Quaternary fluvial sediments in the South Unit.
Figure 43. Generalized cross-section illustrating Pleistocene and Holocene-age terrace and floodplain sediments in the THRO South Unit.
The Pleistocene channel lag deposits are extant on four individual cut terrace treads that range in elevation from 60 to 90 m above the present channel of the Little Missouri River (Figure 43). In the South Unit, these surfaces, designated Pt1 through Pt4, are most evident on Johnson Plateau and Big Plateau. The oldest and highest terrace tread, Pt4, has a mean elevation of 768 m asl while the lowest terrace tread, Pt1, has a mean elevation of 738 m asl.

The sediments associated with each of the terraces have similar lithologic characteristics, although local variation in grain size distribution is evident. The deposits generally consist of 50% to 85% poorly sorted, rounded to subangular gravels (granule to cobble), and 15% to 50% medium to very coarse sand. These are arranged in medium to thick horizontally bounded beds (average thickness ca. 1.3 m). The gravel clasts appear to be both local and non-local in origin, although locally-derived clasts are the most prevalent (i.e., sandstone, shale, scoria, iron concretions). The sediments overlie Paleocene bedrock of the Fort Union Group and are typically mantled by eolian (and occasionally lacustrine) sediments of the late Quaternary Oahe Formation. At every locality, the Oahe Formation sediments are separated from the lag gravels by erosional unconformities. Wood fragments from the gravel deposits at Locality A on Big Plateau yielded a radiocarbon age of 24,230 ± 1510 yr B.P. (Figure 43).

Holocene-age meandering stream and gully deposits are widespread throughout the South Unit lowlands and are characterized by a high degree of temporal, spatial, lithologic, and pedologic variability. This variability is primarily the result of differences in sediment yield and sediment storage both within and between drainage basins. Sediment yield is influenced by a myriad of factors, including vegetation, bedrock lithology, slope aspect, piping, and mass movement, while sediment storage is influenced primarily by hydrological factors related to drainage basin size, and channel and valley morphology.

In spite of variations in sediment yield and storage, fluvial sediments in the THRO South Unit are arranged in a number of temporally equivalent terraces and floodplains. The presence of these surfaces in different drainage basins suggests that all of the major streams in the area responded homogeneously to changes in the base level of the Little Missouri River. The temporally equivalent surfaces consist of four terraces, designated (from lowest to highest) T1 through T4, and the modern floodplain (T0).

The stratigraphic composition of the terrace and floodplain sediments is dependent upon valley location. In valley-margin settings, the fluvial deposits frequently interfinger with, or are capped by, slopewash sediments. These typically consist of massive, poorly sorted medium to
coarse sand arranged in thin to medium plane-beds. At the mouths of tributary streams, the fluvial deposits often exhibit lateral and vertical facies relationships with alluvial fans. Fan sediments, also deposited by slopewash processes, range from coarse, poorly sorted sand and gravel arranged in thin to medium beds to fine-grained massive and laminated mud arranged in medium to thick beds. In mid-valley locations, slopewash deposits are generally absent and the terrace and floodplain fills are comprised almost exclusively of fluvial sediments. In all three settings, terraces are occasionally mantled by thin loess veneers.

Slopewash (and to a lesser extent, loess) deposition has served to limit pedogenesis, especially in valley-margin settings. In areas with the most active slopewash deposition, buried soils are virtually absent and surface soils tend to be poorly developed Entisols. Where slopewash deposition is less pronounced, some buried soils are present (Entisols and occasional Inceptisols). The most well-developed soils are associated with T3 and T4 terrace fills located in mid-valley settings. Here, individual sections may contain from one to seven buried soils. These consist of Entisols, Inceptisols, and more rarely, Mollisols. Although slopewash sediments are absent in the mid-valley locations, surface soils often exhibit simple A - C horizons due to the presence of apparently recent loess veneers.

An idealized mid-valley fluvial sequence consists of coarse-grained channel lag deposits (i.e., imbricated gravels, coarse to medium sand arranged in large-scale trough cross-beds), overlain by upward-finishing point bar sediments (medium to fine sand arranged in small scale trough cross-beds, ripple cross-laminated and parallel-laminated fine sand and mud), and overbank deposits, either natural levee (ripple cross-laminated and parallel-laminated fine sand and mud), or flood-basin (massive fine sand and mud).

In the South Unit study area, the T4 terrace fill is limited to small, isolated remnants concentrated primarily along the margins of larger stream valleys and in the upper reaches of large drainage basins. The terrace surface generally lies from 10 to 12 m above the modern channel floors, although localized elevations vary because of post-depositional erosion and the presence or absence of slopewash sediments. Charcoal recovered from fluvial sediments near the top and bottom of the T4 terrace fill yielded radiocarbon ages of 6115 ± 145/ - 140 yr B.P. and 6660 ± 45 yr B.P. (Figure 43).

The T3 terrace fill is located primarily along valley margins and in occasional mid-valley settings within the larger perennial and intermittent tributaries of the Little Missouri River. Although more prevalent than the T4 fill, the spatial distribution of T3 within the study area is still
quite limited. The T3 tread generally lies from 7.5 to 8.5 m above currently active channel bottoms (Figure 43). Bulk soil organics from a buried soil at one of the T3 sections yielded a radiocarbon age of 5000 ±130/125 yr B.P. (Figure 43).

The T2 terrace fill is extant in a variety of valley locations throughout the South Unit, with the exception of the Little Missouri River and very low-order tributaries. The T2 surface lies from 7 to 5.5 m above the floors of currently active channels and the fill is associated with radiocarbon ages of 2475 ± 60 yr B.P., 2400 ± 60 yr B.P., 2190 ± 130 yr B.P., and 880 ± 25 yr B.P. (Figure 43).

The T1 terrace sediments are well preserved in every meandering stream valley in the study area. Along the smallest ephemeral tributaries, they are the only fluvial deposits present. The T1 surface lies approximately 5 m above active channel bottoms and 2 to 3.5 m above active floodplain surfaces (TO). The fill is associated with five recently acquired radiocarbon ages ranging from 380 ± 120 yr B.P. to 100 ± 0.8% modern (Figure 43).

The chronostratigraphic data from the South Unit indicate that prior to 24,000 years ago the Little Missouri River occupied a surface approximately 90 m higher than at present. Sometime thereafter, apparently in response to base level lowering associated with a glacial advance of Late Wisconsin age, the Little Missouri experienced rapid downcutting on a scale sufficient to initiate badland formation. This downcutting was not a single event, but rather a minimum of four individual episodes of incision and lateral migration between ca. 24,000 and 7000 yr B.P. The first three episodes of downcutting covered a vertical distance of approximately 30 m, while the fourth episode covered a vertical distance of approximately 50 m (Figure 43).

The earliest identifiable episode of Holocene fluvial aggradation occurred between ca. 7000 and 6000 yr B.P. with deposition of the T4 terrace fill. The timing of this aggradation corresponds to the central portion of the "Wolf Creek Unstable Episode" identified by Bluemle and Clayton (1982) as a period of xeric climatic conditions during the middle Holocene. A second period of area-wide fluvial aggradation during the latter portion of the "Wolf Creek Unstable Episode" is evidenced by the T3 terrace fill, which was deposited between ca. 5500 and 4500 yr B.P. The T3 episode of aggradation was apparently followed by a long interval of downcutting and fluvial degradation which ended by ca. 2700 yr B.P. with deposition of the T2 terrace fill. This aggradational episode continued until approximately 800 yr B.P., although there is evidence to suggest that the deposition was interrupted by one or more periods of stability and soil formation. The T2 episode of aggradation corresponds to the xeric late Holocene "unstable episode" identified for the North Dakota region by Bluemle and Clayton (1982). The T2 aggradation was again followed by
downcutting and degradation until deposition of the T1 terrace fill began approximately 400 years ago. The T1 episode of aggradation continued until ca. 150 yr B.P., when the base level of the Little Missouri River dropped some 1.5 to 3.0 m. This lowering for all practical purposes left the T1 surface abandoned. Within the last ca. 150 years, fluvial deposition has been largely confined to the modern TO floodplain and active meandering stream and gully channels.

Slopeswash and Alluvial Fan Depositional Environments

Introduction

The final category of late Quaternary depositional environments involves sediment deposition via surface runoff and slopeswash processes. These processes have been the subject of at least one previous study in the THRO South Unit. In a comprehensive analysis of hillslope lowering using rod and washer techniques, Clayton and Tinker (1971:4) argue that "the dominant hillslope process in the Badlands ... is slopewash." On slopes comprised of Tongue River (i.e., Bullion Creek) sediments, slopewash predominately takes the form of unconcentrated flow, while on slopes comprised of Sentinel Butte Formation sediment, slopewash is concentrated in medium-sized rills (Clayton and Tinker 1971:28). In other words, the term "slopewash", as applied by Clayton and Tinker, includes sediment transported by both overland flow and rillflow. Overland flow is also known as "rainwash" (Schumm 1956b), "unconcentrated wash" (Fenneman 1908), "sheetflow" (Bull 1977), and "sheetwash" (Bryan and Yair 1982). Whatever the name, it refers to sediment-laden water that flows unconcentrated over the ground surface in broad sheet-like fashion. Rillflow refers to the concentration of runoff into networks of small, shallow channels which often coalesce into larger channels or gullies (Horton 1945; Campbell 1989; Savat and De Ploey 1982). I will follow Clayton and Tinker's (1971) lead in combining both overland flow and rillflow under the term "slopewash". In their study, conducted in a portion of the Jones Creek drainage basin, Clayton and Tinker concluded that slopewash accounted for 99.9% of the sediment yield off of hillslopes. This is in spite of the fact that "almost every conceivable type of mass movement occurs in Buffalo Creek basin", including the colluvial processes of dry sliding, soil creep, seepage step retreat (i.e., slumping), and "various types of flows and slides" (Clayton and Tinker 1971:26). These other processes, however, are responsible for only minor amounts of sediment yield (Clayton and Tinker 1971).
Slopowash is ubiquitous throughout the South Unit, and is virtually ongoing as a depositional process. Clayton and Tinker (1971), however, have demonstrated that slopowash deposition is somewhat seasonal, following summer thunderstorms and spring/winter periods of rapid snowmelt. As previously mentioned, slopowash sediments can take the form of sheet-like deposits on slopes, valley-margin deposits, valley-floor deposits, and valley-mouth deposits. Valley-margin deposits invariably interfinger with fluvial sediments, and have already been described. Sheet-like slope deposits are generally thin, poorly sorted, and coarse-grained; similar to sheetflood veneers on arid zone alluvial fans (Harvey 1989) There is a great deal of variability in grain-size, however, dependent upon the lithology of the slope parent material. The dynamic and ongoing nature of their deposition virtually guarantees that slope sheet deposits are going to be quite recent in age (older deposits having long since been transported to valley bottoms). For these reasons, the following discussion will concentrate on the distribution, morphology, and stratigraphy of valley-floor, and valley-mouth (i.e. alluvial fan) deposits; both of which have similar lithologies.

Alluvial Fan and Valley-Floor Deposits

Valley-floor sediments, those that cover the bottom of pre-existing valleys (Figure 44), are found in almost every basin that does not contain a meandering stream (the only exception being youthful V-shaped gullies incised into bedrock slopes). Because of their ubiquitous nature, they were not individually mapped, however the location of some of the more prominent fill deposits immediately adjacent to major streams are illustrated in Appendix A. Many of the valley floor sediments have been incised by gullies, particularly those on south-facing slopes. This aspect-related factor could be the result of increased slopowash in the upper portions of valleys (resulting from increased solar radiation and decreased vegetative cover), which in turn increases the gradient of the valley floor. When the steepening of the gradient exceeds a geomorphic threshold, gullying is initiated (Schumm and Hadley 1957). A good example of incised valley-floor sediments can be found along Beef Corral Wash, an ephemeral gully in the north-central portion of the South Unit (Figure 4). The slopowash deposits are approximately 9 m thick and interfinger with Little Missouri T1 terrace sediments (Figure 45). They are comprised of generally thick plane beds of massive fine to medium sand, occasional medium plane-bedded mud, and thin to medium plane-bedded coarse sand and granules. There are no visible coarse gravel basal units, and sedimentary structure is almost exclusively massive, with fine laminations noted in the mud beds. A bison mandible was
Figure 44. Photograph of typical unincised valley-floor deposits, THRO South Unit.

Figure 45. Photograph of gullied valley-floor deposits at Beef Corral Wash.
recovered from a massive sand unit at a depth of 5.6 m at Locality I (Figure 4). It yielded a bone collagen radiocarbon age of 930 ± 90 yr B.P. (Beta-32163).

Since a fair amount of time is likely to have passed between the time of the animal's death, its decomposition, and transport of the bone clast to the valley bottom, the radiocarbon age suggests that the 5.6 m of overlying slopewash is substantially younger than ca. 900 years.

Alluvial fans are present at the mouth of virtually every valley in the study area. Again, only those that intersected major stream basins were mapped and studied in terms of their stratigraphic relationships with fill terraces. These fans are illustrated in Appendix A. The most comprehensive data on fan thickness, lithology, entrenchment, and facies associations were collected from seven fans in the Knutson Creek valley, two in the Little Missouri River valley, and two in the valley of Paddock Creek (Appendix A, Blocks L and H).

In Knutson Creek, four of the seven fans have been incised by gullies (Fans 93A, B, C, F). Of these, two (Fans 93A and F) have T1, T2, and T3 terrace fills inset against the gully sidewalls, indicating that the fans were deposited prior to the T3 period of aggradation (i.e., prior to ca. 5500 B.P.). At the other two (Fans 93B and C), the T1 and T2 terrace fills are inset against the gully sidewalls, indicated a fan age greater than ca. 2700 years. Of the three ungullied fans, two (Fans 93D and G), overlie T2 terrace fills (indicating that they are younger than T2), and one (Fan 93E) underlies and interfingers with T2 (indicating synchronous and/or slightly earlier deposition). Fan morphology ranges from classic fan shapes (Fans 93E and 93G) to more linear forms (Fan 93C). An interesting feature of three of the gullied fans is that the fan surfaces on either side of the channel are at different elevations (Fans 93A, C, F). This indicates either separate episodes of deposition, non-synchronous deposition because of slope morphology, or differential post-depositional erosion.

Two of the Knutson Creek fans are associated with radiocarbon age determinations on buried soil organics. Fan 93B produced a radiocarbon age of 4300 ± 125 yr B.P. (A-7111) from a truncated Bt horizon buried to a depth of .9 to 1.2 m (Table 5). The age indicates that the lower portion of Fan 93B may be synchronous with the T3 period of aggradation. Fan 93F contained a buried A horizon (at .7 to .8 m below the surface) which yielded a radiocarbon age of 5900 ± 90 yr B.P. (Beta-31872). This age indicates that the deposition of Fan 93F (at least the lower portion) may be roughly contemporaneous with the deposition of the T4 terrace fill (Table 5).

In Paddock Creek, Fan 93A is gullied, and the T1, T2, and T3 terrace fills are inset against the gully sidewalls. The presence of a knickpoint in the gully indicates that there were actually two episodes of entrenchment, one prior to T3 aggradation, and one after the T1 aggradation (i.e., TO
fill is inset against the sidewalls of the lower channel. Fan 93D has a classic fan shape and is not gullied. A clearly visible unconformable contact along the cutbank side of a meander loop indicates that the fan is older than the T3 terrace fill (i.e., the T3 fill is inset against the fan sediments). In a large majority of gullied fans along Paddock Creek, the T1 terrace fill is inset against the gully sidewalls, while a fair number appear more recent; either overlying (or interfingering with), the T1 terrace, or having TO modern floodplain sediments inset against the fan sediments.

Fans along ephemeral streams are both gullied and ungullied. Again, the majority of incised fans are situated at the mouth of south-trending valleys. The gullied fans have modern floodplain sediment inset against the channel side walls, while the ungullied fans tend to overlie the T1 terraces. Both factors indicate a predominance of fan formation subsequent to the deposition of T1. There are also a number of ephemeral valley fans that interfinger with the T1 sediments.

Fans extending into the valley of the Little Missouri River include Fan 93A, which actually consists of two separate slopewash valley-floor deposits that coalesce into a single gullied fan (Appendix A, Block F). Although the fanhead scarp is largely obscured by vegetation, it appears that the T1 terrace fill is inset against the fan sediments (i.e., the fan is older than ca. 400 years). Fan 93B is situated across the river from Cottonwood campground at the location of Locality F, an exposure of the Little Missouri T1 terrace (Figure 4; Appendix A, Block L). The fan is actually a bajada, or series of coalesced valley-wall fans (Figure 46). It has a distinctive fanhead scarp which is 5.5 m above the T1 tread (or 10.5 m above the modern low-water level). The fan is gullied in several places, with the T1 terrace fill inset against the sidewalls of one of the gullies. This indicates that Fan 93B, like 93A, is older than 400 yr B.P.

To reiterate, the lithology of fan and valley-floor deposits is highly variable, ranging from very coarse and poorly sorted gravel and sand, arranged in alternating thin to medium beds (i.e., Knutson Fan 93A) to fine-grained massive and laminated mud, arranged in medium to thick beds (i.e., Knutson Fan 93B). In spite of this variability, the majority of fan and valley-floor deposits do exhibit generally similar lithostratigraphic composition. A typical example is Little Missouri Fan 93B. Here the basal unit consists of a series of medium horizontally bounded beds of silt, medium sand, and granules, overlain by 4.3 m of interbedded silt (in medium to thick plane beds) and massive gravels (predominately granules in thin to medium plane beds). There are also two medium beds of coarser gravels (predominately small cobbles). The top of Unit 1 contains a very weak A-C
soil. Unit 2 consists of alternating thin plane beds of silt and gravels (mostly granules), and is again capped with a weak Entisol. The surface unit consists of .43 m of massive fine sand and the modern surface soil, which is also an Entisol (Figure 47).

In summary, alluvial fans and valley-floor deposits are second only to sheetwash deposits on slopes, in terms of their frequency of occurrence. In general, the age and distribution of fan and valley-floor sediment appears to mirror the pattern observed in the age and distribution of fluvial deposits; that is, the greatest degree of preservation occurs in the larger valleys with perennial-flowing streams, while the least preservation occurs in the smaller valleys with ephemeral discharge, and in the valley of the Little Missouri River. There is little doubt therefore, that the preservation of fan and valley-floor sediments is governed by the same hydrological factors that govern the preservation of fluvial sediments (i.e., drainage basin size, morphology, discharge, and flow velocity). The radiocarbon data from three of the fans (Knutson 93B, 93C, and 93E), coupled with stratigraphic position and observed interfingering, indicate that significant fan deposition occurs at roughly the same time as meandering stream aggradation. The climatic implications of this apparent synchronicity will be discussed in the subsequent chapter.
Figure 47. Stratigraphic profile of Little Missouri Fan 93B.
Summary of Late Quaternary Stratigraphy

Late Quaternary sediments in the THRO South Unit are associated with eolian, fluvial, slopewash, and to a lesser extent, lacustrine and colluvial, depositional environments. These environments are the product of various geomorphic processes that operate within more or less consistent topographic parameters. As a result, eolian processes are the dominant agent of sediment deposition in upland settings while fluvial and slopewash processes are the dominant agents of deposition in the lowlands. This basic pattern is not without exception, but it is consistent enough to enable construction of a composite stratigraphic profile showing the vertical and temporal distribution of principal late Quaternary sediments in the South Unit (Figure 48).

The top of the vertical sequence is comprised of eolian and lacustrine sediments of late Pleistocene and Holocene age. These mantle Fort Union Group bedrock or Pleistocene alluvium at elevations above ca. 738 m asl (Figure 48). The upland sediments are assignable to the Aggie Brown, Pick City, and Riverdale Members of the late Quaternary Oahe Formation (Clayton et al. 1976). Each of these members contain a variable number of stratigraphic units definable on the basis of bounding discontinuities. The Aggie Brown Member lies at the bottom of the sequence and is predominately loess (i.e., very fine silt), with occasional shallow pond sediments (i.e., muds). The member contains multiple soils, including organic-rich Mollisols. Radiocarbon data suggest that the Aggie Brown Member dates from ca. 11,800 to 8500 yr B.P., although the age of the oldest units have yet to be firmly established.

The Pick City Member, comprised entirely of loess (coarse to very fine silt) overlies the Aggie Brown and is distinctive because of the presence of truncated Bk horizons with significant accumulations of calcium carbonate. Radiocarbon data suggest that the Pick City Member dates from ca. 8500 yr B.P. to 4500 yr B.P.

The Riverdale Member overlies the Pick City and is comprised of loess (medium to fine silt) and dune sands (fine sand). The member contains a number of buried soils that are generally more-well developed than those at the surface, but less developed than the distinctive Aggie Brown Mollisols. The Riverdale Member was deposited within the last ca. 4500 years.

The lower portion of the upland sequence consists of Pleistocene-age channel lag sediments which unconformably overly Fort Union Group bedrock (Figure 48). The deposits are associated with four cut terrace surfaces which were created by lateral migration and downcutting of the Little Missouri River. The terrace treads lie at elevations between ca. 738 and 768 m asl. Each of the
Figure 48. Composite stratigraphic profile of principal late Quaternary sediments in the THRO South Unit.
terraces, designated (from lowest to highest) Pt1 through Pt4, contain a relatively thin veneer of coarse-grained alluvium (Figure 48). These deposits have similar lithologies, consisting of 50% to 85% poorly sorted, rounded to subangular gravels (granule to cobble), and 15% to 50% medium to very coarse sand arranged in medium to thick horizontally bounded beds. Radiocarbon data indicate that the downcutting episodes responsible for creation of the terrace surfaces occurred after ca. 24,000 yr B.P.

The lowland sequence is comprised of fluvial and slopewash sediments of Holocene age (Figure 48). The fluvial sediments, which are the product of meandering stream and gully processes, are concentrated in valley-floor and valley-margin settings. The slopewash sediments, which are the product of overland flow and rillflow, are concentrated on slopes, and in valley-floor, valley-margin, and valley-mouth settings. In valleys containing meandering streams, the coexistence of both fluvial and slopewash depositional environments results in complex vertical and lateral facies relationships. In valley-margin settings, fluvial deposits frequently interfinger with, or are overlain by, slopewash sediments. At the mouths of tributary streams, the fluvial deposits often exhibit lateral and vertical relationships with alluvial fans (which also form as a result of slopewash processes). In mid-valley locations, slopewash sediments are generally absent, and fluvial sediments predominate.

The fluvial deposits are characterized by a high degree of temporal, spatial, lithologic, and pedologic variability. This variability is the result of differences in sediment yield, particularly slopewash deposits, and sediment storage. Nevertheless, fluvial sediments throughout the study area are arranged in a number of temporally equivalent terrace and floodplain fills. These consist of TO, the modern floodplain, and four depositional terraces, designated from oldest to youngest, T1 through T4 (Figure 48).

The lithological composition of the terrace and floodplain sediments is dependent upon the previously described relationships between slopewash and fluvial facies. In the absence of slopewash, an idealized fluvial sequence consists of coarse-grained channel lag deposits (i.e., gravels and coarse to medium sand), overlain by upward-finining point bar sediments (medium to fine sand and mud), and overbank deposits (fine sand and mud).

The degree to which pedogenesis has altered the terrace and floodplain sediments varies according to the age of the fills and the amount of active slopewash deposition at any given locality. Soils are generally absent in the modern floodplain and T1 terrace fills, with the exception of poorly developed surface soils (i.e., Entisols). Soil development increases progressively in the T2, T3 and
T4 terraces fills, however this development is governed primarily by the presence or absence of slopewash. Terraces located in areas of active slopewash deposition tend to have a lack of soils except for those at the surface, which are generally poorly developed Entisols. Where slopewash deposition is less pronounced, some buried soils are present. These tend to be Entisols and Inceptisols. The most well-developed soils are associated with the older T3 and T4 terrace fills located in areas without active slopewash deposition (i.e., mid-valley settings). Here, individual sections may contain from one to seven buried soils (primarily Entisols, Inceptisols, and rarely, Mollisols).

The distribution of the T4 terrace fill is limited to small, isolated remnants extant along the margins of larger stream valleys and in the upper reaches of large drainage basins. The terrace tread generally lies from 10 to 12 m above the floors of active channels. Radiocarbon data indicate that aggradation of the T4 terrace fill occurred between ca. 7000 and 6000 yr B.P.

Although more prevalent than the T4 fill, the spatial distribution of the T3 fill is limited to valley margins and in occasional mid-valley settings within the larger perennial and intermittent tributaries of the Little Missouri River. The surface of the T3 terrace ranges in height from 7.5 to 8.5 m above active channel bottoms. The aggradation of the T3 fill occurred between ca. 5500 and 4500 yr B.P.

The T2 terrace is present in a variety of valley locations, with the exception of the Little Missouri River, and very low-order tributaries, where it is rare to non-existent. The T2 surface lies from 7 to 5.5 m above active channel bottoms. The aggradation of the T2 fill occurred between ca. 2700 and 800 yr B.P.

Sediments associated with the T1 terrace are well preserved throughout the lowlands. In valleys occupied by the lowest order tributaries they are the only fluvial deposits present. The T1 surface lies approximately 5 m above the active channel bottoms and 2 to 3.5 m above active floodplain surfaces. The T1 fill was deposited between ca. 400 and 150 yr B.P.

The TO, or modern floodplain, was established approximately 150 years ago, when the base level of the Little Missouri River dropped some 1.5 to 3.0 m. Within the last ca. 150 years, fluvial deposition has been largely confined to the TO floodplain and to the channels of meandering streams and gullies.

The deposition of slopewash sediments is more or less ongoing, although periods of significant slopewash deposition appear to correspond with the major episodes of fluvial aggradation. As previously mentioned, slopewash sediments can occur in the form of sheet-like
deposits on slopes, and as valley-margin, valley-floor, and valley-mouth deposits. The lithology of the slopewash sediments varies according to the lithology of the slope parent material. Ideally, sheet-like slope deposits are generally thin, poorly sorted, and coarse-grained; similar to sheetflood veneers on arid zone alluvial fans. Valley-margin sediments frequently interfinger with, or overlie, fluvial deposits, and most commonly consist of massive, poorly sorted medium to coarse sand arranged in thin to medium horizontally-bounded beds. Valley-floor sediments cover the bottoms of valleys that do not currently contain meandering streams. These and valley-mouth sediments, or alluvial fans, have similar lithologies, ranging from coarse, poorly sorted sand and gravel arranged in thin to thick beds, to massive and laminated muds arranged in medium to thick beds. Valley-floor and valley-mouth sediments are often incised by gullies, many of which exhibit cut and fill relationships with the terrace and floodplain fills.
CHAPTER VI

LATE QUATERNARY PALEOENVIRONMENT

Introduction

The purpose of this chapter is to complete the elucidation of the natural environmental context of the South Unit archaeological record. A significant portion of this context was identified in the preceding chapters by outlining and describing major late Quaternary sedimentary depositional environments (Chapter V), and principal landform characteristics (Chapter II). Both of these are important constituents of "environment" in general, defined by Hassan (1985b:87) as "a subdivision of the earth's surface distinguished and delimited on the basis of physical, chemical, and/or biological criteria." Hassan recognized three principal components of the natural environment as viewed from a geologic and geomorphological perspective. These components are: (1) climatic-morphogenetic systems; (2) geomorphic units; and (3) depositional systems (Hassan 1985b:Table 4.2). In that geomorphic units and depositional systems were described in preceding sections, this chapter provides data relevant to late Quaternary climatic-morphogenetic systems (Hassan 1985b; Peltier 1950). Climatic-morphogenetic systems are broad zones characterized by specific rainfall and temperature regimes and associated with particular sets of geomorphic units (Butzer 1976b; Hassan 1985b; Peltier 1950).

The importance of climate as an agent in the formation and genesis of geomorphic units and depositional environments cannot be overstated. The term climate refers to the more salient physical properties of the atmosphere, particularly temperature, humidity, precipitation, evaporation, wind, and barometric pressure. With its direct influence on vegetation, climate is the driving force behind the weathering, erosion, transportation, and deposition of sediment (Hassan 1985b:Figure 4.1). As noted by Barnosky (1989) and Barnosky et al. (1987), there is a real dearth of significant paleoclimatic and paleoecologic data from large portions of the Northern Great Plains. This chapter will attempt to help alleviate that deficiency by presenting new information on climatic conditions in the Little Missouri Badland region during the last ca. 24,000 years (especially the last 12,000 years), and by relating these climatic conditions to changes in the "geomorphic landscape"; defined by Hassan (1985b:88) as the association of individual geomorphological units. The data are derived primarily from stable carbon isotope analysis, stratigraphy, and paleopedology. These are the
success stories. Less successful was the attempt to procure climatic data from fossil pollen; a problem of poor preservation that is not uncommon in arid and semiarid regions (cf. Barry 1983). The following chapter includes a brief introduction to some of the basic principles and models associated with the study of late Quaternary climates, a summation of the recently acquired isotopic data, and a reconstruction of the late Quaternary climatic and geomorphic history of the study area.

Basic Principles and Models

The dominant paradigm in contemporary studies of climate change is the astronomical model of latitudinal and temporal (both long-term and seasonal) variation in solar insolation; variations driven by changes in the precession, tilt, and eccentricity of the earth’s orientation and orbit around the sun (COHMAP 1988; Davis 1984). These models have shed considerable light on climatic conditions during the late Quaternary, but have also highlighted the pitfalls associated with attempting to make generalizations between regions, physiographic sections, and even individual ecosystems (Davis 1984; Davis and Sellers 1987; Hollliday 1989b). Because of variability and periodicity in the precession of the equinox, insolation maxima occur repeatedly over relatively short time intervals; each associated with a specific season. For instance, the most recent maxima for early summer occurred ca. 13,000 yr B.P., for midsummer ca. 8000 yr B.P., and for late summer ca. 5000 yr B.P. (Davis 1984). These maxima, however, affected different ecosystems at different times and rates, depending upon the elevation, composition, and physical requirements of the ecosystem (Davis 1984). Likewise, the influence of climate upon any given locality is affected by other highly variable, but inter-related, environmental characteristics such as aspect, thermal conductivity of soils and vegetative cover, reflectivity, and topography (Lamb 1977). Of importance to the present study, with its emphasis on changing landscapes, are models of biogeomorphic response to climate change. As illustrated in Knox (1972), and Selby (1985), abrupt climate change, depending upon the magnitude and frequency of the event, often leads to significant vegetative and landform change resulting from the disruption of long-term equilibrium. The resultant change, however, be it to vegetative communities or landforms, frequently lags behind the more abrupt change in climate, sometimes on the order of hundreds of years (cf. Bryson et al. 1970; Bryson and Wendland 1967; Knox 1972). In a geomorphic landscape as dynamic as the Little Missouri Badlands, both sudden climate change, and subsequent lagged adjustment, can have important implications concerning the
preservation or visibility of archaeological materials. Some of these implications are addressed during the course of the subsequent discussion.

It is safe to say that the climate in southwestern North Dakota during the last 24,000 years was dominated by the adjustment from glacial to postglacial conditions (Porter 1983). In the contiguous United States, glacial conditions persisted until ca. 14,000 years ago, and continued to be the major influence on climate until at least 12,000 years ago, although the influence was time transgressive from west to northeast (Barry 1983; COHMAP 1988; Wright 1992). The maximum extent of continental glaciation occurred about 18,000 years ago, although solar insolation in general reached a minimum approximately 5000 to 7000 years earlier (Barry 1983:395). During maximum glaciation, annual temperatures in the United States averaged 3° to 8°C lower than at present, but extremes approached 14° to 17°C lower in the Rocky Mountains (Barry 1983). The temperature of the sea-surface was approximately 2.3°C lower than at present, with maximum cooling of up to 15°C occurring in the northern latitudes. This cooling was partially the result of strong anticyclonic circulation around the Laurentide ice sheet. There is evidence that the time interval from 22,000 to 19,500 yr B.P. was characterized by interstadial conditions; although Harmon et al. (1979) cite evidence for cooling on the order of 0.6°C - 1.0°C per millennia subsequent to a relatively warm and moist period at approximately 23,000 yr B.P. in the Great Plains (Barry 1983:395,399). The ca. 18,000 B.P. glacial maximum was also responsible for a split in the jet stream across North America during the winter months (COHMAP 1988:1048); the southward displacement of which contributed to high pluvial lake levels in the American Southwest between ca. 18,000 and 14,000 yr B.P. The period from 12,000 to 6000 B.P. was a time of increased seasonality because of greater thermal contrast between the oceans and land masses (COHMAP 1988). Haynes (1991) cites evidence for a cold-dry drought in western North America between ca. 11,300 -10,900 yr B.P.; a climatic episode that appears to correspond with the Younger Dryas event in northern Europe (Dansgaard et al. 1989).

Postglacial insolation maxima (with temperatures 2° - 4° C higher than at present) occurred before 9000 yr B.P. in western North America, however in eastern North America, the summer thermal maxima was delayed until approximately 6000 B.P. due to the slowly retreating ice sheet (COHMAP 1988). In general, summer temperatures at 6000 B.P. were 2° to 4° C higher than present throughout much of the North American interior (COHMAP 1988:1048).

Traditional models (Antevs 1948, 1955) characterized the climate of western North America during the Holocene as tripartate, with the early Holocene being wet and cool, the middle Holocene warm and dry, and the late Holocene wet and cool (but essentially modern). Each change was
interpreted as a gradual transition from one regime to the other. As more extensive paleoclimatic data became available, an increasingly complicated and diverse Holocene paleoclimatic history emerged. Among the first in the United States to recognize this diversity were Bryson and Wendland (1967) and Bryson et al. (1970), who identified a series of distinct quasi-stable climatic episodes divided by abrupt changes. A review of environmentally relevant radiocarbon ages suggested global correlation with the European Blytt-Sernander sequence (Bryson et al. 1970). As a result, Bryson and others argued that the earlier "indistinct transitional model" of Antevs be replaced by the "episodic model" of Blytt-Sernander (Bryson et al. 1970; Reeves 1973). The latter has since found widespread acceptance among Great Plains archaeologists (cf. Gregg 1985a; Greiser 1985). The postglacial climatic episodes of the Blytt-Sernander/Bryson et al. (1970) model are as follows: Late Glacial (pre-10,500 yr B.P., cold and wet); Pre-Boreal (10,500 - 9650 yr B.P., warm and wet); Boreal (9650 - 8450 yr B.P., cold and wet); Atlantic (8450 - 4680 yr B.P., warm and dry); Sub-Boreal (4680 - 2890 yr B.P., cold and dry); Sub-Atlantic (2890 - 1690 yr B.P., warm and dry). Several more recent, post Sub-Atlantic, episodes were subsequently identified by Wendland (1978) as follows: Scandic (ca. 1690 - 1390 yr B.P., warm and dry); Neo-Atlantic (ca. 1390 - 950 yr B.P., warm and moist); Pacific (ca. 950 - 480 yr B.P., warm and dry); and Neo-Boreal (post 480 yr. B.P., cool and moist). Many aspects of this model have proven to be of great utility in paleoclimatic, paleoecological, and archaeological research (cf. Toom 1992b). As previously mentioned, however, subsequent research indicates that Holocene climates were characterized by a higher degree of regional, physiographic, seasonal, and ecological variation than that reflected in the Blytt-Sernander model. Consequently, several components of the model have been criticized (cf. Wright 1974), and the application of European-based chronologies for some regions and ecosystems in North America are disputed (Davis 1984:618). As argued by Holliday (1989b:74), high climatic variability and problems associated with the resolution of previous models, necessitates a focusing of paleoclimatic investigations "on specific time intervals in specific areas." A major goal of this chapter will be to do just that for the Little Missouri region.

Fossil Pollen Analysis

As stated, data relevant to the late Quaternary climatic-morphogenetic history of the THRO South Unit were derived from the analysis of stable carbon isotope composition, stratigraphy, and paleopedology. A fourth potential source of data, fossil pollen analysis, was undertaken but proved
contradictory and potentially unreliable. Nevertheless, the results of the analyses have important implications for future archaeological and paleoecological investigations in the region, and are therefore worthy of mention.

Eolian sediments from upland loess sections were twice examined for the presence of fossil pollen. The results, or at least the interpretation, of these analyses are highly disparate. The first effort, conducted by PaleoResearch Laboratories of Denver, Colorado, was limited to the examination of samples collected from the A horizons of three buried soils at Locality O on Petrified Forest Ridge. These are S2, radiocarbon dated at ca. 11,070 ± 280 yr B.P.; S8, radiocarbon dated at ca. 10,730 ± 460 yr B.P.; and S12, radiocarbon dated at ca. 2160 ± 70 yr B.P. (See section on eolian and lacustrine depositional environments). Using standard processing and counting techniques, Cummings (1990:2) reported that: “the three paleosols yielded similar pollen records to one another, indicating that conditions during the formation of paleosols during the late Pleistocene and early and late Holocene may have been similar.” Noted differences, however, included a greater quantity of *Pinus* pollen in the earliest soil, higher Cheno-am frequencies in the late Pleistocene and early Holocene soils, and greater quantities of *Artemisia*, Compositae, and Gramineae pollen in the late Holocene soil (Cummings 1990). The author goes on to infer relatively low effective moisture during formation of the late Pleistocene soil, warmer conditions and a retreat of pine from the area during the formation of the early Holocene soil, and relatively high effective moisture but a decline in arboreal pollen during the late Holocene period of pedogenesis (Cummings 1990:3). The fossil pollen was counted to “...a total of 100 to 200 pollen grains at a magnification of 430x”, while preservation was reported as varying “from good to poor” (Cummings 1990:1).

The recovery of apparently reliable palynological data from Locality O during the first analytical effort was far from duplicated during a more extensive analysis conducted a year later at the Palynology Laboratory at Texas A&M University (Kuehn 1991). The analysis involved 14 samples collected from the identified stratigraphic units at Locality O, one sample collected from the S1 (early Holocene) soil at Locality L, and one sample from the S1 (late Pleistocene) soil at Locality C (Kuehn 1991). In no instance could traditionally sufficient pollen counts (i.e., from 200 - 300 grains) be achieved (cf. Martin 1963). Indeed, the largest number of pollen grains observed on any one slide was 48, from the modern surface soil at Locality O. The total number of taxonomically identifiable grains from all 16 samples was only 236 (barely sufficient to qualify as reliable even if all the grains came from a single slide). Of these, 33% were recovered from the three most recent (i.e., late Holocene) stratigraphic units, as compared to only 30% from the nine units
dating from the late Pleistocene and early Holocene. This decrease in pollen frequency with time suggests pollen deterioration (Hall 1981). Additional evidence for deterioration is indicated by the percentage of indeterminable pollen grains, which ranged from 28.8% of the total count from the late Pleistocene/early Holocene samples, 12.9% of the count from the middle Holocene samples, and 7.3% of the total count from the late Holocene samples. Remember, these are percentages of already extremely low counts. Pollen concentration values, a frequently useful tool in palynology, were highly discrepant (i.e., ranging from 376 grains/ml to 9,040 grains/ml) and hence, of little interpretive value. Finally, of those grains that could be taxonomically recognized, fully 28.8% were Chenopods, 22% were Poaceae, and 21% were Pinus. All three of these contain distinctive features which make them easily recognizable, even when highly degraded (Bryant and Schoenwetter 1987:38). As recently emphasized by Bryant and Hall (1993) very low overall pollen counts, a high percentage of indeterminable pollen grains, and the predominance of grains with easily recognizable features, cumulatively indicate that the pollen data from a site or group of sites are likely to be suspect. Such is the case with the samples from Localities O, C, and L.

The evidence for decreasing pollen frequencies through time and with depth (coupled with the strikingly low grain counts and the high percentage of indeterminable grains), are factors which suggest poor pollen preservation. This appears to be the result of two characteristics of the natural environment. First, a measurement of pH indicates high alkalinity in each of the sampled sediments. Thirteen of the samples had a pH above 8.0 and three had a pH above 8.5. There is abundant evidence to suggest that such high pH values are not conducive to pollen preservation (Bryant and Hall 1993; Bryant and Holloway 1983; Dimbleby 1957). Second, the localities are all well-drained and appear to have been subjected to frequent wetting and drying. This is indicated by soil permeability (moderately rapid), soil structure (moderate to strong subangular blocky and prismatic), and soil compaction (weak). Repeated episodes of sediment hydration and dehydration exert tremendous pressure on the exine of pollen and are therefore highly destructive (Bryant and Hall 1993; Holloway 1989).

With poor pollen preservation noted in every analyzed sample, and with the natural environment of the sampled localities shown to be highly destructive to fossil pollen, the reliability of the previous interpretations is put into question. The cause of the discrepancy between the two sets of analyses is not known. The lack of data on pollen concentration levels and exact pollen counts from the first analysis makes resolution of the disparity even more difficult. A potential clue, however, may lie in the high percentage of indeterminate pollen grains (41% to 29%) noted by the
previous investigators (Cummings 1990:Figure 1). These high frequencies should be considered as a warning of potential problems with data reliability (Bryant and Hall 1993:281). It therefore appears (based on the suspect nature of the original data and the lack of substantive information from the more recent effort), that fossil pollen may be of little interpretive value to the reconstruction of late Quaternary environments in the badlands region.

**Stable Carbon Isotope Analysis**

In certain light elements the separation, or fractionation, of isotopes is mass dependent. Those elements with greater differences in isotopic mass experience greater fractionation during certain physical, chemical, and biological processes (Herz 1990). As one of these elements, carbon is particularly useful as a climatic/ecological indicator because of the isotopic fractionation that occurs during plant photosynthesis (cf. Cerling et al. 1989). While converting CO$_2$ into carbohydrates, plants follow one of three metabolic pathways that can be seen in the carbon isotopic composition of the plant tissue (van der Merwe 1982, from Herz 1990). These pathways, C3, C4, and CAM, reflect the evolutionary adaptation of plants to different climates. For example, C4 plants convert CO$_2$ into tissue in less time and with less water than do C3 plants. They also discriminate less against $^{13}$CO$_2$ than C3 plants, and therefore end up with lower $\delta^{13}$C values (Park and Epstein 1960). C4 plants thrive in semiarid (and warm) environments such as tropical and subtropical grasslands, while C3 plants thrive in more humid (and cooler), temperate climates such as the forested regions of northeastern North America. The former have an average $\delta^{13}$C value of -12.5%, while the latter have an average $\delta^{13}$C value of -26.5% (Bender 1968, from Kelly et al. 1991). The percentage of C3 and C4 plants in any given area is primarily the result of climatic factors, particularly precipitation and temperature (cf. Boutton et al. 1993; Herz 1990; Nordt et al 1994). Upon decay and transformation into humus, the isotopic composition of plant-derived material changes little (Nadelhoffer and Fry 1988). Therefore, the isotopic composition of soil organic matter is a reflection of the plant community that produced it. Since the C3/C4 biomass of any particular plant community is largely a product of climatic conditions, an analysis of the carbon isotopic composition of soil organic matter (via organic carbon) can be a valuable indicator of both modern and paleoclimatic conditions (cf. Boutton et al. 1993; Cerling et al. 1989).

As described in Chapter III, 50 samples collected from the upland stratigraphic sequences at Localities A, O, C, and L were analyzed for organic carbon isotopic composition at the Texas
A&M stable isotope laboratory (Table 7 and Appendix C). Before discussing the results and implications of these analyses, consideration must be given to potential sources of organic carbon within the samples.

Most isotopic investigation of soils and sediments involves the analysis of organic matter and/or carbonates (cf. Cerling et al. 1989; Kelly et al. 1991). Organic matter in soils is normally considered to be the product of plant (and animal) decay and pedogenic processes; the latter being responsible for the translocation of organic matter down the solum. The measurable constituent of organic matter, organic carbon, can however, enter a soil or stratigraphic profile by non-pedogenic, or inherited, sources (cf. Nordt et al. 1994). The most common inherited source of organic carbon is geologic parent material. Likewise, carbonates, while often pedogenic in origin, can also be inherited from the parent material. As an example, Kelley et al. (1991:1654) demonstrated that much of the disseminated carbonate in six Great Plains soils was inherited from loess parent material, while segregated carbonate was almost certainly pedogenic in origin. In the case of the four South Unit upland sections, the analysis centered on the isotopic composition of organic carbon, as the inorganic (i.e., carbonate) carbon was removed during pretreatment with HCL. The remaining organic carbon reflects both inherited and pedogenic sources, however the former does not appear to be a significant factor in the resultant δ¹³C values, as the organic carbon content of the parent material appears to be quite low. The parent material at all four South Unit localities is eolian in origin (i.e., loess or atmospheric dust) with the possible exception of Unit Ia at Locality A, whose geologic source remains unknown (but likely eolian). The four samples recovered from pedogenically unaltered parent material (D28 and D29 in Unit IIc; D35 and D36 in Unit If at Locality A) have the lowest percentages of organic carbon in the section; ranging from 0.198 to 0.223% (Figure 17; Appendix C). Compared to an average value of 0.703% organic carbon in pedogenic samples (the A horizons of which frequently exceed 1.1%), the low percentages in the parent material samples certainly suggest minimal influence on overall δ¹³C results. As expected, organic carbon is generally highest in surface or subsurface A horizons and progressively decreases down the soil profile (Table 7; Figure 17; Appendix C). Percentages are low in buried soils lacking A horizons, but are still higher than the unaltered parent material. This "normal" pedogenic pattern indicates that the organic carbon δ¹³C values are reflective of the isotopic composition of the vegetative communities that occupied both present and former (i.e., buried) soil surfaces. The values are therefore accepted as potentially reliable indicators of paleoclimatic conditions (cf. Cerling et al. 1989).
As illustrated in Figure 49, the isotopic composition of organic carbon from the two principal South Unit eolian sections demonstrates a steady decrease in C3 biomass from the late Pleistocene through the middle Holocene, and then an increase in C3 biomass in the most recent portion of the late Holocene (i.e., in the A horizons of the surface soil). In broad general terms, the data (together with available chronostratigraphic evidence) indicate that relatively cool and moist conditions predominated from ca. 11,800 to ca. 9000 yr B.P., followed by increasingly warm and dry conditions until sometime after ca. 2100 yr B.P., when the climate witnessed a return to more mesic conditions. This then is the general trend. On the whole it is similar to major paleoenvironmental trends identified elsewhere in the Northern Great Plains (Barnosky et al. 1987). The data, however, do indicate that there are differences in the δ13C values between the two principal sections. They also permit the drawing of some more specific, albeit tentative, inferences concerning paleoclimatic conditions.

The first substantive difference that is evident is the persistence of high negative δ13C values over a greater vertical depth at Locality A. There, negative values greater than -22.1% dominate the lower 2.3 m of the section, as compared to Locality O, where similar values are limited only to the lower 0.9 m (Figure 49). The chronostratigraphic data indicate that this difference is at least partially due to the greater volume of late Pleistocene and early Holocene sediment extant at Locality A. Of the ten identified stratigraphic units, 9 or 90%, are early Holocene or older in age (i.e., greater than ca. 7255 B.P.) (Figure 15). At Locality O, temporally equivalent sediments comprise nine of the 13 units, or 69% of the total (Figure 14). The remaining four units are middle to late Holocene in age (i.e., younger than ca. 6700 yr B.P.). In addition, Locality A has higher overall negative δ13C values in those units dating from the late Pleistocene and early Holocene (i.e., pre ca. 9000 yr B.P.). Samples from these units yielded an average δ13C value of -23.3% as compared to an average value from equivalent units at Locality O of -22.5% (Figures 14 and 15; Tables 7 and 9). This greater volume of C3 biomass could be related to the fact that Locality A has stratigraphic and pedogenic evidence for at least two possible episodes of ponding during the period between ca. 11,830 and 10,480 B.P. While the organic carbon percentages from these units do not indicate bog-like conditions, the presence of standing water (and associated plant species) could account for the higher negative δ13C values at Locality A.
Figure 49. Vertical sequence of $\delta^{13}C$ values of organic carbon at Localities A and O.
As for the drawing of more specific climatic inferences. The peak in low negative $\delta^{13}C$ values evident at 200 cm below the surface at Locality O (Figure 49) could reflect a short-lived period of aridity that has been hypothesized for the western United States during the period from ca. 11,300 to 10,700 yr B.P.; possibly correlating to the Younger Dryas climatic event of northern Europe (Dansgaard et al. 1989; Haynes 1991). The increase in C4 biomass at 200 cm is associated with a stratigraphic unit radiocarbon dated at ca. 10,730 $\pm$ 460 yr B.P. (Table 5; Figure 14). The sudden termination of drought-like conditions was followed by an increase in effective precipitation and a rise in water-tables in the southwestern United States (Haynes 1991). This precipitation increase could be reflected in the high negative $\delta^{13}C$ value at 230 cm below the surface at Locality A (Figure 49); a sample which was collected from a stratigraphic unit radiocarbon dated at 10,480 $\pm$ 82 yr B.P. Perhaps more convincing, however, is the stratigraphic and pedologic evidence for ponding during this period. The peak in high negative $\delta^{13}C$ values at 170 cm at Locality O and 200 cm at Locality A could indicate another return to more mesic conditions, the last of the early Holocene (Figure 49). The increase in C3 biomass is associated with stratigraphic units estimated to be between ca. 9000 and 9500 years of age on the basis of bracketing radiocarbon ages and a radiocarbon date of 9200 $\pm$ 70 yr B.P. from Locality L (Table 5). This time period corresponds to the Boreal climatic episode of the Blytt-Sernander chronology (Bryson et al. 1970:63). It is appropriate to point out, however, that changes in $\delta^{13}C$ values on the order of .1 or .2 per mill are not considered significant, as they fall within the range of the standard deviation resulting from possible measurement error (T. Boutton, personal communication, 1994). On the other hand, a $\delta^{13}C$ value change of 1.0 per mill translates into a ca. 7% change in C3/C4 biomass (T. Boutton, personal communication, 1994). Therefore the peak in negative $\delta^{13}C$ values at 200 cm reflects a 7% to 10% increase in C3 biomass as compared to the $\delta^{13}C$ value of the underlying sample (Figure 49). In a relatively stable grassland community such as the one on Big Plateau, this increase in C3 biomass could indicate a change in climate.

The strongest evidence for changing climate conditions at either locality is the steady decrease in negative $\delta^{13}C$ values after ca. 9000 yr B.P. (Figure 49). There is little doubt that the increase in C4 vegetation reflects a general trend towards warmer and dryer conditions during the middle Holocene. Stratigraphic units IIc at Locality A and II at Locality O yielded radiocarbon ages of 7255 $\pm$ 170 yr B.P. and 6580 $\pm$ 160 yr B.P. respectively, and produced $\delta^{13}C$ values of -19.7% and -20.6% (Figure 49). These data, together with previously mentioned stratigraphic and pedologic evidence, suggest that the deposition, subsequent stability, and erosion of these, and additional units,
occurred under conditions of reduced effective precipitation and increased temperature. What is rather enigmatic, however, is the fact that the lower negative δ¹³C values peak in those units immediately above (i.e., post-dating) the middle Holocene (Figures 14, 15, 49). At Locality A this peak occurs at the bottom of Unit II, estimated to be late Holocene in age on the basis of chonocorrelation (Figures 15 and 49). At Locality O, the peak occurs in Unit IIIb, radiocarbon dated at ca. 2160 ± 70 yr B.P. (Figures 14 and 49). It is important to note, however, that Unit IIIa, radiocarbon dated at ca. 4070 ± 130 yr B.P., demonstrates a 0.6 per mill decrease in negative δ¹³C value over the underlying middle Holocene unit. A similar decrease of 0.4 per mill is evident between the upper-most sample in Unit IIc (middle Holocene) and the lowest sample in overlying Unit III at Locality A (late Holocene). This apparent slight increase in C₄ biomass after the middle Holocene period of aridity could be the result of a number of factors. First, it is possible that the increase in C₄ is due to grazing by large ungulates (i.e., Bison bison). It has been demonstrated that grazing can significantly alter the vegetative community in any given area (French 1979). In the mixed grass prairie of the Northwestern Great Plains, heavy grazing could result in the elimination or significant decrease of C₃ mid and tall grasses, leaving the C₄ short grasses (T. Bouton, personal communication, 1994). This change would then leave its signature on the stable carbon isotopic composition of soil organic matter. The increase in C₄ values in the post-middle Holocene soils could suggest that bison were absent, or much less frequent, in the badlands during periods of severe aridity, but then returned in significant numbers when drought conditions ameliorated.

Dillehay (1974) hypothesized the total disappearance of bison from the Southern Plains during the Altithermal. Subsequent research indicates that bison populations in the Southern Plains were indeed reduced during two episodes of middle Holocene aridity (as the result of reduced vegetative cover); the reductions, however, were not as universally great as that proposed by Dillehay (Holliday 1989b; Johnson and Holliday 1986). Further north in the Central High Plains, Greiser (1985:39) hypothesizes that, during the Altithermal "...large herbivore populations were more drastically reduced than other species", with bison populations being more seriously affected than antelope. In the Northern Great Plains, no specific research has been conducted on bison populations during the arid portions of the middle Holocene, although Frison (1978) suggests that bison (and humans) may have been forced into oasis-like areas such as the Black Hills. Morgan (1980) presents evidence that bison on the Canadian Plains followed a two-field rotation system, or annual migrations between summer and winter ranges, because of seasonally-induced changes in forage availability. She also admits that significant climatic fluctuations could initiate significant
changes in grassland vegetation (and hence changes in average migration patterns) (Morgan 1980:147). Hanson (1984) on the other hand, argues that in the North Dakota region, bison did not follow large seasonal migrations, but were present in both large and small herds in all grassland zones throughout the year. He does hypothesize, however, that drought conditions likely forced bison herds into refugia areas such as major river valleys (Hanson 1984:110). Lastly, Bamforth (1987) stresses that studies of bison movement must take into account specific climatic and ecological conditions due to a direct link between annual forage production and precipitation. Therefore, bison, like all large ungulates, can be expected to significantly alter their otherwise relatively predictable patterns of movement in response to climatically-induced changes in forage production (Bamforth 1987).

While hard evidence one way or the other is absent, conditions of severe aridity during the Middle Holocene could certainly have forced bison out of the badlands region. Their return in significant numbers could also have reduced the amount of C3 biomass in the study area, resulting in low negative δ¹³C values in soil organic matter. The low values, however, could also result from other natural agents such as fire (cf. McPherson et al. 1993; Nordt et al. 1994). A third possibility is that erosion has removed significant amounts of middle Holocene sediment, and therefore the complete carbon isotopic record from this period is not available. Indeed, all of the middle Holocene stratigraphic units are capped with truncated Bk soil horizons, indicating the erosional loss of at least one A horizon in each unit, and perhaps even entire units. These erosional episodes, in the form of wind deflation, are likely to occur during the most severe portions of droughts, when vegetative cover is the most significantly reduced (cf. Clayton et al. 1976). This is precisely the time when C4 values would be expected to be greatest. The irony then, is the possibility that the isotopic evidence for severe aridity may have been lost as the result of that aridity.

Finally, the 0.5 per mill peak in low negative δ¹³C values at 35 cm below the surface at Locality O, and the similar 0.7 per mill peak at 10 cm below the surface at Locality A, could reflect a brief period of warm and dry climate that coincides with the Sub-Atlantic climatic episode of the Blytt-Sernander chronology (Bryson et al. 1970). Although highly tentative, this correlation is possible on the basis of the radiocarbon age of 2160 ± 70 yr B.P. that was obtained from bulk soil organics in Unit IIIb at Locality O (which is the unit that produced the low negative δ¹³C value) (Figure 49). The Sub-Atlantic episode has an estimated temporal range of ca. 2890 to 1690 yr B.P. (Bryson et al. 1970:63). The return to relatively high negative δ¹³C values at the top of both sections
is consistent with the composition of the modern vegetative communities at the two localities, which are both decidedly C3 dominated (T. Boutton, personal communication, 1994).

**Climatic-Geomorphic Interpretation**

The discussion thus far has centered on continental and global climatic conditions subsequent to the last advance of the Laurentide ice sheet, and on the isotopic evidence for general climatic trends at the local level during the last ca. 12,000 years. Prior to that, the stratigraphic and pedologic composition of those packages of sediment associated with major late Quaternary depositional environments were described. These deposits are largely the product of unique geomorphological processes (Reineck and Singh 1980). Because of the close association between geomorphology and climate (cf. Butzer 1976b; Derbyshire 1976), additional inferences on paleoclimatic conditions in the study area can be drawn from a re-examination of the previously described stratigraphic and pedologic data. This re-examination can best be accomplished by utilizing all available data in a diachronic reconstruction of the late Quaternary climatic and geomorphic history of the study area. Similar pluralistic approaches are commonly utilized during environmental reconstruction for geological (Reineck and Singh 1980:4-6), and archaeological (Hassan 1985b:89), purposes.

Before embarking upon such a reconstruction, it is necessary to first examine some relevant aspects of the relationship between climate and geomorphology. As mentioned, climate, particularly insolation and moisture, has a direct influence on vegetation. Together both climate and vegetation are the driving forces behind the weathering, erosion, transportation, and deposition of sediment (Butzer 1976b; Hassan 1985b). Landscape form is therefore determined by the balance between deposition, erosion, and stability (Butzer 1976b; Hassan 1985b). Large-scale landform characteristics, however, are controlled by rock structure and lithology, just like the specific composition of a particular vegetative community is also controlled by rock type, drainage patterns, and hillslope processes. As summarized in Butzer (1976b:333), Erhart's (1967) model of biostasis and rhexistasis, although dated and somewhat simplistic, is appropriate to the present discussion. During periods of landscape stability, hillslopes are covered by a protective mat of vegetation and there is a balance between weathering (soil formation) and denudation. Weathering during stable episodes produces a relatively continuous mantle of soil. During periods of landscape instability, vegetation is significantly reduced, if not completely eliminated. On slopes, disequilibrium results
when erosion outpaces soil formation; resulting in denudation and the truncation or complete loss of soils. The removal of vegetation and soils leads to the lateral transfer of sediment to lower slopes and valley bottoms. Large-scale alterations between landscape stability and instability can be caused by climate change, eustatic sea level fluctuations, tectonic activity, and human activity (Butzer 1976b). At the scale of individual geomorphic landscapes, change can also occur as a result of complex response to intrinsic variables (Schumm 1973). It was previously argued that in the case of the tectonically stable Little Missouri Badlands, large-scale changes in fluvial landscapes are the result of fluctuating climatic conditions, although as mentioned earlier, myriad variations in sediment yield can lead to complex response in the smaller drainage basins or in ephemeral tributaries. While there is also a great deal of small-scale variation in eolian (i.e., upland) landscapes, area-wide changes in these settings appear to be climatically driven as well.

If climate is the driving force behind landscape stability and instability in the badlands as a region, what are some of the more salient aspects of the process-response association between climate and sediment deposition, erosion, and weathering? As demonstrated by Langbein and Schumm (1958), runoff is most directly affected by precipitation, but temperature is also a key factor because of its influence on evapotranspiration. The role of climate and vegetation in controlling the amount of runoff generated by surface flow (i.e., slopewash as defined for our study area) is critical, as surface flow is a major contributor to erosion and aggradation in fluvial systems (Knox 1984a, 1984b). In semi-arid regions of the western United States, including the Little Missouri Badlands, the affects of temperature and precipitation on runoff and sediment yield are particularly acute (Knox 1984b; Langbein and Schumm 1958). In these regions, rather modest changes in annual precipitation (i.e., 10%) and temperature (i.e., 2 degrees C) can result in dramatic changes in mean annual sediment yield. As illustrated in Knox (1984b:Table 4) climatic variations of the magnitude just described can influence sediment yield in semi-arid regions in the following manner: (1) cool/wet conditions = major reduction; (2) warm/wet conditions = little change to modest reduction; (3) cool/dry conditions = no change; and (4) warm/dry conditions = major increase. Therefore, periods of warm and dry climate are most likely to be associated with episodes of eolian, fluvial, and slopewash aggradation, as hillslopes and upland surfaces are stripped of vegetation and sediment is eroded, transported, and redeposited. Grassland vegetative communities are particularly susceptible to drought conditions, with decreases in effective cover as high as 80 to 95% reported during the onset of droughts in the Central Great Plains (Brice 1966; Knox 1984a:29; Tomanek and Hulett 1970).
These basic climatic-geomorphic principles, including Erhart's (1956) and Butzer's (1976b) hypotheses on landscape stability, are clearly evident in the model developed by Clayton et al. (1976:8-9) for climate, vegetation, and stability in the North Dakota region. The model formed the conceptual basis for their interpretation of the depositional environments associated with the Oahe Formation. Its central theme is as follows:

By analogy, it is probable that dryer periods in the northern Great Plains during the late Quaternary were periods when steep hillslopes were unstable, when valley bottoms were alluviating, when a large amount of sediment was available to the wind, when dust storms were numerous, and when silt was actively being deposited on gentle slopes. In contrast, during more moist periods, steep hillslopes were stable, rivers were downcutting, little sediment was available to the wind, there were few dust storms, and soils were formed (Clayton et al. 1976:8).

The available South Unit data suggest that this relatively simple model is remarkably well suited to the Little Missouri Badlands. It will therefore form the framework for the subsequent discussion on late Quaternary climate and large-scale changes in geomorphic landscapes. As with the stratigraphic and pedological investigations, the reconstruction focuses on those landscapes associated with eolian (i.e., upland), and fluvial/slopeswash (i.e., lowland) depositional environments.

Approximately 24,000 years ago the study area was in the grip of glacial climatic conditions. The Little Missouri River flowed over a surface that lies some 90 m above the current river level. This surface is 54 m lower than the Miocene/Pliocene surface on adjacent Petrified Forest Plateau, a figure that provides a rough estimate as to the depth of the Little Missouri valley at that time. Following a relatively mild and moist interval around 23,000 yr B.P., the climate experienced cooling on the order of between 0.6 and 1.0 degrees C per 1000 years (Harmon et al. 1979, from Barry 1983). By ca. 17,000 B.P., the mean annual temperature in the study area was somewhere between -7°C and -10°C (Barry 1983). The last advance of the Laurentide ice sheet to extend into southwestern North Dakota and hence have a direct influence on the geomorphology of the study area reached its maximum extent about 20,000 yr B.P. This advance, termed Phase D by Clayton and Moran (1982), is poorly documented in western North Dakota, although it may correlate with
the well-known Napoleon and Long Lake advances in central North Dakota, and with ice-margin MT3 in northeastern Montana (Clayton and Moran 1982:65; Mickelson et al. 1983:9).

It is interesting to note that a map showing the maximum extent of Phase D ice at ca. 20,000 B.P. (Clayton and Moran 1982:Figure 4), shows a proglacial lake in the valley of the Little Missouri River south from the southern margin of the ice sheet in the vicinity of "The Bend" (Figure 2). A similar lake, this one extending south from "The Bend", but also east to near the mouth of the Little Missouri, is illustrated in Bluemle (1991:Figure 67) but is not associated with a specific ice-margin advance. Although little substantive data are available as to the age and spatial extent of these lakes (cf. Bluemle 1991:70), both references to proglacial damming of the Little Missouri valley could have important implications concerning the downcutting associated with abandonment of the Pleistocene cut terraces described in the previous chapter. As argued, this downcutting appears to have occurred after ca. 24,000 B.P., based on a radiocarbon age recovered from the highest terrace surface. This, together with the age of ice margin D at its maximum extent, suggests that the incision was related to late Wisconsin glacial activity. It was noted earlier that traditional theory holds that the downcutting was triggered by the eastward diversion of the Little Missouri in "The Bend" area by an advancing ice sheet (Bluemle 1972, 1977, 1991; Laird 1950; Leonard 1916). While the geological literature is ambiguous as to the timing of this diversion, Bluemle (1991:5,16) most recently estimated that it occurred between 600,000 and 700,000 years ago. This estimate is incongruent with the 24,000 B.P. radiocarbon age recovered from the Pt4 terrace sediments on Big Plateau; a factor which indicates that either: (1) the 600,000 to 700,000 B.P. estimate is incorrect; (2) the 24,000 B.P. radiocarbon age is in error; or (3) that the downcutting of the Little Missouri was not triggered by the major episode of glacial diversion. What is obvious is that the incision was caused by a rapid and significant lowering of the Little Missouri River base level. Although not stated, Bluemle's estimate is likely based on chronostratigraphic data recovered from correlated glacial deposits. If these data are reliable, the stated time of the original (i.e., major) diversion could indeed be accurate. On the other hand, the 24,000 radiocarbon age, and Clayton and Moran's (1982) estimate for the timing of the maximum extent of ice margin D (i.e., 20,000 B.P.) appear too closely related to be mere coincidence. Therefore, the 24,230 yr B.P. radiocarbon age does appear to accurately reflect a time of channel lag deposition prior to rapid downcutting. If this is the case, the long-assumed link between the major episode of glacial diversion and the rapid lowering of the Little Missouri base level may no longer be an a priori assumption. As illustrated on the Figure 2 geologic map, a large abandoned, southeast-trending, river channel is located near the present-day mouth of
the Little Missouri (Clayton 1980). It is possible that the river flowed through this channel after the first glacial diversion, but over a gradient that was not significantly steeper than that of the pre-diversion channel. Therefore the diversion may not have led to the downcutting which initiated the formation of the badlands. If this is correct, then that short portion of the Little Missouri between the abandoned channel and the present-day mouth must have been established as the result of a more recent ice margin advance (i.e., ice margin D?). This smaller diversion may have been the actual event responsible for the rapid lowering of the Little Missouri base level.

There is, however, another possible scenario. It is rather puzzling that large-scale badlands are limited only to the Little Missouri River and its tributaries, and not to other tributaries of the Missouri River. This suggests that the other tributary streams did not experience a similar rapid lowering of base level even though the underlying parent material is identical throughout the region. It also suggests that the base levels of the Little Missouri and Missouri Rivers may not have been congruent. This is where the evidence for proglacial lakes becomes relevant. If a lobe associated with the Phase D ice margin dammed the mouth of the Little Missouri, a proglacial lake would have been created in the upper end of the valley, and lowering of the Little Missouri base level would have temporarily ceased. If meanwhile, the Missouri River below the mouth of the Little Missouri was not dammed, it could have continued to downcut, perhaps even rapidly if there was a great deal of meltwater discharge. When the glacial ice retreated from the region, the damming of the Little Missouri would have ended, and its base level would have been significantly higher than that of the downstream Missouri. Rapid downcutting of the Little Missouri in a time-transgressive progression upstream would have followed as its base level adjusted to that of the trunk stream.

While there is a dearth of stratigraphic and chronologic information concerning late Wisconsin proglacial lakes in western North Dakota, such information is not totally absent. Bluemle (1972:2191-2192), citing morphologic and lithologic data, suggests that a late Wisconsin proglacial lake was present in the Missouri River valley a short distance upstream from the Little Missouri mouth. If this is correct, a different version of the above mentioned scenario could have occurred. That is, the ice-walled lake could have stopped or significantly reduced discharge into that portion of the Missouri downstream from the dam. A breaking of the dam by ice-margin retreat could have resulted in rapid incision and lowering of the Missouri base level downstream as a result of the sudden release of water. If this base level lowering was limited only to a short portion of the Missouri valley that included the Little Missouri mouth area (because of pre-existing valley positions
and differential ice-margin locations), the Little Missouri River may have been the only significant tributary forced to rapidly adjust its base level.

Although these scenarios are offered only as possibilities, what is fairly certain is that sometime after ca. 24,000 yr B.P. the Little Missouri River experienced at least three episodes of downcutting and lateral migration prior to the base level maintained during the Holocene T4 period of aggradation (ca. 6000-7000 yr B.P.). These terrace surfaces and their associated channel lag gravels are the only evidence of fluvial activity during the period between approximately 24,000 and 7000 yr B.P. The 90+ m of Little Missouri incision after 24,000 B.P. led to badland formation as tributary streams downcut to meet the new trunk stream base level, and as headward erosion caused the dissection to encroach upon the unglaciated Missouri Plateau. During this period, newly created hillslopes experienced both downwearing and backwearing on a massive scale. Hillslope lowering was not uniform, however, because of lithologic variations in the Paleocene bedrock (cf. Campbell 1989). These, and other badland processes such as gullying, piping, and mass wasting, were also operative (and still are). Blumle (1991:16) estimates that, since their inception, 40 cubic miles of sediment have been eroded and removed from the Little Missouri Badlands.

While the basic geomorphological trend after the initiation of rapid downcutting was one of general degradation (particularly valley incision and hillslope lowering), the evidence of at least four Pleistocene terrace levels, and their relatively broad spatial extent, indicates that the periods of incision were followed by channel migration. Downcutting, therefore, was not constant, and valleys may have experienced some brief periods of stability and even aggradation. The lack of fluvial sediment other than channel lag, however, suggests that degradation far outpaced any significant aggradation, especially vertical accretion. On the other hand, the lack of recognizable point bar sediments, in light of the evidence for channel migration, suggests that some fluvial sediment must have been eroded after the surfaces were abandoned by downcutting.

As fluvial incision and badland formation continued, portions of the Missouri Plateau surface were spared denudation. Some of these remain articulated to the Missouri Plateau, extending into the badlands in peninsula-like fashion, and others were isolated from the Plateau by erosion, standing today as buttes or ridges. Other upland surfaces adjacent to the Little Missouri River represent the previously discussed Pleistocene terraces.

Elevational differences between the Miocene-Pliocene surface extant on Petrified Forest Plateau (822 m asl), the P4 terrace surface on Big Plateau and Johnson Plateau (768 m asl), and the bottom of the active Little Missouri channel (676 m asl), indicate that 63% of the total vertical relief
associated with the Little Missouri Badlands was established within the last ca. 24,000 years (based on the radiocarbon age recovered from the Pt4 terrace sediments on Big Plateau). These elevations also suggest that prior to the abandonment of the Pt4 surface, the Little Missouri River flowed through a valley that was ca. 50 - 60 m deep. This is similar to the depth of present-day valleys in the middle to lower reaches of other streams in western North Dakota (i.e. the Heart and Cannonball Rivers) (cf. Clayton 1980 map). These modern drainage systems are not associated with large-scale badland development, hence there is no reason to believe that significant badlands existed within the Little Missouri drainage prior to the abandonment of Pt4. In addition, since no temporal data are available concerning the upper 54 m of surface lowering (i.e., the lowering from the Miocene/Pliocene surface to the surface of the Pt4 terrace), estimates concerning long-term denudation rates are provided by measuring the amount of incision that occurred after abandonment of the Pt4 surface. Since the vertical distance between this surface and the bottom of the active Little Missouri River channel is 92 m, and the known time span associated with this downcutting is less than 24,000 years, a denudation rate of at least .38 cm/\text{year} is indicated. The rate is even greater if the estimate is based solely on the amount of incision that occurred between the time of the Pt4 abandonment (ca. 24,000 yr B.P.) and the first subsequent recognizable episode of aggradation (i.e., deposition of the T4 terrace fill beginning at ca. 7000 years B.P.). The vertical distance between the Pt4 surface and the base of the T4 fill (ca. 85 m) and the time interval between the two events (ca. 17,000 years) suggest an average rate of lowering on the order of 0.5 cm/\text{year}. Both of these rates (0.38 cm/\text{yr} and 0.50 cm/\text{yr}) are not unlike those reported from other badland regions in North America with similar bedrock lithologies. These include the badlands of the Chaco drainage basin in New Mexico (0.3 - 0.5 cm/\text{yr}), the Steevesville Badlands in Alberta, Canada (0.46 - 2.98 cm/\text{yr}), and the White River Badlands of South Dakota (1.01 - 1.90 cm/\text{yr})(Campbell 1970; Schumm 1956b; Wells and Gutierrez 1982).

Except for the lag gravels capping the Pt1-Pt4 terraces, sedimentation in the uplands is not evident until ca. 12,000 yr B.P., when ponding in at least one closed upland basin resulted in the deposition of carbonate-rich muds. As stated earlier, the pond sediments themselves most likely have an eolian origin, with deposition on the surface of the pond associated with increased rainfall, which was responsible not only for pond formation, but also for carrying the fine grained eolian sediment out of atmospheric suspension (Greeley and Iversen 1985). The timing of this episode (ca. 12,000 to 11,500 B.P.), as indicated by a radiocarbon age from the lowest identified pond sediment (Unit 1b) at Locality A. corresponds with the evidence for increased discharge and effluent
conditions in the southwestern United States immediately prior to ca. 11,200 yr B.P. (Haynes 1991:441). In the prominent pluvial lakes of the west and southwest, there is abundant evidence for a brief period of lake level expansion around 12,000 yr B.P. (Smith and Street-Perrott 1983:205). Closer to the study area, pollen records from lakes in the Northwestern Great Plains of Montana indicate mesic conditions from ca. 12,200 to 11,500 yr B.P. (Barnosky 1989). These conditions are substantiated by the stable carbon isotope data from Locality A, where the pond sediments exhibit high negative $\delta^{13}$C values (-23.4‰).

The episode of ponding at Locality A was followed by a brief period of stability, as indicated by the presence of a weakly developed Entisol at the top of Unit 1b2. Subsequent to this, but prior to ca. 11,000 yr B.P., the first recognizable episode of loess deposition occurred in the uplands. This episode is represented by Aggie Brown Units 1a and 1b at Locality O, and Unit 1c at Locality A. Clayton et al. (1976:8) and Jorstad et al. (1986) argue that significant loess deposition associated with the eolian lithofacies of the Oahe Formation occurred under conditions of dryer climate resulting from increased sediment availability and an increase in dust storms. While the paleoenvironmental data from Middle Holocene loess units in the study area do tend to support this argument, the late Pleistocene and early Holocene deposits, like those represented by Unit 1b at Locality O, have relatively high negative $\delta^{13}$C values (i.e., -23.8‰) which are not indicative of xeric climatic conditions. Therefore, the causal factors behind early loess deposition in the study area are not yet fully understood.

The pre-11,000 B.P. episode of loess deposition appears to have occurred rapidly and was followed by a second, more substantial, period of upland soil formation which was terminated by burial sometime shortly after 11,000 years ago. This temporal estimate is based on a radiocarbon age recovered from very near the top of the S2 A horizon at Locality O (Table 5). Here the period of stability associated with S2 soil formation was followed by multiple episodes of loess deposition punctuated by short periods of stability and the formation of weak soils prior to ca. 10,700 B.P. Each of the units associated with these episodes of loess deposition is topped with a thin A horizon. The upper-most of these, S8, is associated with an apparent increase in C4 biomass, as indicated by a peak in low negative $\delta^{13}$C values (-20.4‰)(Figure 49). This factor, coupled with radiocarbon ages of ca. 10,700 yr B.P. from Localities O and C, suggests that pedogenesis occurred during an episode of relatively xeric climatic conditions. As mentioned, this period is temporally equivalent to the Clovis-age drought reported by Haynes (1991) from several archaeological sites in the Southwest and Great Plains.
A brief return to cooler/moister climatic conditions prior to ca. 10,500 B.P. is indicated by a second episode of ponding evident at Locality A, and by the high negative $\delta^{13}C$ values (i.e., -23.0 to -23.3%) associated with the pond sediments (Figure 15). The episode corresponds to a mesic period and water table rise in the southwest subsequent to the previously mentioned Clovis drought (Haynes 1991). The deposition of lacustrine sediment (Unit ld2) was followed by a period of stability that ended shortly after 10,500 B.P., as indicated by a radiocarbon age on charcoal recovered from the near the surface of the S3 soil at Locality A (Table 5). Similar evidence of ponding was not recovered from the other upland sections. Factors of variable sediment preservation aside, Locality A is the only upland section with evidence of standing water. All of the other temporally equivalent sections were apparently well-drained. It is assumed that while ponding and lacustrine deposition was occurring in poorly drained areas like Locality A, other upland locations were experiencing stability and soil formation. As previously mentioned, there is no available evidence for landscape conditions in the lowlands during this time period.

Loess deposition resumed shortly after 10,500 B.P. At Locality O this deposition is associated with a thick eolian unit (Unit li). Deposition ceased shortly thereafter, and by approximately 9200 years ago the uplands were experiencing a period of landscape stability. This pedogenic episode is represented by buried soils at Localities A, O, C, L, and K (Figure 23). Bracketing radiocarbon ages and an age of 9200 yr B.P. from Unit I at Locality L suggest that the stable episode may have been synchronous with the so-called Boreal climatic episode, a period of cold and wet climate between 9650 and 8450 yr B.P. (Bryson et al. 1970). Additional evidence for mesic conditions during this period is suggested by the increase in C3 biomass evident in Units li at Locality O ($\delta^{13}C$ value of -21.8%) and le and If at Locality A ($\delta^{13}C$ values of -23.6% and -23.5%). The latter (Unit If) also exhibits a clay increase, possibly indicating yet a third episode of ponding (Tables 6 through 9).

Data from some of the key sections tend to suggest that significant periods of pedogenesis in the uplands were associated with generally mesic climatic conditions; an observation first offered by Jorstad et al. (1986) on the basis of their research on Cinnammon Creek Ridge. As we have seen, there are, however, middle Holocene exceptions to this pattern.

The abundance of eolian stratigraphic units dating from ca. 11,000 to 9000 B.P. is consistent with the pattern of general landscape instability throughout much of North America during the late Pleistocene, as vegetative communities and geomorphic landscapes adjusted to changing climatic conditions associated with the retreat of continental glaciation (cf. Schumm and Brakenridge
1987:238). There were, however, periods of stability, as indicated by prominent Leonard-aged paleosols. The presence of a Mollisol at the top of Unit Ib at Locality O is a strong, but not infallible, indicator of grassland vegetation in the area by ca. 11,000 B.P. (cf. Fenton 1983:129).

As stated, the stable carbon isotope data indicate a general trend toward warmer/dryer conditions from the late Pleistocene until relatively recent times. The stratigraphic and pedologic data, however, indicate that xeric conditions were most severe during the middle Holocene. Contrary to earlier models which recognized a single Altithermal episode of extremely long duration (Antevs 1955), the THRO data suggest that there were multiple warm/dry periods during the middle Holocene. This evidence is more in keeping with the "two-drought" Altithermal recognized by Benedict (1979) and Benedict and Olson (1978) for the western United States, and by Johnson and Holliday (1986) for the Southern Great Plains. Indeed, Holliday (1989b) more recently acknowledged that even the "two-drought" hypothesis is probably overly simplistic.

In the uplands, the period from ca. 8500 to 7000 B.P. may have witnessed as many as three episodes of loess deposition and subsequent stability. These landscape conditions are particularly evident at Locality A (Units IIa, IIb, IIc). As stated in chapter V, each of the units is capped by an eroded Bk soil horizon indicative of pedogenesis under conditions of reduced precipitation.

The middle Holocene landscape pattern for upland settings appears to have been one of loess deposition during periods of increased aridity, followed by landscape stability under less severe, but still relatively xeric, conditions, and subsequent erosion (via wind deflation) as more severe drought conditions returned (cf. Holliday 1989b; Reeves 1966). Archaeological excavations at numerous upland locations in the badlands suggest that episodes of wind deflation during this period were widespread. In every documented instance, field investigators encountered sediments dating to less than ca. 5300 years old directly overlying Paleocene bedrock (Artz 1992; East et al. 1985; Jorstad et al. 1986; Kuehn 1990, 1993; Wyckoff 1982). The THRO data indicate that older (i.e., late Pleistocene to early Holocene) sediments are limited to small, isolated basins which served to trap sediments except under the most extreme erosional conditions (Kuehn 1993:329).

Significant, but apparently nonsynchronous, episodes of loess deposition and wind deflation in the uplands continued during the period from ca. 7000 to 6000 yr B.P. This interval also witnessed periods of substantial hillslope erosion, which in turn led to alluvial fan deposition and the earliest recognizable episode of Holocene fluvial aggradation in the study area (Table 9). The latter is represented by the poorly preserved T4 terrace fill, evident in the larger perennial and
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intermittent stream valleys, and associated with radiocarbon ages of 6115 and 6660 yr B.P. (Table 5). Synchronous fan deposition is evident at Knutson Fan 93F, which contains a shallowly buried soil radiocarbon dated at ca. 5900 yr B.P. This suggests that most of the fan sediment was deposited prior to this time of soil formation. Loess deposition during this period is represented by Unit II at Locality O (Table 5), and by Pick City Member units extant at other upland localities (Figure 23). The S10 episode of pedogenesis again occurred under conditions of reduced precipitation (as indicated by heavy accumulations of carbonates). This middle Holocene pedogenic episode was followed by a period of wind deflation and upland erosion, evident by disconformable unit contacts at several of the upland sections (Figure 23). The erosion may signal a return to more severe drought conditions and the concurrent reduction or elimination of vegetative cover. A radiocarbon age from the buried soil at Knutson Fan 93F tentatively suggests that by ca. 6000 yr B.P. the area may have witnessed another period of landscape stability. The variable and incongruent nature of paleosols in the study area, however, indicates that caution should be taken when attempting to infer area-wide landscape conditions on the basis of a single soil, especially one in the highly dynamic lowlands.

During the latter portion of the middle Holocene (i.e., from 5500 - 4500 yr B.P.), loess deposition is not yet evident in the South Unit stratigraphic record, perhaps because of a predominance of wind deflation. Eolian sediments dating from this period, however, are reported from several other localities in the badlands and adjacent areas (Jorstad et al. 1986; Kuehn et al. 1987; Simon et al. 1982). Meanwhile, in the lowlands, high sediment yield from eroding hillslopes resulted in the second recognizable episode of Holocene fluvial deposition, the T3 terrace fill. The time of this aggradational episode is tentatively estimated at between 5500 and 4500 yr B.P. (See Chapter V). The T4 and T3 periods of aggradation recognized in the THRO study area generally correspond to the warm and dry Wolf Creek Unstable Episode (ca. 8800 - 4700 yr B.P.) as defined for the North Dakota region by Bluemle and Clayton (1982). They are also roughly synchronous with episodes of alluviation in the White River Badlands of South Dakota, which were prevalent from ca. 10,000 to 5000 yr B.P. (White and Hannus 1985).

The Wolf Creek Unstable Episode, like Antev's (1955) model of the Altithermal, is represented as a more or less continuous episode of warm and dry climate, although the drought-like conditions were time-transgressive from southwest to northeast (Bluemle and Clayton 1982:Figure 8). Again, the single-drought model is not consistent with the THRO data, which clearly indicate that climatic conditions fluctuated during the middle Holocene. While the stable carbon isotope record reflects a rather uniform increase in C4 biomass after ca. 8500 yr B.P. (Figure 49), the
stratigraphic and pedogenic evidence suggest that the uplands experienced as many as three episodes of loess deposition and subsequent soil formation between ca. 8500 and 7000 yr B.P., and at least one episode of deposition and soil formation between ca. 7000 and 6000 yr B.P. The THRO data also indicate that the lowlands experienced two separate episodes of fluvial aggradation during this period, one from ca. 7000 to 6000 yr B.P. (T4) and one from ca. 5500 to 4500 yr B.P. (T3). The 2.5 to 3.5 m of downcutting associated with the abandonment of the T4 surface suggests (in lieu of other extrinsic variables) that sediment yield to the lowlands decreased significantly sometime between ca. 6000 and 5500 yr B.P. Given the above described relationship between geomorphology and climate, this decrease most likely resulted from an increase in effective precipitation. In other words, drought-like conditions appear to have ameliorated at least temporarily. In light of this new evidence, it is suggested that the Wolf Creek Unstable Episode in southwestern North Dakota be subdivided into two episodes of warm and dry climate; one (the Lower Wolf Creek Unstable Episode) from ca. 8500 to ca. 6000 yr B.P., corresponding to the deposition of the oldest Pick City loess units and to the aggradation of the T4 terrace fill, and one (the Upper Wolf Creek Unstable Episode) from ca. 5500 to ca. 4500 yr B.P., corresponding to the deposition of the youngest Pick City loess units and to the aggradation of the T3 terrace fill (Table 9). The unnamed period from ca. 6000 to 5500 yr B.P. is hypothesized as an interval of relative landscape stability in the uplands and fluvial degradation in the lowlands. Like Holliday's (1989b) interpretation of middle Holocene climatic conditions in the Southern Great Plains, the proposed division of the Wolf Creek Unstable Episode in southwestern North Dakota into two events is likely to prove overly simplistic as more paleoenvironmental data become available. Nevertheless, the division is in keeping with the realization that climatic conditions during the middle Holocene were more variable than the earlier "single-drought" models would suggest (cf. Benedict 1979; Benedict and Olson 1978; Holliday 1989b; Johnson and Holliday 1986).

The multiple periods of middle Holocene drought ended in the badlands region by ca. 4500 yr B.P. The presence of organic-rich soils associated with radiocarbon ages from Unit IIIa at Locality O and Knutson Fan 93B indicate that a significant period of landscape stability may have occurred in both upland and lowland settings between ca. 4500 - 3800 yr B.P. This return to presumably cooler and more stable climatic conditions corresponds to the early Sub-Boreal episode of the Blytt-Sernander chronology (Bryson et al. 1970). It is also synchronous with stable climatic conditions in the Midwest (Knox 1984a; May 1986). In North Dakota, the event falls within the generally cool and more moist Thompson Stable Episode (ca. 4700 to 2700 yr B.P.) as described by
Bluemle and Clayton (1982:13-14). The South Unit soils are equivalent to the lower Thompson Paleosol(s) of the Riverdale Member, which are believed to have formed under mesic conditions (Clayton et al. 1976). While the existing δ13C record from this period is inconclusive, the temporal estimates established for the Thompson Stable Episode are supported by the THRO stratigraphic and pedologic data.

With the deposition of Unit IIIa at Locality O and Knutson Fan 93B likely occurring prior to ca. 4500 B.P. (during the last middle Holocene drought), sediment deposition in the study area is not evident again until sometime prior to 2500 yr B.P. While data are sketchy, the absence of deposition suggests that badland landscapes during this long interval were undergoing simultaneous stability and erosion. More specifically, the uplands appear to have witnessed stability and soil formation, while the lowlands were experiencing significant degradation. This hypothesis is supported by the widespread occurrence of Thompson-aged paleosols throughout the uplands (Figure 23), and by the paucity of T3 and T4 fluvial sediment in the lowlands. In their late Quaternary paleoenvironmental reconstruction for the North Dakota area, Clayton et al. (1976:Figure 7) estimate that the lower Thompson paleosols formed during a period when mean annual precipitation was 54% greater and mean annual temperature was 2.5 to 3 degrees C cooler than the preceding warm and dry middle Holocene. Following the climatic-geomorphic model previously described, (i.e., Knox 1984b: Table 4), this significant increase in precipitation and decrease in temperature is likely to have triggered a major reduction in sediment yield, which in turn led to stream entrenchment and fluvial degradation. This degradational episode may have been responsible for widespread erosion of the T3 and T4 terrace fills. Synchronous episodes of stream entrenchment are recognized throughout the Short-Grass Prairie and Western Great Plains (Albanese and Wilson 1974; Knox 1984a:35), and Midwest regions (May 1986). In the White River Badlands, fluvial erosion dominates the period from ca. 5000 to 2500 yr B.P. (White and Hannus 1985). Along the Missouri River in northwestern North Dakota, incision is documented for the interval between 3500 and 3100 yr B.P. (Coogan 1983). In the badlands, the degree to which older fluvial and alluvial sediment was voided appears related to factors of drainage basin size, valley geometry, and discharge (cf. Albanese 1978; Gonzalez 1987; Hadley and Schumm 1961).

Sediment deposition in the study area resumed by ca. 2500 yr B.P. In the uplands, this deposition is represented by Unit IIIb at Locality O, and by several undated loess units at other localities. In the lowlands, it is represented by the fairly extensive T2 terrace fill; the deposition of which is estimated to have occurred between 2700 and 800 yr B.P. (see Chapter V). Synchronous
fan formation also took place during this period, as indicated by facies relationships at Locality CC in Knutson Creek, and by fan/terrace correlations at other localities. In the uplands, loess deposition terminated and was followed by landscape stability by ca. 2200 B.P. This pedogenic episode is represented by numerous Thompson-aged paleosols scattered throughout the study area. A second episode of stability may have occurred in both the uplands and lowlands around 1500 yr B.P., based on the presence of a radiocarbon-dated Mollisol at Locality B, and a similar radiocarbon age from a soil in the T2 terrace fill at Locality P. Again, this episode is highly tentative as it cannot be automatically assumed that pedogenesis occurred synchronously in all geomorphic landscapes.

The timing of the T2 terrace fill deposition corresponds closely to the warm and dry late Holocene "Unstable Episode" (ca. 2700 to 1000 yr B.P.) in North Dakota (Bluemle and Clayton 1982:Figure 8). It is also virtually synchronous with a second episode of Holocene aggradation in the White River Badlands, dated at ca. 2500 to 850 yr B.P. (White and Hannus 1985). The earlier portion of the depositional event falls within the Sub-Atlantic climate episode (ca. 2890 - 1690 yr B.P.) of the Brynt-Sernander/Bryson et al. (1970) model. That the aggradation (in both the uplands and lowlands) occurred during a return to more drought-like conditions may be reflected in the low negative δ¹³C values associated with samples collected from 35 cm and 10 cm below the surface at Localities O and A (Figure 49). The δ¹³C value at Locality O (-19.5‰) is associated with bulk soil organics which yielded a radiocarbon age of 2160 ± 70 yr B.P. (Table 5, Figure 14).

The T2 period of fluvial aggradation was followed by stream incision which led to the creation of the T2 terrace. While the THRO δ¹³C data are again ambiguous, this incision was likely triggered by the change to wetter conditions associated with the onset of the Mandan Stable Episode (ca. 1000 - 800 yr B.P. to present) in North Dakota. The timing of this event is more or less synchronous with a period of maximum stability and soil formation in the Big Bend-Lake Sharpe region of central South Dakota, estimated by Toom (1992b) to have occurred between ca. 950 to 700 yr B.P. This stable period, which is pedogenically evident over a large portion of the Northern Plains, is correlated with the warm and moist Neo-Atlantic climatic episode (Toom 1992b). According to Gonzalez (1987:Table 9), temporally analogous periods of valley incision in the Great Plains region are recognized by Brice (1966), Leopold and Miller (1954), and Albanese and Wilson (1974).

It is appropriate to emphasize at this point that fluvial chronologies are absent, or are poorly documented, from many portions of the Northern Great Plains. Those that are available indicate a highly complex and variable geomorphic history, especially in terms of post-depositional sediment
preservation and/or modification. Albanese and Wilson (1974:17) and Albanese (1978a:389) explicitly argue that the correlation of Holocene terrace sequences in the Northwestern Plains beyond the scale of individual drainage basins is hampered, if not precluded, by widespread variation in hydrologic regimes, sedimentation rates, and sediment preservation. Similar difficulties are encountered in the correlation of terraces along different portions of the Missouri River in North Dakota (Artz 1992; Reiten 1983).

The one terrace that is more or less universally recognized is also the most recent. In Wyoming, this terrace fill is assigned to the Lightning Formation, broadly dated at between ca. 700 and 100 yr B.P. (Leopold and Miller 1954). Temporally similar alluvium is recognized throughout the Northwestern Plains (Albanese and Wilson 1974; Knox 1984a), and the Central Great Plains (Brice 1966; May 1986). Its equivalent in the South Unit study area is the T1 terrace fill, deposited between ca. 400 and 150 yr B.P. (see Chapter V). The timing of this aggradation is very similar to that of the Knife River B2 terrace fill in west-central North Dakota (Reiten 1983) and to the deposition of the MTO terrace fill of the Missouri River near Williston, North Dakota (Coogan 1983). Albanese and Wilson (1974:17) suggest that the widespread recognition and correlation of the Lightning Formation is made possible by its relatively recent age, which has reduced the cumulative effects of post-depositional modification.

The T1 period of aggradation was accompanied by widespread eolian and slopewash deposition throughout the study area; factors which suggest a brief return to more xeric conditions. The presence of very weakly developed Entisols on most older terrace surfaces is incongruous with the age of the terrace fill deposition and indicates more recent loess and/or slopewash veneers. A similar situation is evident in the uplands, where many youthful surface sediments barely qualify for classification as soils (i.e., they support plant growth). In some areas, such as Locality B, sand dune deposition appears to have continued until recent times. Likewise, the mapping and correlation of alluvial fans in major stream valleys indicates widespread interfingering of slopewash and T1 fluvial sediments (Chapter V: Appendix A).

Approximately 150 years ago, and definitely prior to the 1880's, the Little Missouri underwent a 1.5 to 3 m lowering of its base level; an event apparently not associated with significant bedrock incision. Everitt (1968) argues for a rearrangement of the channel and floodplain sediment during a catastrophic flood event, while a similar hypothesis involving channel in-filling and subsequent erosion was offered in the previous chapter. In any case, Gonzalez (1987:70) presents fairly convincing tree-ring evidence that Paddock Creek experienced an episode of incision between
ca. 200 and 115 yr B.P. This event created the T1 terrace and led to the formation of the T0 floodplain. Whether or not the lowering of the Little Missouri base level was associated with bedrock incision, the timing of the creation of T1 does correspond to the initiation of the modern cycle of stream erosion in the Northern Great Plains; an episode well dated at between 150 and 100 yr B.P. (Albanese and Wilson 1974:17). Since then, climatic conditions have varied within modern parameters, and streams appear to have remained relatively stable or have tended to erode slightly. The uplands have been relatively stable by badland standards, and have been dominated by C3 vegetation. Fairly significant, but brief, mesic and xeric periods have been historically documented throughout the last century. One of the more severe droughts occurred during the 1930's, and was associated with widespread eolian deposition and gully incision in western North Dakota (Bluemle 1991:16; Hamilton 1967). Slopewash deposition, the dominant form of sedimentation in the badlands, has continued throughout the study area, and numerous alluvial fans have encroached upon the T1 surface.

In light of the geomorphic and climatic data just presented, it is now possible to refine certain aspects of the Mandan Stable Episode in southwestern North Dakota. The episode is described as a period of generally cool and moist climate characterized by reduced slopewash and eolian deposition during the last ca. 800 years, preceded by approximately 200 years of warmer temperatures, decreased precipitation, and increased aggradation (Bluemle and Clayton 1982:13). While the proposed warm and dry interval from ca. 1000 to 800 yr B.P. corresponds to the latter portion of the T2 episode of valley aggradation (and is therefore supported by the THRO data), the Mandan Stable Episode does not recognize the subsequent T1 period of aggradation that is now well-documented in the South Unit study area from ca. 400 to 150 yr B.P.; a period which also witnessed widespread eolian and slopewash deposition. These geomorphic events, together with the evidence for area-wide valley incision approximately 150 years ago, suggest that the term "Mandan Stable Episode" in southwestern North Dakota be limited to the period from ca. 800 to 400 yr B.P., which corresponds to an interval of landscape stability and soil formation in the uplands and fluvial degradation followed by possible stability in the lowlands (Table 9). The unnamed period from ca. 400 to 150 yr represents the T1 episode of valley aggradation and synchronous loess and slopewash deposition, while the final unnamed episode, from ca. 150 yr B.P. to present, corresponds to an interval of downcutting responsible for the abandonment of the T1 surface and the subsequent establishment of the modern TO floodplain (Table 9).
The Little Missouri Badlands today remain a dynamic region, one where localized landscape stability, erosion, and deposition can occur simultaneously in response to myriad small-scale variations in geomorphic processes. The South Unit data suggest, however, that large-scale changes in geomorphic landscapes have been driven by changes in climate and its a priori relationship with vegetation and soils. That the badlands appear to correspond so well to fundamental process-response models of climate, vegetation, and sediment yield is not at all surprising, considering that data derived from badland areas were influential in the development of many traditional geomorphic models. These include models of drainage evolution (Leopold et al. 1966; Schumm 1956a; Strahler 1950), landscape textural variation (Smith 1950), slope evolution (Schumm 1956b), and equilibrium theory (Strahler 1950, 1956). This reliance on badlands in the development of early large-scale geomorphic models has been based on three major assumptions: (1) that badland processes are extremely rapid; (2) that landforms in badlands are analogous to major fluvial landforms; and (3) that badland processes are easily studied (Bryan and Yair 1982). While the South Unit data suggest that these assumptions are essentially valid at the large scale, geomorphological research indicates that badland processes are more complex than first thought (Bryan and Yair 1982; Campbell 1989). This complexity is the result of a high degree of variability in bedrock lithology, runoff generation, slope morphology, and erosion rates (Campbell 1982; Wise et al. 1982; Yair et al. 1980; Yair et al. 1982). As a result, geomorphologists have combined traditional short-term processual studies with long-term studies of badland development. The latter incorporate the analysis of regional trends, the interpretation of stratigraphic relationships, and the dating of lithostratigraphic units (cf. Wells and Gutierrez 1982). Just as the geomorphological study of badlands in general has become increasingly more complex, so should additional efforts at paleoenvironmental reconstruction in the Little Missouri Badlands. The reconstruction presented here has concentrated on large-scale and fairly obvious climatic/geomorphic relationships. Such is the nature of many initial scientific investigations. Future research, however, should strive for finer spatial and temporal resolution; a diminution in scale that will hopefully be mirrored by future archaeological investigations.
CHAPTER VII

CONCLUSIONS: THE ARCHAEOLOGICAL RECORD IN A STRATIGRAPHIC AND PALEOENVIRONMENTAL CONTEXT

Introduction: A Matter of Perspective

In the three previous chapters, individual summaries of the archaeology, stratigraphy, and paleoenvironment of the South Unit study area were presented. This rather compartmentalized approach in many ways mirrors earlier archaeological research efforts in the badlands, which have tended to separate archaeology from the natural environment, especially as the two are related diachronically. In most prior research reports, the archaeology has been individually summarized and little attention was paid to the natural environment other than to describe modern conditions in sections set aside as individual chapters or appendices. Furthermore, since the inception of large-scale archaeological research efforts in the late 1970's, the archaeological record in the badlands has been viewed primarily as though it were part of a static natural system (Kuehn 1993:319). As a result, archaeologists adopted the perspective, albeit inadvertently, that archaeological material remains have been little affected by changing geologic conditions (Kuehn 1993:314). With the procurement of significant new data on sedimentological, geomorphic, and paleoclimatic history, attention can now turn to integrating the cultural and natural components of the landscape into a more contextual framework. By doing so, the archaeology of the Little Missouri Badlands can be open to interpretation from a new perspective; one that views the archaeological record as an interactive part of a highly dynamic natural environment.

A Re-examination of the Archaeological Data Base

The dissected topography illustrated in Figure 50 is the product of numerous interactive geomorphic processes, including wind deflation (which stripped eolian sediments from most upland surfaces), hillslope erosion (which greatly reduced the size of upland landforms and resulted in the transportation of sediment to the lowlands), and alluvial degradation (which flushed most of the redeposited sediment from the lowland valleys). By now it should be clear that these geomorphic changes were widespread to the point that the extant archaeological record cannot be assumed to
wholly, or even significantly, reflect the entire range of prehistoric human behavior that occurred there. While relevant to archaeological research in many geomorphic and physiographic regions, in a region as dynamic as the Little Missouri Badlands this truism is a critical prerequisite to the drawing of meaningful archaeological inference (Butzer 1982; Schiffer 1987; Schoenwetter 1981; Stein and Farrand 1985; Waters 1992). Simply recognizing that the archaeological record may not be complete, however, is not sufficient. What is needed is a more specific identification of those non-cultural processes that were active in creating the record as it is now available to scientific study. In the THRO South Unit, these processes are best elucidated by re-examining key elements of the archaeological data base in light of the newly presented stratigraphic, geomorphic, and paleoenvironmental data. While doing so, previous explanations as to the causal factors behind widely recognized patterns in archaeological site location and age are also presented. It is hoped that by reiterating earlier explanations, the usefulness of a more contextual approach will become self-evident.

Site Age, Cultural Affiliation, and Natural Setting

These components of the archaeological record have long served as building blocks for the development of basic theories on prehistoric human behavior and diachronic culture change. In the
Little Missouri Badlands, as elsewhere, the first two topics formed the basis for the building of a regional cultural chronology (Beckes and Keyser 1983; Lau 1981; Loendorf et al. 1982), while the third was used to develop hypotheses concerning prehistoric settlement patterns (Beckes and Keyser 1983; Campbell et al. 1983; Hill 1988; Metcalf 1984). If formulated on the basis of the extant archaeological record, are these chronologies and settlement hypotheses valid? It was noted in Chapter IV that all of the major cultural traditions and periods common to the Northwestern Plains are to some degree represented in the 178 prehistoric sites located during the course of the THRO South Unit investigations. In this sense, the cultural chronology itself does appear valid, perhaps because it mirrors larger inter-regional chronologies (cf. Frison 1978; Gregg 1985a). It was also noted however, that the temporal distribution of various cultural components within the overall site aggregate is decidedly unequal. The same applies to their natural setting in terms of landform and sedimentary depositional environments. It is these disparities in the age and setting of sites, and the hypotheses developed to explain them, not the basic chronology or overall distributional pattern, that are open to re-interpretation.

As summarized in Chapter IV, components associated with named archaeological units in the South Unit study area include: the Hell Gap-Agate Basin complex (n = 1), Logan Creek-Mummy Cave complex (n = 1), McKean Complex (n = 4), Pelican Lake complex (n = 5), Late Archaic period (n = 2), Besant complex (n = 4), Plains Village tradition (n = 4), and Late Prehistoric period (n = 8). More useful than the number of named unit terms are the time periods represented by the various components. Viewed from this perspective, the temporal composition of the South Unit archaeological data base is as follows (based on the percentage of components that have been relatively or numerically dated):

<table>
<thead>
<tr>
<th>Time Period</th>
<th>Components</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>10,500 - 9500 yr B.P.</td>
<td>n=1</td>
<td>3.4%</td>
</tr>
<tr>
<td>7500 - 5250 yr B.P.</td>
<td>n=1</td>
<td>3.4%</td>
</tr>
<tr>
<td>4500 - 2600 yr B.P.</td>
<td>n=4</td>
<td>13.8%</td>
</tr>
<tr>
<td>3500 - 1700 yr B.P.</td>
<td>n=5</td>
<td>17.2%</td>
</tr>
<tr>
<td>3500 - 2000 yr B.P.</td>
<td>n=2</td>
<td>6.9%</td>
</tr>
<tr>
<td>2050 - 1200 yr B.P.</td>
<td>n=4</td>
<td>13.8%</td>
</tr>
<tr>
<td>2000 - 250 yr B.P.</td>
<td>n=8</td>
<td>27.6%</td>
</tr>
<tr>
<td>950 - 150 yr B.P.</td>
<td>n=4</td>
<td>13.8%</td>
</tr>
</tbody>
</table>

These ages clearly indicate an increase in the number of temporally identified components through time. Only 6.8% of the components are older than ca. 4500 yr B.P., while the majority
(55.2%) are younger than ca. 2050 years. There are two obvious explanations for such a temporal distribution. The first is that the archaeological record has been influenced by factors of site preservation and/or visibility. The older the material remains, the more likely they are to have been destroyed by erosion or obscured from view by sediment deposition (Kuehn 1993:330). The second is that the pattern reflects an increase in the size of prehistoric populations through time. As we will see, the chronostratigraphic data show a strong correlation between the temporal distribution of sites and the temporal distribution of late Quaternary sediments. This correlation favors the first explanation. The same stratigraphic data also suggest that prehistoric population densities in the badlands region are likely to be difficult to quantify because of the incomplete and highly disparate nature of the preserved late Quaternary landscape.

Archaeologists working in the badlands recognized very quickly that there was an absence, or extreme paucity, of cultural materials associated with the Paleoindian and Early Plains Archaic traditions (Kuehn 1993). In spite of this fact, some researchers either ignored, or made only cursory reference to, natural site formation processes as a causal factor. Loendorf et al. (1982:46-48), for instance, opted for the single explanation that the badlands received little or no late Pleistocene and early Holocene cultural utilization because of deterioration of local resource potential resulting from worsening post-glacial climatic conditions and associated landscape instability. Beckes and Keyser (1983:173) attributed the temporal pattern to either erosional influences on the archaeological record, or to cultural preference in favor of larger river valleys like the Missouri. Similarly, Simon and Keim (1983:25) hypothesized that the badlands were avoided by earlier groups because the region was made largely uninhabitable by extensive downcutting, or that a large portion of the landscape may have been eroded by erosional processes.

Support for the hypothesis that natural processes have played a major role in the lack of late Pleistocene and early Holocene-aged cultural materials is dependent upon two conditions. First, it must be demonstrated that groups associated with the Paleoindian and Early Archaic traditions did in fact utilize the badlands region. Second, there must be sufficient geomorphic and stratigraphic evidence to document the widespread erosion or burial of late Pleistocene and early Holocene land surfaces. The first criteria was briefly addressed in Chapter II, with a summation of Paleoindian projectile points recovered from the badlands, including the Agate Basin specimen collected from a site in the South Unit. Additionally, Simon and Keim (1983:25) discussed examples of both fluted and parallel-oblique flaked projectile points recovered from localities near the badlands. They also made the highly relevant observation that: "it is puzzling that these distinctive diagnostics have been
found in areas adjacent to, and surrounding, the Little Missouri drainage (north, south, east, and west) and yet none have been reported within the breaks themselves" (Simon and Keim 1983:25). While a number of projectile points were recovered from the badlands since then, Simon and Keim's observation highlights the fact that greater numbers of Paleoindian artifacts are located in adjacent areas. Beckes and Keyser (1983:173) documented the recovery of isolated, late Paleoindian (parallel-oblique flaked) projectile points from the Little Missouri National Grasslands, and also reported on the identification of fluted projectile points from nearby localities. They argued that the artifacts "...demonstrate(s) the presence of Paleoindian hunters on the Grasslands" (Beckes and Keyser 1983:173). The Little Missouri National Grasslands are administered by the U.S. Forest Service and encompass a large portion of the badlands region (Figure 1). As these examples illustrate, there is sufficient projectile point evidence to document the presence of Paleoindian groups in this area. The data also suggest, however, that the frequency of Paleoindian materials is greater in peripheral (i.e., non-badland) areas. Likewise, the vast majority of projectile points recovered from the badlands are associated with later Paleoindian groups, while fluted (i.e., Clovis and Folsom) specimens are totally absent. Again, the oldest artifacts are the least common. As for Early Plains Archaic materials, it was mentioned in Chapter II that a number of Logan Creek/Mummy Cave complex sites were reported from western North Dakota, and that at least one such component was identified within the South Unit study area. Beckes and Keyser (1983:176) stated that occasional large side-notched projectile points of the Hawken (i.e., Simonsen) variety, ca. 7500 to 5250 yr B.P., were identified in the badlands and nearby areas. Most of the specimens were recovered from surface contexts, although two were associated with subsurface components, (Kuehn et al. 1987; Simon and Keim 1983). More common, but still quite rare, are projectile points associated with the Oxbow complex (ca. 5500 - 3500 yr B.P.). These distinctive bifurcated points are reported from subsurface excavations in peripheral areas (Beckes and Keyser 1983), and from ridgetop sites in the badlands (East et al. 1985; Borchert and Wermers 1994). In the few cases where Oxbow and Simonsen projectiles points are associated with radiocarbon ages, the components date from ca. 5300 to 5000 yr B.P. (Kuehn 1990:120).

Like Paleoindian, the recovery of Simonsen and Oxbow projectile points from sites in the badlands does indicate that Early Plains Archaic groups were present in the region. Other material remains associated with their occupation, however, are extremely rare. The key to understanding this paucity may lie in Waters' (1992:97) summation of a basic geoarchaeological principle: "the nature and completeness of the buried archaeological record parallels the nature and completeness
of the late Quaternary stratigraphic sequence." It was demonstrated in Chapters V and VI that much of the late Quaternary stratigraphic sequence in the South Unit study area has been obliterated or buried by natural processes of erosion and deposition. Furthermore, the temporal composition of the extant sedimentological record closely mirrors that of the extant archaeological record. Recall that in the uplands, sediments older than the Riverdale Member (ca. 5000 - 4500 yr B.P.) were only encountered at eight stratigraphic sections, and these are all limited to small basins and leeward areas where sediment was trapped and preserved (Figures 4 and 23). The oldest deposits, dating from ca. 12,000 to 9000 yr B.P., were only identified at five localities (Figure 23). Of the ca. 24,000 linear meters of upland edges examined during the course of the South Unit investigations, the visible horizontal extent of all identified Aggie Brown and Pick City Member sediment totals less than 550 meters, or 2.3% of the total length of the surveyed escarpments. This figure, of course, does not take into account deposits that may be located away from the eroding edges and therefore not recognizable by surface reconnaissance. It does, however, illustrate the paucity of visible Pick City and Aggie Brown-aged sediments in the study area. In addition, stratigraphic investigations undertaken in conjunction with numerous archaeological excavations in upland settings have invariably failed to encounter sediments older than ca. 5300 yr B.P. (Artz 1992; Jorstad et al. 1986; Kuehn 1990). This characteristic of the stratigraphic sequence led Wyckoff (1982) to suggest that the uplands were subjected to widespread wind deflation during the Altithermal. Although overgeneralized, this hypothesis does tend to be supported by the presence of truncated soils at the top of middle Holocene units at Localities A, O, K, and L (Figure 23). At Locality A, at least three individual post-pedogenic erosional events are evident in the middle Holocene portion of the section. At all four localities, every middle Holocene-aged unit is bounded by an upper erosional unconformity. This magnitude of eolian erosion is not evident in any of the other Oahe Formation members (i.e., Aggie Brown or Riverdale). Archaeological materials located in such eroded areas would have remained behind as deflation lag, to be re-buried by post-middle Holocene deposition, and perhaps mixed with later cultural occupations. Furthermore, site locational evidence suggests that Middle Archaic through Late Prehistoric groups may have concentrated their upland activities along the edges of ridges and buttes, rather than in the interiors, in order to take advantage of superior views (Kuehn 1990:192). If this were true of Paleoindian and Early Archaic groups as well, the affects of hillslope erosion on the archaeological record must also be taken into account. In their study of hillslopes in the THRO South Unit, Clayton and Tinker (1971) documented an average rate of slope lowering by slopewash of 0.41 inch per year on west-facing Sentinel Butte slopes, 0.14 inch
per year on southwest-facing Bullion Creek slopes, and 0.11 inch per year on northeast facing Bullion Creek slopes. It cannot be assumed that these rates were constant throughout the late Quaternary; indeed they may have been even far greater during the more severe periods of middle Holocene aridity. As a result, hillslope erosion must be considered a significant geomorphic process and one that certainly qualifies as a factor in archaeological site formation.

Given the evidence for widespread deflation during the middle Holocene, and the ongoing effects of hillslope lowering, the potential for intact Paleoindian and Early Plains Archaic components in the uplands is possible but must be considered low. The potential for similar components in the alluvial lowlands is far less. The oldest sediments encountered to date in a lowland setting are those associated with the T4 terrace fill, deposited during an apparent episode of Middle Holocene aridity between ca. 7000 and 6000 yr B.P. This factor alone argues against the possibility of any Paleoindian sites remaining in the lowlands. Furthermore, the poor preservation and limited distribution of the T4 sediments suggests that Early Plains Archaic sites, especially those associated with the Logan Creek/Mummy Cave complex, are likely to be extremely rare. To illustrate, T4 terrace fill sediments were only identified within the Knutson Creek and Paddock Creek drainage basins, and these are comprised of only 10 small localities. The remnants, whose locations are illustrated in Appendix A, have an estimated combined sediment volume of less than 45,000 cubic meters. Remember, the bottom portion of the terrace fill is comprised of lateral accretion sediments deposited under high energy subaqueous conditions that are not conducive to the preservation of archaeological sites in primary context (cf. Waters 1992:126,138). This limits the potential for intact Logan Creek-Mummy Cave materials to the T4 vertical accretion sediments and to the surface of T4. As far as the latter is concerned, sites that may be located on the terrace surface face the possibility of a loss of integrity through mixing with later cultural components (Waters 1992:93) and because of the widespread occurrence of post-occupational slopewash deposition. In addition, they could be obscured from view by quite recent eolian terrace veneers. There is also the possibility that sites from this time period could be located in or on top of alluvial fans deposited more or less synchronously with the T4 fill, however only one such fan was identified in the South Unit (Knutson Fan 93F). The generally coarse-grained and poorly sorted lithology of the fan sediments also appear unfavorable to site preservation.

The extremely limited distribution of T4, coupled with the available paleoclimatic and geomorphic evidence, suggests that most of the T4 fill was eroded during one or more episodes of fluvial degradation that occurred subsequent to the T4 period of aggradation. These could have
included periods of incision believed to have taken place between ca. 6000 - 5500 yr B.P. and 4500 - 2700 yr B.P. (see Chapter VI). Similar stream erosion may have been responsible for the removal of even older fluvial sediment (i.e., late Pleistocene and early Holocene) although there is no evidence to support this possibility in the study area.

Whether we are talking about upland eolian settings or lowland alluvial settings, the South Unit stratigraphic data do indicate a general paucity of sediment older than ca. 5500 - 4500 years. The processes responsible for this sediment loss are wind deflation and hillslope erosion in the uplands, and fluvial degradation (i.e., downcutting and lateral stream migration) in the lowlands. Similar sedimentological voids in other portions of North America have been identified. For instance, in Wyoming and Colorado, Albanese (1977) argues that there is extreme variability in the preservation of fluvial sediment older than ca. 7500 yr B.P. In the Upper Midwest, Thompson and Bettis (1982) recognize a significant void in the alluvial record of small stream valleys from ca. 8000 to 3500 yr B.P. In the Southern High Plains, Blum et al. (1992) present evidence for the widespread erosion of sediment from alluvial landscapes during the late Pleistocene through the middle Holocene, while in the arid Southwest, Waters (1988) demonstrates that virtually all alluvial sediment older than 5500 yr B.P. has been flushed from the Santa Cruz River valley in Arizona. In North Dakota, Coogan (1983) found that early Holocene alluvium was lacking from the Missouri River valley near Williston, while a similar situation was encountered at the mouth of the Knife River by Reiten (1983). These additional examples of sediment loss indicate that the Little Missouri Badlands are far from unique in this regard. They also reinforce the need to articulate the relationship between late Quaternary landscape evolution and archaeological site preservation.

If the widespread obliteration of late Pleistocene and early Holocene-aged sediment is a major factor behind the paucity of Paleoindian and Early Archaic components, what about the increased frequency of more recent archaeological materials? Again, the basic principle that the completeness of the archaeological record is dependent upon the completeness of the stratigraphic record appears relevant. For examples, we begin first with the middle to late Holocene McKean complex.

From a low of 3.5% for the earlier groups, the percentage of identified McKean components (ca. 4500 - 2600 yr B.P.) in the study area increases to 13.8%. A similar increase is evident in the badlands as a whole. Compared to a complete absence of definitive Paleoindian components, and only a handful of components associated with the Early Plains Archaic (i.e., Logan Creek/Mummy Cave and Oxbow complexes), the number of McKean components identified in the badlands region
(as of 1990) totals at least 40 (Hill 1988; Kuehn 1990). This dramatic increase led earlier researchers to hypothesize that McKean represented "the earliest intensive use" of the Little Missouri area (Loendorf et al. 1982:51). Indeed, the widespread occurrence of McKean sites compared to those associated with earlier groups is a phenomenon that was recognized throughout the Northern Great Plains (Frison 1978; Gregg 1985a; Wormington and Forbis 1965). The traditional explanation for this increase is that improved climatic conditions after the amelioration of the "Altithermal drought" greatly enhanced the resource potential and overall carrying capacity of the Great Plains (cf. Gregg 1985a; Greiser 1985). Considering the evidence for multiple episodes of middle Holocene aridity, and the possibility that these episodes may have been associated with decreased bison populations, this certainly could have been the case, although data from the badlands region must still be considered inconclusive. Whether or not there was a dramatic increase in McKean utilization of the region, one thing is certain, McKean sites by and large post-date the last major episode of middle Holocene drought. This factor does have important sedimentological and archaeological implications.

It has already been suggested that periods of middle Holocene aridity were probably responsible for widespread wind deflation in the uplands. While this indicates that sediments older than ca. 4500 yr B.P. (the apparent end of the last middle Holocene drought in the study area) are likely to be rare, additional evidence indicates that the reverse is also true. That is, the most widespread sediments in the uplands appear to be those dating to the last 4500 years. At seven of the eight upland sections where Pick City Member sediments are identified, the stratigraphic evidence indicates that the most recent episode of middle Holocene erosion was followed by renewed loess deposition and widespread landscape stability. At one of these sections (Locality O), the soil topping the post-erosional unit was dated at ca. 4000 B.P. At the eighth section (Locality C) the Pick City Member unit is topped with a buried soil apparently equivalent in age to the lower Thompson Paleosol (Clayton et al. 1976). Similar Thompson-aged soils are common in loess deposits throughout the region (Jorstad et al. 1986; Kuehn 1990; Wyckoff 1982).

The argument that McKean-aged (i.e., post-Altithermal) sediments are common in the uplands is strengthened by the landform distribution of McKean components themselves. In the South Unit, all four recorded McKean sites are located on the tops of ridges in association with loess deposits. A similar pattern is evident in the distribution of McKean sites in the badlands as a whole. In a synthesis of all sites in the Little Missouri National Grasslands with known cultural and temporal affiliations (n = 160), Hill (1988) noted that fully 78% of McKean sites were located on
upland landforms (Kuehn 1993). This percentage is substantially higher than that of any other cultural group; an aspect of the badlands archaeological record that was generally attributed to factors of cultural preference or resource availability (Beckes and Keyser 1983:177-178; Kuehn 1993:318).

The total lack of McKean sites in the South Unit lowlands, and their scarcity in similar settings throughout the badlands, appears related to the overall age of fluvial and alluvial sediments. Within the study area, the potential for intact McKean components in the lowlands appears limited to the surfaces of the T4 and T3 terraces and synchronous alluvial fans. Remember, the estimates for the deposition of the T4 and T3 fills are ca. 7000 - 6000 and 5500 - 4500 yr B.P., respectively. Although the McKean complex has a general temporal range of ca. 5000 - 2600 yr B.P. in the Northern Plains and Rocky Mountain regions (Gregg 1985a:108-112), no sites older than ca. 4500 yr B.P. are documented in western North Dakota. In fact, sites older than that are limited primarily to the Colorado Front range and the Big Horn Basin area of Wyoming (Benedict and Olson 1973; Symns 1969, from Gregg 1985a:108). That being the case, McKean components in western North Dakota are likely to post-date the deposition of T3. The older T4 terrace has already been shown to have very limited spatial distribution. As illustrated in Appendix A, the T3 fill is somewhat more widespread, but still must be considered rare compared to the distribution of younger fluvial sediment. Like T4, the T3 terrace is primarily limited to the Knutson and Paddock Creek drainage basins, although it is also extant along Boicourt Wash, and there are a few small remnants along Sheep Creek and Petrified Forest Creek (Appendix A). Like earlier components, McKean sites located on the T4 and T3 surfaces also face the possibility of artifact mixing, disturbance by slopewash deposition, and burial by terrace veneers.

The temporal parameters of the McKean complex appear to correspond to a return to more mesic climatic conditions. As argued, these conditions resulted in substantial periods of landscape stability in the uplands. In the lowlands, McKean appears to post-date the deposition of the T3 terrace fill, but was synchronous with widespread fluvial degradation associated with reduced sediment yield from adjacent hillslopes. This reduction in sediment, along with greater stream discharge, resulted in the removal of large amounts of earlier alluvial fill. Along with the removal of sediment went much of the potential for intact McKean sites.

The trend toward increasing site frequencies through time continues with the Late Plains Archaic, ca. 3500 - 1700 yr B.P. Late Archaic components account for 24.2% of the total South Unit assemblage, as compared to 13.8% for McKean. The former include five components assigned to
the Pelican Lake complex and two to the general Late Archaic temporal period. Late Archaic sites are evenly distributed between uplands and intermediate landforms (42.9% each), although they are the oldest cultural group in the study area to be associated with a site in the lowlands (14.2%). Hill (1988) noted a more significant change in the badlands as a whole, where the frequency of sites in the uplands decreased from 78% for McKean to 60% for Pelican Lake. Likewise, the number of lowland sites increased from 18% for McKean to 23% for Pelican Lake. These differences prompted Hill (1988:16) to suggest that Pelican Lake groups adopted a different settlement/subsistence strategy than McKean in response to the increased aridity of the Sub-Atlantic climatic episode. As far as the upland stratigraphic record is concerned, Riverdale Member sediments equivalent in age to the Late Archaic are documented in the South Unit and elsewhere. At Locality O, a buried soil in Unit IIb was radiocarbon dated at ca. 2100 yr B.P., suggesting that it correlates with the more recent of the lower Thompson Paleosols (Clayton et al. 1976). Synchronous units and soils are evident at other upland localities in the study area, including Locality U (Figure 23). Outside of THRO, Jorstad et al. (1986) report on the presence of two radiocarbon-dated paleosols from the Late Archaic period on Cinammon Creek Ridge, while other Late Archaic dates from upland localities are reported by Simon and Borchert (1981a), Simon et al. (1982), and Floodman et al. (1983), among others. The uplands during the Late Archaic were therefore characterized by episodes of landscape stability and loess deposition; some of the latter apparently significant judging from the thickness of deposits such as Unit IIb at Locality O, and Unit IIb at Locality U (Figure 23).

The deposition of relatively thick loess units is indicative of a return to somewhat more arid conditions, and indeed the last 1000 years of the Late Archaic period correspond to Bluemle and Clayton's (1982) "Unstable Episode". It was during this period that fluvial aggradation in the study area resumed with the deposition of the T2 terrace fill (ca. 2700 to 800 yr B.P.). The deposition was not continuous, however, as the fill contains several buried soils. In any case, the fact that the last 1000 years of the Late Archaic period are synchronous with the aggradation of T2 suggests that the terrace fill does have the potential to contain Late Archaic components in primary context. The widespread distribution of T2 in the study area could explain the increase noted by Hill (1988) in the number of Pelican Lake sites in lowland settings. T2 sediments are extremely common along Knutson and Paddock Creeks, and are not infrequent along Sheep Creek, Jules Creek, Boicourt Wash, and Petrified Forest Creek (Appendix A). In addition, they are the only Holocene fluvial sediments older than T1 and T0 to be located in the Little Missouri Valley, although only two small remnants have thus far been identified. Late Plains Archaic materials therefore can be expected to
occur in Riverdale Member eolian units throughout the uplands, on the T4 and T3 terrace surfaces, in or on top of synchronous alluvial fans, in association with slopewash deposits on intermediate landforms such as foothills and slopes, and finally, in the fill of the T2 terrace.

The Besant complex of the Plains Woodland tradition, ca. 2050 - 1200 yr B.P., accounts for 13.8% of the South Unit components (n = 4). Rather surprisingly, all of these are associated with loess deposits in the uplands. In the North Unit of THRO, however, an additional 7 Besant sites were located, two of which are in the fluvial lowlands and five in intermediate settings (i.e., slopewash covered foothills). In the badlands region, total Besant components number around 40 and their landform distribution shows a dramatic decrease in upland settings. Compared to 60% for Pelican Lake, the number of Besant sites in the uplands drops to 38%. Conversely, the number of components in lowland (i.e., fluvial) settings increases to 27% (Hill 1988). Again, this pattern was attributed to a change in settlement strategy, with Besant groups adopting "a river orientation" (Hill 1988:18).

Geomorphic conditions in the uplands appear to have changed little during Besant times (as compared to Pelican Lake), with a continuation of loess deposition and one or more episodes of landscape stability. A buried Mollic epipedon at Locality B produced a radiocarbon age of 1480 ± 80 yr B.P., indicating temporal equivalency with the oldest of the upper Thompson Paleosols (Clayton et al. 1976). Synchronous soils appear to be present at Localities O, U, and Z, and may be extant in the near surface units at many other upland sections (Figure 23). Locality Z, however, was the only other section associated with a temporally equivalent Mollisol.

The entire Besant temporal period corresponds to the continuing aggradation of the T2 terrace fill and synchronous alluvial fans. This suggests that hillslopes remained poorly vegetated and sediment yield to the valleys continued to outpace degradation. Just as in Pelican Lake times, however, Besant-aged paleosols in the lowlands suggest that there were some episodes of at least localized landscape stability during this period. One such soil at Locality P yielded a radiocarbon age of 1210 ± 60 yr B.P. Terrace veneers were also deposited at this time, as indicated by a dated fire hearth excavated from a loess unit capping the T3 terrace in Paddock Creek (Locality N).

Areas with the potential to contain Besant components include Riverdale Member eolian units in the uplands, the surface of the T4 and T3 terraces and synchronous alluvial fans, intermediate landform settings such as slopewash-covered foothills, the T2 terrace fill, and contemporaneous slopewash deposits. It is clear from this relatively long list that the number (and/or volume) of geomorphic units with the potential to contain intact archaeological materials increases as the age of the cultural
components decrease. This appears at least partially related to the reduction in the amount of time that the archaeological materials have been exposed to natural site formation processes. Although this principle is more or less universal (cf. Blum et al. 1992; Butzer 1982; Schiffer 1987; Waters 1992), in the Little Missouri Badlands the relationship between time and site preservation appears quite strong, and could explain the increase in the number of sites younger than ca. 2050 years B.P. While demographic explanations (i.e., population changes) may contribute to the above described patterns, it has already been argued that the incomplete nature of the stratigraphic record is such that meaningful population estimates in the badlands region may be difficult, if not impossible, to formulate.

With Besant already discussed, we turn now to the last remaining cultural components; those assigned to the Late Prehistoric period (ca. 2000 - 250 yr B.P.) and the Plains Village tradition (ca. 950 - 150 yr B.P.). Together these account for 41.4% of all temporally identified components in the study area. When Besant sites are added, the percentage increases to 55.2%. This indicates that over 50% of the identified sites are concentrated in 1/5 of the known temporal range of human occupation in the region.

Late Prehistoric components in the South Unit are the most diverse in terms of landform setting, with 50% situated in the loess mantled uplands, 12.5% on slopes, 25% in fluvial lowlands, and 12.5% in slopewash lowlands. Of the four Plains Village components, 50% are located in fluvial terraces, 25% in alluvial fans, and 25% in intermediate slump deposits. These percentages represent a significant increase in the number of lowland sites through time (75% for Plains Village and 37.5% for Late Prehistoric, as compared to 14.2% for Late Archaic and 0% for Besant, McKeans, Early Archaic, and Paleoindian). A similar pattern at the regional level has long been recognized by archaeologists, who attributed it to cultural preference, or settlement pattern changes (Beckes and Keyser 1983:203; Hill 1988:18-25, from Kuehn 1993:318).

As far as geomorphic and climatic conditions during Late Prehistoric and Plains Village times, the first 1200 years of the Late Prehistoric correspond with continuing aggradation of the T2 terrace fill. Slopewash deposition also continued during this period, including the aggradation of the Beef Corral valley fill, which yielded a radiocarbon age of 930 ± 90 yr B.P. (Beta-32163). With the advent of more mesic conditions around 800 yr B.P. (i.e., the onset of the Mandan Stable Episode), fluvial and slopewash aggradation ceased and valleys underwent a period of incision and sediment degradation. This led to the creation of the T2 terrace.
The uplands, meanwhile, continued to witness loess deposition concurrent with the T2 period of aggradation. The uppermost eolian units at every study section are Late Prehistoric in age, as indicated by stratigraphic position, Oahe Member correlations, and radiocarbon ages. The widespread distribution of Late Prehistoric-aged loess is also evident on the basis of archaeological data. Numerous diagnostic Plains side-notched arrow points (see Chapter IV) have been recovered from eolian sediments both within and outside of the study area, and a fair number of Late Prehistoric components have been identified and radiocarbon dated during the course of archaeological excavations on prominent ridge systems (Aivazian 1981; Floodman et al. 1983; Jorstad et al. 1986; Kuehn 1986; Simon and Keim 1983).

By approximately 400 yr B.P., climatic conditions became more xeric and sediment yield to the lowlands increased once again. This resulted in the aggradation of the widespread T1 terrace fill and contemporaneous alluvial fans. This episode corresponds to the Plains Village tradition and to the latter portions of the Late Prehistoric period. The predominance of components associated with these groups in lowland settings, particularly Plains Village, may be a direct reflection of the predominance of the T1 fill. These recent sediments are ubiquitous in the major perennial and intermittent stream valleys such as Paddock and Knutson Creeks, and are virtually the only terrace deposits remaining in the smaller ephemeral valleys such as Jones and Jules Creeks, and in the valley of the Little Missouri River (Appendix A). Buried soils appear to be largely absent from the T1 fill, a factor apparently related to its young age. During the T1 period of aggradation, the uplands may have experienced a minor increase in loess deposition, as evident by the presence of shallowly buried Plains Village and Late Prehistoric sites, and by the presence of thin surface units at some localities (i.e., Locality S).

In terms of Late Prehistoric and Plains Village site potential, components associated with these groups are likely to be located in the mid to upper units of the Riverdale Member eolian lithofacies, on the surface of upland landforms, in intermediate landforms such as foothills and slopes (in association with slopewash or Paleocene bedrock), on the surfaces of the T4, T3, and T2 terraces and synchronous alluvial fans, in the fill of the T2 and T1 terraces, and in the fill of temporally equivalent fans. Late Prehistoric materials have the potential to be more common in the T2 terrace fill than Plains Village, but the available archaeological evidence suggests that Plains Village is more common in the T1 fill. The latter may be related to the fact that the Little Missouri Badlands are ethnographically known to have been part of the secondary or tertiary territories of the Mandan and Hidatsa (Bowers 1948; Wilson 1928), who could have controlled access to the region.
Finally, data from outside of the study area indicate that there is also the potential for extremely recent Plains Village and Late Prehistoric sites in relatively unstable settings like hillslopes and toe slopes. Sites of this nature include wooden conical timbered lodges and bison jumps.

The preceding discussion illustrates that the site formation processes most responsible for the destruction of earlier (i.e., Paleoindian, Early to Middle Archaic) cultural components are fluvial degradation (both incision and lateral migration), wind deflation, and hillslope erosion. Remember, however, that a limited number of upland settings do have the potential for early cultural materials. Lateral migration appears to have been virtually ongoing in all meandering streams, while episodes of downcutting, which were responsible for the flushing of sediments from lowland valleys, are estimated to have occurred between ca. 6000 - 5500 yr B.P., 4500 - 2700 yr B.P., 800 - 400 yr B.P., and ca. post-150 yr B.P. Wind deflation was ongoing in a localized sense, but was apparently widespread during periods of middle Holocene drought (i.e., from 7000 to 6000 yr B.P. and 5500 to 4500 yr B.P.?). Hillslope erosion has also been more or less continuous, but the more severe episodes were probably contemporaneous with drought periods as well. Processes most responsible for the burial of sites (thus obscuring them from view) include colluvial loess deposition in the uplands (and on the surfaces of terraces and fans), meandering stream aggradation, and slopewash deposition.

Site Type and Prehistoric Settlement

Now that significant temporal and spatial voids in the archaeological record have been identified, and previous hypotheses formulated to explain them have been reinterpreted, we turn our attention to those research topics more relevant to the study of prehistoric human behavior.

Near the beginning of Chapter IV it was argued that the analysis of site type, via Binford's (1980) model of hunter-gatherer settlement, is an effective approach to understanding aspects of intra-site activity and prehistoric settlement strategy. It was also demonstrated, however, that inferences relevant to these topics cannot be drawn from a simple examination of the badlands archaeological record (because the record itself is far from complete). It is therefore necessary to first determine the relationship between natural site formation processes and site type, and then consider the implications of this relationship to the future study of prehistoric settlement and subsistence in the badlands region.

Recall that all four of Binford's (1980) hunter-gatherer site type categories were identified in the badlands, but only three were located in the South Unit study area. Of these, 56.2% are classified as field camps, 43.3% as locations, and 0.5% as stations (Table 3).
Locations, or resource procurement areas, are overwhelmingly lithic raw material sites associated with outcrops of fluvial gravels. These are decidedly concentrated in the uplands on surfaces representing Miocene/Pliocene and Pleistocene river levels. No temporally or culturally diagnostic artifacts have yet been recovered from the sites, although their presence in the uplands suggests they could have been utilized since Paleoindian times.

A curious and potentially significant aspect of the South Unit locations is the paucity of components associated with floral and faunal procurement. Only two such sites were identified, a bison kill site and a possible eagle trapping pit. This paucity is indeed puzzling given the fact that the badlands contain abundant natural resources within a variety of ecosystems. As mentioned in Chapter II, these different ecosystems, which include riparian woodlands, hardwood draws, and upland grasslands, are the result of the badlands high topographic relief, which is dramatic when compared to the relatively featureless terrain of the surrounding Missouri Plateau. Ethnographic evidence indicates that Plains Village groups living along the Missouri River in west-central North Dakota frequently ventured into the badlands on hunting forays because of the abundance of game (Bowers 1948; Wilson 1928). Small groups (i.e., nuclear family and/or husband and wife teams) occasionally spent entire winters there, as the badlands are also extremely well sheltered (Bowers 1948). In historic times, the badlands were considered some of the richest game hunting territory in the Northern Great Plains (Nelson 1946; Simon 1982:56). In addition to the abundance of game animals such as bison, deer, elk, and bighorn sheep, the badlands were also rich in floral resources such as chokecherry, wild plum, gooseberry, and buffaloberry, to name but a few. Why then is there a lack of evidence for floral and faunal procurement? As suggested in Chapter IV, one reason may be that target resources themselves are organic and are not conducive to preservation in an archaeological context. If processing implements such as grinding stones or scrapers were curated and carried away from the site of procurement, the only evidence remaining would be the rapidly deteriorating organic waste (i.e., pits, bone fragments, etc.). If maintenance-related flintknapping was also conducted at the location, flaking debris could very well be the only material remaining, in which case the site was likely classified as a field camp.

Factors of material preservation and artifact curation aside, the stratigraphic and geomorphic data also provide clues for a second explanation. Of all the target floral and faunal resources mentioned, only bison were likely to be found in abundance on the grass-covered buttes and ridge systems in the badlands. Floral resources were (and are still) concentrated in intermediate and lowland landform settings, particularly wooded slopes, hardwood draws (i.e., heavily vegetated
sloping valleys), and stream-side riparian areas. These sheltered locations and slopes were also the preferred habitat for deer, elk, and bighorn sheep. What do we now know about the geomorphic history of areas such as these? They are all dynamic landscapes subject to the degradational processes of hillslope erosion, gullying, piping, slumping, and downcutting. Likewise, while the frequent occurrence of bison in the South Unit uplands today is testament to the likelihood that they were abundant in similar settings prehistorically, it is unlikely that they were dispatched by hunting groups in such exposed areas. On the contrary, the evidence from throughout the Northwestern Plains suggests that bison were driven off cliffs and cutbanks, trapped in the headwall portions of gullies and arroyos, or mired in wet spring-fed muds (Albanese 1978; Frison 1978; Frison et al. 1976; Loendorf et al. 1982). These settings are also geomorphically unstable. It is therefore probably no coincidence that the only bison procurement site located in the study area is Plains Village in age and is situated in slump deposits at the toe of the Little Missouri Escarpment. Older sites in similar settings are not likely to have survived to become part of the extant archaeological record. For these reasons, the absence of floral and faunal resource procurement sites could very well be a reflection of the geomorphic setting of the localities where the procurement was conducted. In a region as dynamic as the badlands, such geomorphic settings are not conducive to archaeological site preservation. This possibility has implications for future research in that extant floral or faunal locations, if identified, are likely to be quite recent (i.e., late Holocene in age), with cultural affiliations dominated by Plains Village, Late Prehistoric, and Besant.

The largest category of site type in the study area, field camps, are also conducive to reexamination within a geomorphological context. Field camps are found in virtually every landform setting and major depositional environment in the study area. Since it has been demonstrated that the ages of these landforms and environments are highly disparate, so then are the ages of the field camps associated with them. Camp sites located in the eolian uplands have the potential to date back to Paleoindian times, but the stratigraphic data suggest that the majority are going to be younger than ca. 5500 - 4500 yr B.P. (i.e., younger than the last episodes of middle Holocene drought). In the alluvial lowlands, no field camps older than ca. 7000 yr B.P. are to be expected, and the vast majority are likely to date from the late Holocene (i.e., in association with Besant, Late Prehistoric, and Plains Village groups). This is a reflection of the temporal and spatial pattern of the extant sedimentological and archaeological records; a pattern that has two very important implications to the study of prehistoric settlement and subsistence in the badlands region. First, it limits archaeological knowledge about Paleoindian, Early Archaic, and McKean lifeways
to those activities conducted in upland settings. Consequently, archaeologists are not likely to gain an understanding of the nature and extent of field camp use by these groups in the lowlands. In light of the previous discussion on the distribution of floral and faunal resources, it also suggests that information about Paleoindian, Early Archaic, and McKean subsistence behavior is likely to be limited as well.

The second important implication concerns seasonal utilization of the region. It has been argued by previous researchers (i.e., East et al. 1985; Kuehn 1990), that field camps located in the high, exposed uplands were not likely to have been occupied during the winter months, given the normal severity of winters in the Northern Plains. Winter occupation, like that ethnographically documented for the Mandan and Hidatsa (Bowers 1948) was no doubt concentrated in the more sheltered lowlands. Therefore, if field camps associated with Paleoindian, Early Archaic, and McKean groups are stratigraphically limited to upland settings, their settlement and subsistence activities during the winter season are going to remain unknown. This is unfortunate considering that most prehistoric hunter-gatherer groups in the Northern Plains are believed to have followed a seasonal round (Buchner 1979; Frison 1978; Keyser and Davis 1984; Nicholson 1988). Again, another possible research deficiency resulting from natural formation processes.

Although not identified in the South Unit study area, residential bases are an important constituent of hunter-gatherer settlement (Binford 1980). At least three residential bases are identified in the THRO North Unit (Kuehn 1990). Two additional sites are identified a short distance south of Medora (Johnson 1983; Metcalf and Schweigert 1987), and a possible third was excavated in the extreme southern portion of the badlands (Toom 1992a). Of significance is the fact that all of the sites are late Holocene in age. Four are assigned to the Besant complex, (Kuehn 1990; Toom 1992a), one to the late Extended Confluent variant of the Plains Village tradition, ca. 300 - 240 yr B.P. (Metcalf and Schweigert 1987), and one to the Plains Village tradition, Scattered Village complex, ca. 600 - 300 yr B.P. (Johnson 1983). Equally significant is the fact that all of the residential bases are located within the valley of the Little Missouri River. Two of the Besant sites from the North Unit are situated in slopewash-covered foothills along the valley margin, while the third is situated in T2 terrace fill near the mouth of an intermittent stream (Kuehn 1990). The Besant site investigated in the southern badlands is associated with eolian sand dune deposits at the base of a prominent butte (Toom 1992a). Both of the Plains Village sites are situated in the fill of the Little Missouri T1 terrace, although one (the Connel Ranch site) is associated with slopewash rather than fluvial sediments (Johnson 1983; Metcalf and Schweigert 1987). All of these are appropriate
settings for residential bases in that base camp location appears to have been dependent upon proximity to certain critical resources such as water, shelter from the elements, and fuel (Binford and Binford 1969; Greiser 1985:30-46; Hanson 1983:1405; Jochim 1976). Upland locations, which tend to be flat, exposed, and grass-covered, do not generally offer these resources in sufficient quantities to make extended occupation practical. It is therefore hypothesized that residential bases in the badlands were concentrated in riverine settings (i.e., the Little Missouri River); an hypothesis supported by the location of the five residential bases thus far identified. Given what is known about the age of sediments in the Little Missouri valley, it is not surprising that these same sites are late Holocene in age. The oldest sediments yet identified along the Little Missouri are those associated with the T2 terrace fill, deposited between ca. 2700 - 800 yr B.P., while the vast majority are associated with the T1 fill, deposited between ca. 400 and 150 yr B.P. This limits the potential for base camps to Late Plains Archaic, Besant, Late Prehistoric, and Plains Village groups. The fact that no Late Archaic base camps have yet been identified may be related to the apparent paucity of extant T2 terrace sediment in the Little Missouri valley (only two small terrace remnants have yet been identified). Again, previous researchers have argued instead that this paucity is the result of settlement pattern change; in this case, that Late Archaic groups abandoned the use of base camps (Hill 1988:16). In addition to not accounting for natural site formation processes, this argument is also inconsistent with existing archaeological and ethnographic evidence which suggests that base camps were an important constituent in the settlement patterns of virtually all prehistoric hunter-gatherers in the plains region. Greiser (1985), for instance, presents both theoretical and archaeologically-based evidence for base-camp use since Late Glacial times in the Central High Plains. Keyser and Davis (1984) and Keyser (1985) likewise argue that the Middle Archaic McKean complex in the Northwestern Plains utilized base camps as part of a central based circulating settlement pattern which also included field camps and locations. Hanson (1983), following Binford's (1980) model in a conceptual sense, but modified for the North Dakota region on the basis of ethnographic evidence, argues that the same three basic site types were utilized by all hunter-gatherer groups in the region. He does, however, identify four different levels of group size and group coalescence in response to seasonal changes in resource availability. Consequently, there is no reason to believe that the use of base camps in the badlands region was limited only to Besant and Plains Village populations. The fact that these are the only cultural groups currently associated with residential bases is more than likely a reflection of the original location of these types of sites (in riverine settings), and the geomorphic history of such settings. Once again, the evidence suggests
that a large portion of the prehistoric archaeological record is absent; this time, most of the record concerning extended occupation prior to Besant times. The natural site formation processes responsible for this particular research deficiency are the multiple episodes of fluvial degradation that appear to have flushed virtually all early and middle Holocene-aged sediment from the Little Missouri valley. It is ironic that there is a higher degree of sediment preservation in the valleys of perennial tributary streams such as Paddock and Knutson Creeks, but these streams were probably not preferred locations for extended occupation.

Given this rather bleak view of the extant archaeological record, just what can be done, if anything, to improve archaeological understanding of prehistoric human behavior in the badlands region?

Recommendations: The Need for Continuing Contextual Research

The preceding discussion identified some of the more significant natural processes that produced the arrangement of artifacts and ecofacts we call the archaeological record. Although these materials are static in the sense that they are no longer part of a living cultural system, they are indeed part of an extremely dynamic natural system. If we as archaeologists are to develop sound inferences about prehistoric human behavior in the badlands on the basis of the extant material evidence, then we must continue to expand our understanding of this natural system. We must also "anchor" our inferences on those elements of the archaeological record that are already developed and founded on solid evidence (Binford 1982:131). In other words, we must continue to pursue an interdisciplinary, contextual approach in order to further define voids or deficiencies in the record, while at the same time, expand on existing archaeological inferences that have been proven sound. The latter include basic elements of the cultural chronology, hypotheses on McKean through Plains Village lithic technology and raw material utilization, inferences concerning the morphology and function of middle to late Holocene sites in upland settings, and conclusions concerning the late Holocene utilization of lowland settings (cf. Artz 1986; Beckes and Keyser 1983; East et al. 1985; Floodman et al. 1983; Kuehn 1990; Metcalf and Schweigert 1987; Simon and Keim 1983; Wermers et al. 1991).

Just as some previously formulated inferences have proven well founded, others are open to question and may need to be replaced. The more significant of these are arguments that early cultural groups completely avoided the badlands region, that Middle Archaic groups had a preference for upland landforms, and that more recent groups had a greater reliance on riverine
settings. As described in the following statement by Schiffer (1987:341), the building and replacement of hypotheses is a normal process in the development of archaeological thought:

A great many important archaeological inferences have been established prematurely on the basis of small numbers of sites and inadequate coverage of study areas. Such inferences are quite vulnerable to new discoveries and, consequently, are overturned and replaced at a prodigious rate [1987:341].

The first step in alleviating significant archaeological research deficiencies in a particular region is to recognize their existence. In the case of the Little Missouri Badlands, these deficiencies are the result of significant spatial and temporal gaps in the archaeological record. Now that these gaps, and the causal factors behind them, have been identified, it is fairly obvious how we as archaeologists can begin to rectify them. For instance, the data suggest that sites affiliated with the Paleoindian tradition are only likely to be found in the eolian (and in some cases lacustrine) facies of the Aggie Brown Member of the Oahe Formation, which are limited to isolated pockets of sediment in upland localities. These are the areas that should be searched if we want to find Paleoindian material. Fortunately for archaeologists, the Aggie Brown Member is easily recognizable by the presence of distinctive, dark-colored Leonard Paleosols which invariably lie near the bottom of the stratigraphic sequence. A similar situation has been documented in the nearby Yellowstone Badlands, where intact Paleoindian sites appear limited to a loess/slopeswash veneer on top of the Makoshika Bench and to soils associated with the Lewis Surface (Eckerle and Aaberg 1990). Like there, finding Paleoindian sites in the Little Missouri Badlands will involve some geoarchaeological detective work.

The search for Early Plains Archaic sites should be concentrated in the eolian facies of the Pick City Member of the Oahe Formation, which like the Aggie Brown, is limited to isolated sediment traps in upland settings. Again, the Pick City is easily recognizable on the basis of its generally light color, lack of organic material, and by the presence of carbonate-rich, and often truncated, soils. The search for Early Archaic could also include the fill of the T4 terrace, which unfortunately, is very poorly preserved and apparently only extant in the larger perennial tributaries of the Little Missouri River.
While a fair amount of archaeological information is already available on the Middle Archaic McKean complex, most of it has been recovered from sites associated with lower portions of the eolian facies of the Riverdale Member. These sediments are apparently widespread throughout the uplands. The deficiency therefore, lies in the paucity of McKean sites investigated in lowland settings, and hence in our knowledge about the role of the lowlands in McKean settlement and subsistence. Unfortunately, this deficiency may be difficult to alleviate, as McKean times were not synchronous with known episodes of fluvial aggradation. On the contrary, McKean appears contemporaneous with one or more periods of fluvial degradation associated with the return to mesic conditions subsequent to the last episode of middle Holocene drought. Nevertheless, the search for McKean components should include the surfaces of the T4 and T3 terraces and synchronous alluvial fans, which although quite rare in the lowlands, are often capped with loess veneers which could have served to seal the components and thereby reduce the potential for mixing with later cultural occupations.

Late Plains Archaic components are also well documented, again mostly from Riverdale Member sediments in the uplands, but also from an increasing variety of lowland settings. The latter include the fill of the widespread T2 terrace and in similarly aged fans. In spite of this expanded data base, existing archaeological knowledge can best be augmented by concentrating future research in alluvial settings because the available data on Late Archaic settlement and subsistence are still biased in favor of upland utilization.

The potential for archaeological understanding increases with the more recent cultural groups (i.e., Besant, Late Prehistoric, and Plains Village) because their sites are the least affected by natural formation processes. This is particularly true of overall settlement and subsistence behavior, which for the first time, includes substantive data on the utilization of riverine environments. These data include information on site type (particularly extended occupation), and subsistence strategy (particularly floral and faunal resource utilization). This increased potential is just beginning to manifest itself in the archaeological literature because most previous investigations have focused on Middle and Late Archaic components in upland ridge systems (Kuehn 1993:318-319). Fortunately, there has been a significant increase in the number of excavations conducted in lowland settings in recent years (Borchert and Wermers 1994; Metcalf and Schweigert 1987; Peterson and Foster 1991; Toom 1992a).

While research efforts such as these can help alleviate some of the current deficiencies, there will probably always remain significant gaps in available information: an unfortunate legacy of the
dynamic geomorphic forces which have acted to shape the badlands region. Some of the more significant data losses can be expected to include: (1) most information on Paleoindian and Early Plains Archaic occupations in the area; (2) a large portion of the archaeological record reflecting the nature and extent of Middle Plains Archaic utilization of lowland ecosystems; and (3) a basic understanding of general adaptive strategies involving floral and faunal resource procurement, seasonality, and extended occupation in the badlands prior to the late Holocene. While these aspects of the archaeological record do appear missing, it may be possible to mitigate some of the informational voids through the use of ethnographic and archaeological analogy drawn from data recovered from adjacent areas, or regions ecologically and topographically similar to the Little Missouri Badlands. The latter could include the White River Badlands of South Dakota (cf. Hannus 1990), or the Yellowstone Badlands of Montana (cf. Eckerle and Aaberg 1990). Continuing growth in the study of badlands prehistory requires that archaeologists draw upon as many available sources of information as possible.

Summary

The principal objective of this dissertation was to identify those natural site formation processes that were the most influential in shaping the archaeological record of the Little Missouri Badlands. It was hoped that by doing so, the resultant information would serve as a foundation for the difficult task of building sound archaeological inference out of temporally and spatially fragmented materials. The research strategy followed the contextual paradigm in archaeology (cf. Butzer 1982; Schoenwetter 1981; Stein and Farrand 1985; Waters 1992), by integrating various methodological and conceptual aspects of archaeology, stratigraphy, pedology, geomorphology, and paleoclimatology.

The research, concentrated primarily in the South Unit of Theodore Roosevelt National Park, indicates that fluvial degradation, wind deflation, and hillslope erosion are the natural processes most responsible for the destruction of archaeological materials. Various forms of sediment deposition, particularly loess, vertical accretion sediment, and slopewash, have also served to bury archaeological sites and thus obscure them from discovery by conventional surface inventory.

The elucidation of natural site formation processes involved the identification of principal late Quaternary depositional environments and a description of the lithologic, pedologic, and chronologic composition of their associated geomorphic units. The latter include the eolian facies
of the Oahe Formation, in addition to meandering stream, slopewash, and occasional lacustrine (i.e., pond) facies. The eolian and lacustrine deposits are concentrated in upland (i.e., ridgetop and buttetop) landforms, while the meandering stream deposits are extant in the larger valleys. Slopewash deposits are ubiquitous throughout the region but are most concentrated on hillslopes, foothills, the floors of preformed valleys, valley margins, and valley mouths (i.e., alluvial fans).

The accumulated lithostratigraphic, chronostratigraphic, and pedostratigraphic data formed the basis for a reconstruction of the late Quaternary paleoenvironmental and geomorphological history of the study area. The reconstruction was aided considerably by an analysis of the stable isotopic composition of organic carbon at two of the more temporally complete loess sections. While the isotopic data indicate a steady decrease in C3 biomass from the late Pleistocene through the middle Holocene, and then an increase in C3 biomass in the most recent portion of the late Holocene, a more precise picture of paleoenvironmental conditions emerged when the isotopic data were supplemented with the stratigraphic and pedologic evidence.

Finally, the results of this research have demonstrated the value of a contextual approach to archaeology by providing a more clear picture of the inexorable link between archaeological materials and the natural environment. Indeed, the potential usefulness of contextual archaeology may be at its highest in badland environments because of their inherent sensitivity to changes in climate, vegetation, and sediment yield. As such, similar interdisciplinary approaches are likely to prove indispensable to all archaeologists working in highly eroded landscapes. Lessons learned from the Little Missouri Badlands suggest that these efforts should precede, rather than follow, conventional archaeological investigations.
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Wolman, M. G., and L. B. Leopold

Wood, J. J.

Wood, W. R.

Wood, W. R., and D. L. Johnson

Wormington, H. M., and R. G. Forbis
Wright, H. E., Jr.


Wyckoff, J.

Yaalon, D. H.

Yair, A., P. Goldberg, and B. Brimer

Yair, A., R. B. Bryan, H. Lavee, and E. Adar
APPENDIX A

PLANIMETRIC MAPS OF LATE QUATERNARY ALLUVIAL SEDIMENTS,
SOUTH UNIT, THEODORE ROOSEVELT NATIONAL PARK
Block B

1, 2, 3, 4 Holocene Terraces
Alluvial Fan (valley mouth)
Slopewash Deposits
(valley margin, valley floor)

Scale
0 0.25 0.50 0.75 1.00 km

Boicourt Wash
1, 2, 3, 4 Holocene Terraces

- Alluvial Fan (valley mouth)
- Slopewash Deposits (valley margin, valley floor)

Scale:

0 0.25 0.50 0.75 1.00 km

Sheep Creek
Little Missouri River
APPENDIX B

U.S.G.S. 7.5 MINUTE TOPOGRAPHIC MAP OF
THE SOUTH UNIT, THEODORE ROOSEVELT NATIONAL PARK,
SHOWING THE LOCATION OF ALLUVIAL PLANIMETRIC MAPS
AND PRINCIPAL EOLIAN DEPOSITS
APPENDIX C

LOCALITY A-SUMMARY OF ELEMENTAL AND ISOTOPE ANALYSIS,
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VITA

David Duane Kuehn
1826 Briar Oaks Drive
Bryan, Texas 77802

Personal Background

Sex: Male
Birth Date: November 25, 1952
Place of Birth: Elgin, North Dakota
Marital Status: Married, no children
Citizenship: USA

Educational Background

B.A. Anthropology, University of North Dakota, 1974
M.A. Anthropology, Northern Arizona University, 1981
Ph.D. Anthropology, Texas A&M University, 1995

Professional Background

1994-1995 Senior Prehistoric Archaeologist, Center for Environmental Archaeology, College of
Liberal Arts, Texas A&M University.

1994-1995 Assistant Lecturer, Department of Anthropology, Texas A&M University.

1991-1993 Chief Archaeologist, Fort Buford Archaeological Project. State Historical Society of
North Dakota.

1981-1991 Associate Research Archaeologist, Department of Anthropology, University of North
Dakota.


1979 District Archaeologist-McKenzie District, USDA Forest Service, Custer National
Forest.
