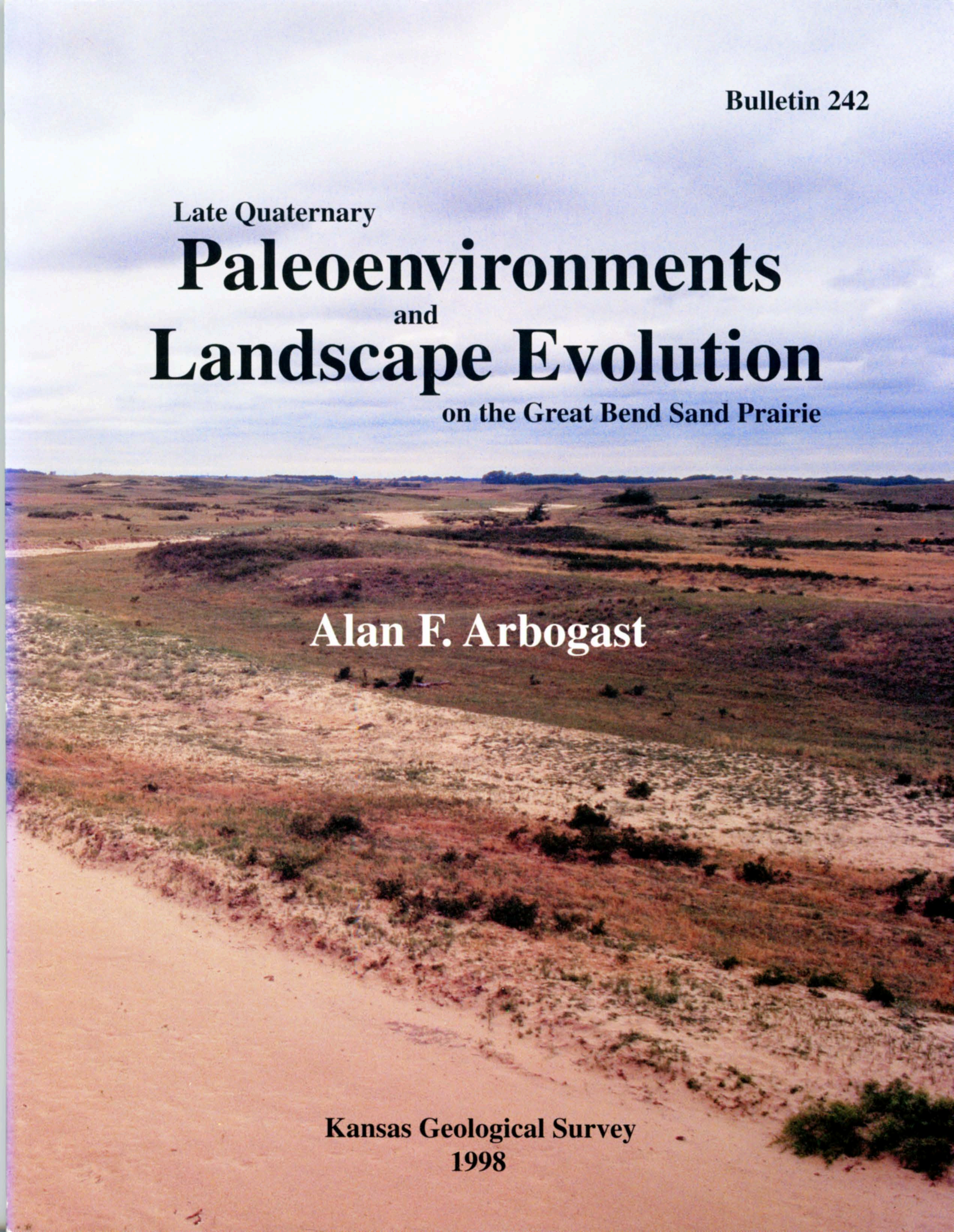

Bulletin 242

Late Quaternary Paleoenvironments and Landscape Evolution on the Great Bend Sand Prairie

Alan F. Arbogast

Kansas Geological Survey
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Cover: View to the southwest of a well-developed dune field in the southwestern part of the Great Bend Sand Prairie. The dry bed of Rattlesnake Creek is in the foreground and is a probable source for eolian sand in the dune field.

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Abstract

Global circulation models (GCM's) project enhanced warming and drying in the central Great Plains during the next few decades in response to elevated levels of atmospheric CO_2. Given the sensitivity of sand-mantled landscapes to climate changes, paleoenvironmental research has focused on the potential response of sand sheets and dune fields to increased aridity. Although appreciable research has been conducted in Nebraska, Colorado, and Texas, no detailed information has as yet emerged from Kansas.

Geomorphic research on the Great Bend Sand Prairie, a large sand sheet in south-central Kansas, indicates that two late Quaternary stratigraphic units occur in the region. The oldest deposits are late Wisconsinan, and the youngest are Holocene. Late Wisconsinan deposits are widespread, consisting largely of sand and silt (defined herein as silty sand), and probably accumulated in a very low energy fluvial environment. Radiocarbon ages from the lower part of the silty sand range from about 20,000 yr B.P. to around 9,000 yr B.P. At all localities, the silty sand contains one or two extremely well developed buried soils with stacked Bt horizons, indicating long-term landscape stability following deposition. Floral (*Picea* cf. *glauca*) and faunal (e.g., *Discus cronkhitei*) remains, as well as $\delta^{13}C$ values (e.g., -25.6‰) derived from the silty sand indicate that the late Wisconsinan climate was cooler and had more effective moisture than the climate during the Holocene. Northwesterly winds prevailed, as indicated by the orientation of Wilson Ridge, a late Wisconsinan lunette.

Overlying the silty sand are eolian sands of varying thicknesses. Radiocarbon ages from the upper 5 cm (2 in) of the underlying silty sand provide an estimate of the maximum-limiting age of dune development. At three sites, ages on the upper silty sand are late Wisconsinan, suggesting that overlying eolian sands accumulated during the Woodfordian. In most instances, however, the upper silty sand dates from 7,000 yr B.P. to 800 yr B.P., indicating that overlying dunes are largely Holocene deposits.

Mapping of Holocene landforms on uplands recognizes six categories, ranging from level sand sheets to parabolic dunes. In comparison to late Wisconsinan deposits, dune sands are well sorted, with a mean particle size of very fine to fine sand. Values of $\delta^{13}C$ (e.g., -15.0‰) derived from dunes imply a warmer climate during the Holocene than the Woodfordian. The orientation of parabolic dunes indicates prevailing, southwesterly winds. Dunes usually contain one to two, weakly developed buried soils with A/AC/C horizonation, representing brief periods of landscape stability. Calibrated radiocarbon ages at standard deviation (2σ) on buried soils imply six periods of pedogenesis during the Holocene, with the center of probabilities at ca. 6,300, 2,300, 1,500, 1,000, 700, and 200 yr B.P. Surface soils are generally poorly developed, suggesting that dunes can easily be mobilized if increased aridity occurs.

Introduction

This study reconstructs the history of late Quaternary climate change and landscape evolution on the Great Bend Sand Prairie, a large sand sheet in south-central Kansas. Prior to this investigation, research of a similar nature had been conducted only in surrounding areas of the Great Plains, including northeastern Colorado, Wyoming, Nebraska, and Texas. As a result, little was known about the late Quaternary paleoenvironmental and geomorphic history of the sand sheets in Kansas, specifically the record of desertification and eolian sand mobilization. This lack of information has precluded prediction of landscape response should Kansas experience intensive drought in the future, as is anticipated by greenhouse-warming scenarios.

It is generally thought that increased levels of greenhouse gases (e.g., CO_2, methane) in the atmosphere will result in global warming (e.g., Washington and Meehl, 1984; Hansen et al., 1988). Climatic modelling of greenhouse scenarios (e.g., Hansen et al., 1988; Schlesinger, 1989) has demonstrated that the degree of warming may reach levels reconstructed for the last few interglacials. Although the models show regional variation, greater warming is expected to cause increased dryness for the already subhumid to semiarid Great Plains (Hansen et al., 1988; Wetherald and Manabe, 1988; Wendland, 1993). Landscapes within the region that consist of unconsolidated sand are particularly sensitive to moisture loss because they easily destabilize when protective vegetation is reduced. This is significant within the context of regional warming because large portions of the Great Plains are mantled by sand dune fields and sand sheets (fig. 1) that are presently stable (e.g., Smith, 1940; Muhs, 1985, 1991; Madole, 1986, 1994; Holliday, 1989; Forman and Maat, 1990; Swinehart, 1990; Johnson, 1991).

Periodic desertification and mobilization of eolian sand has been demonstrated elsewhere in the Great Plains during the late Quaternary when intervals of drought reduced vegetative cover (Madole, 1986, 1994; Holliday, 1989; Swinehart, 1990). Dune migration apparently occurred during such cool-arid periods as the late

Wisconsinan (e.g., Wright et al., 1985; Forman and Maat, 1990) as well as such warm-arid intervals as the middle Holocene (Ahlbrandt et al., 1983; Forman and Maat, 1990). The conditions under which these geomorphically sensitive areas would desertify again is the focus of increased speculation in response to predictions of greenhouse warming and drying (e.g., Swinehart, 1990; Forman and Maat, 1990; Johnson, 1991; Muhs, 1991; Forman et al., 1992).

Although late Quaternary paleoenvironmental reconstructions have been reported for sand sheets in northeastern Colorado (Muhs, 1985, 1991; Madole, 1986, 1994, 1995; Forman and Maat, 1990; Forman et al., 1992), Nebraska (Ahlbrandt and Fryberger, 1980; Ahlbrandt et

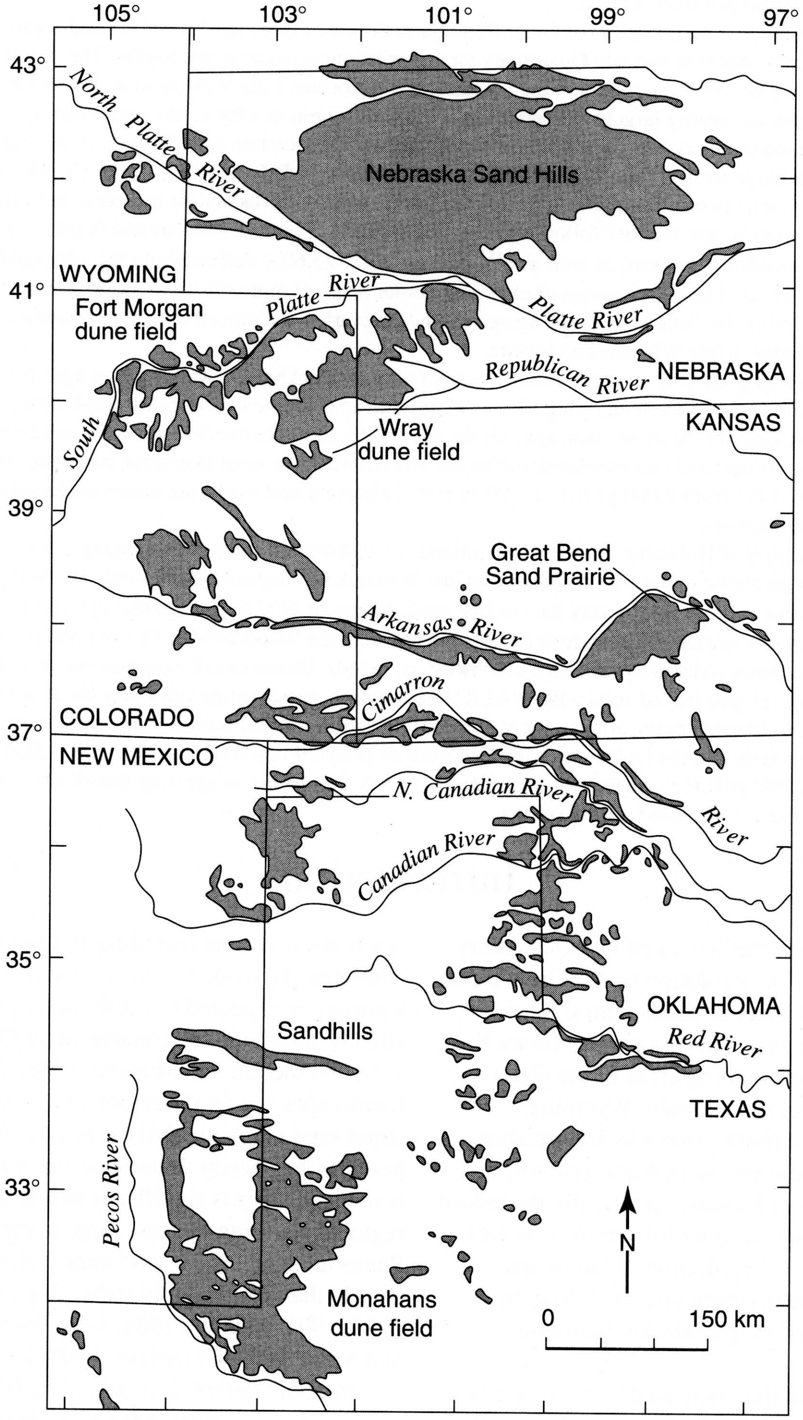


FIGURE 1—Distribution of dune fields and sand sheets in the central Great Plains (modified from Muhs and Holliday, 1995).

al., 1983; Swinehart, 1990), and Texas (Holliday, 1985, 1989), little is known about the response to climate change and the chronology of geomorphic events on sand-covered landscapes in Kansas. Significant portions of the south-central and southwestern part of Kansas are mantled by unconsolidated sand. Most of this area is agricultural land, portions of which were destabilized and intensively eroded by wind during the dust bowl of the 1930's (Smith, 1938; Latta, 1950; Simonett, 1960). Although some research has been conducted in the dune fields of the region (e.g., Moore, 1920; Courtier, 1934; Smith, 1938, 1940; Simonett, 1960), it was generally descriptive in nature because of the lack of absolute dating techniques.

The largest sand sheet in Kansas, the Great Bend Sand Prairie (fig. 1), was chosen for detailed paleoenvironmental and geomorphic investigations because of its potential to provide significant baseline data for future studies. Increased levels of aridity are predicted for the future in the central Great Plains, including the Great Bend Sand Prairie, if greenhouse warming is realized. Because sand sheets have been especially sensitive to increased warming in the past (e.g., 1930's), understanding the timing and magnitude of landscape response is critical if destabilization is to be curtailed. Accordingly, the four specific objectives of this research were to (1) construct a detailed map of surficial geology; (2) determine the number, character, and relative ages of late Quaternary stratigraphic units; (3) reconstruct the history of paleoclimatic change by analyzing the sediments of major stratigraphic units; and (4) construct a chronology of desertification and sand mobilization on the Great Bend Sand Prairie and compare it with that derived from other sand sheets in the central Great Plains.

Study Area

The research discussed in this report was conducted on the Great Bend Sand Prairie, a large sand sheet located within the "great bend" of the Arkansas River (fig. 1). Approximately 4,500 km^2 (1,100 mi^2) in size, it includes all of Stafford County, and portions of Barton, Edwards, Kiowa, Pratt, Reno, and Rice counties. The vast majority of investigations occurred within the jurisdiction of Groundwater Management District 5 (GMD 5). The study area is situated mostly within the Arkansas River Lowlands physiographic province, except for the southern one-half of Pratt county, which lies in the High Plains physiographic province (Schoewe, 1949). Major tributaries to the Arkansas River in the region are the North Fork Ninnescah River, which flows generally to the southeast, and Rattlesnake Creek, a northeasterly trending stream that bisects the study area (fig. 2).

Geology

The geology of the Great Bend Sand Prairie is complex. Structurally, the study area lies on the southwestern flank of the Central Kansas uplift (Barton arch) and the northern one-half of the Pratt anticline (Merriam, 1963) (fig. 3). Basement rocks are Permian and early Cretaceous in age, with Cretaceous rocks present and forming the bedrock surface only in the western one-half of the study area. Permian rocks, consisting of the Ninnescah Shale, Stone Corral Formation, Harper Sandstone, Salt Plain Formation, Cedar Hills Sandstone, and undifferentiated strata (including Whitehorse and Dog Creek Formations) are often referred to as "red beds" because they contain red to brown shale, siltstone, and sandstone with minor beds of limestone, dolomite, and anhydrite. Cretaceous rocks

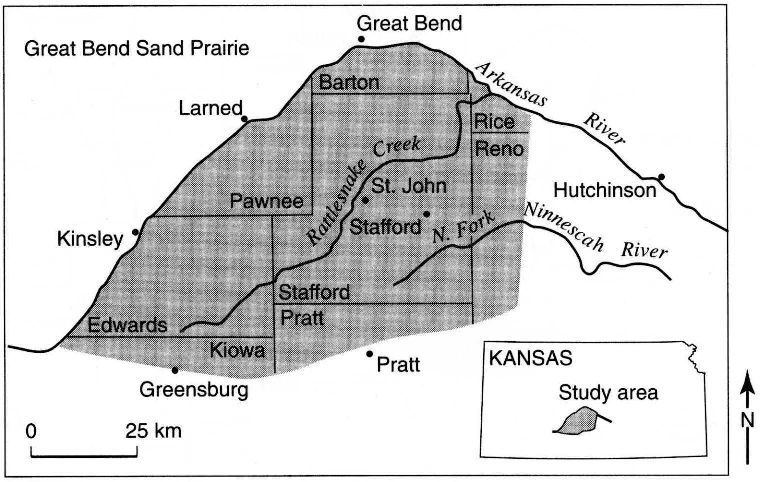


FIGURE 2—Location of major towns, county boundaries, and tributaries on the Great Bend Sand Prairie.

include the Cheyenne Sandstone, Kiowa Formation, and Dakota Formation, and generally consist of interbedded shales, sandy shales, and fine- to coarse-grained sandstones.

Overlying the Permian and Cretaceous basement rocks on the Great Bend Sand Prairie are varying thicknesses of unconsolidated Tertiary and Quaternary deposits. Tertiary deposits consist of the Pliocene Ogallala Formation. In general, the Ogallala is characterized by deposits of silt and fine sand with interbedded caliche that were derived from the Rocky Mountains (Fader and Stullken, 1978).

The surficial geology of the entire Great Bend Sand Prairie is dominated by unconsolidated Quaternary deposits of eolian and alluvial origin. Quaternary sediments of the region have a maximum thickness of about 110 m (360 ft) (table 1). The kinds of minerals (e.g., quartz, feldspar, granite) found in most Quaternary deposits suggests a Rocky Mountain origin with the Arkansas River serving as the primary source. The bend of the Arkansas River is thought to have migrated laterally from the south to its present position via successive captures by its own northern tributaries, leaving a thick deposit of sand, silt, and clay behind (Fent, 1950). In general, five lithostratigraphic units have been recognized (bottom to top): (1) basal sand and gravel; (2) alternating sequences of sandy silt-clay, sand, and gravel; (3) near-surface silt-clay bed; (4) loess; and (5) dune sand (Rosner, 1988). Of specific concern in this study is the surficial mantle of dune sand, from 0 m to 15 m (0–49 ft) thick (Johnson, 1991), and its relationship to underlying deposits of silt-clay and loess (e.g., fig. 4).

Soils

Soils within the Great Bend Sand Prairie are classified as Mollisols, Alfisols, Entisols, or Inceptisols. Categorization is based on landscape position and parent-material associations. Upland soils have formed in undulating to hilly sandy sediments; in so-called old alluvium that is predominantly sandy clay loam, silty clay loam, and clay loam in texture; and in loess. Floodplain soils are those that have developed in areas with a seasonally high water table or in stream drainageways (or both).

The best-developed soils in the study area are Typic Argiaquolls (Carwile Series), Udic Argiustolls (Naron Series), Pachic Argiustolls (Blanket and Farnum Series), and Vertic Argiustolls (Tabler Series). These soils are loamy, generally considered to have formed in old alluvium, and occur on the broad landscapes of relatively low relief between large dune fields. Soils in the Tabler Series have the finest texture, generally occupy depressional positions, and are the least well drained. Carwile soils occur in similar topographic positions as Tabler soils but are more coarse textured and slightly better drained. Naron and Farnum soils contain the highest proportions of sand, occupy slightly higher landscape positions, and are better drained. Blanket soils are found only in Stafford County where a significant deposit of loess occurs.

Soils that have evolved in the complex, wind-modified dune topography of the study area consist of Psammentic Haplustalfs (Pratt Series), Typic Ustipsamments (Tivoli Series), and Aquic Ustipsamments (Dillwyn Series). Each has formed in sediments classified as loamy fine sand. Dillwyn soils are deep, somewhat

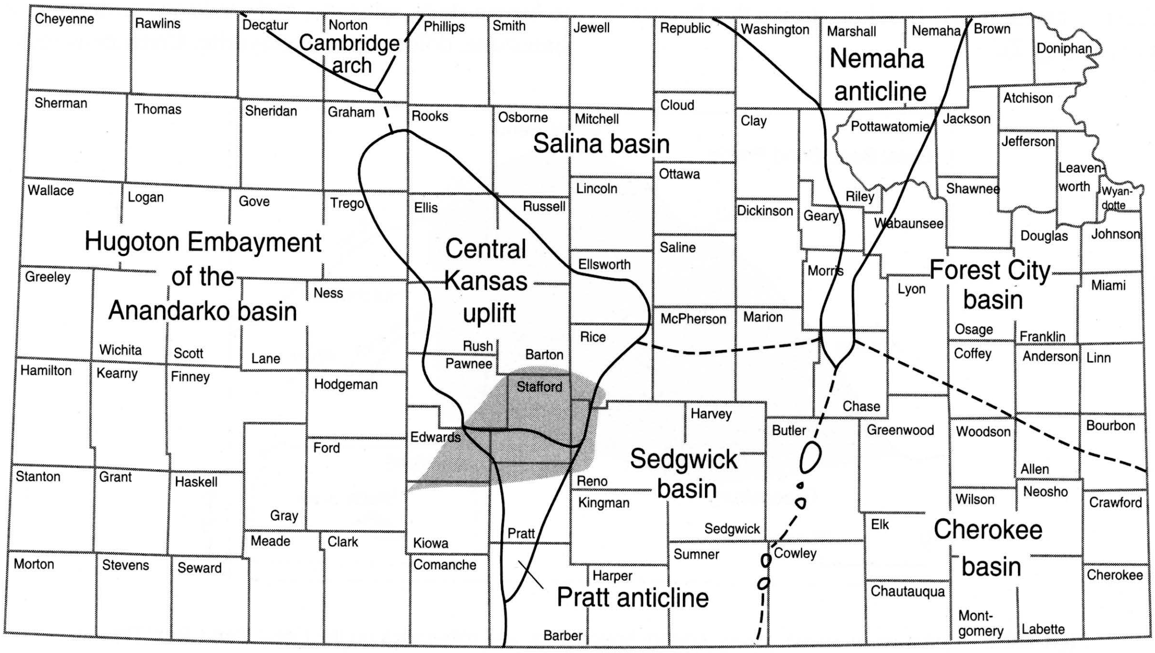


FIGURE 3—Structural elements of Kansas portraying the position of the Great Bend Sand Prairie (screened area) relative to the Central Kansas uplift and Pratt anticline.

TABLE 1—GENERALIZED COLUMNAR SECTION OF GEOLOGIC UNITS ON THE GREAT BEND SAND PRAIRIE (modified from Fader and Stullken, 1978).

System	Geologic Unit	Maximum thickness (m)	Physical Character
Quaternary	undifferentiated Pleistocene deposits	110	Unconsolidated deposits of sand and gravel with interbedded lenses of clay, silt, and caliche. Eolian sand occurs over most of the area.
Tertiary	Ogallala Formation (Pliocene deposits)	20	Unconsolidated deposits of silt, fine sand, and interbedded caliche.
Cretaceous	undifferentiated Lower Cretaceous rocks	115	Upper unit (Dakota Formation): brown, fine- to medium-grained sandstone interbedded with shale. Middle unit (Kiowa Formation): dark-gray shale interbedded with sandstone. Lower unit (Cheyenne Sandstone): gray to brown medium-grained sandstone interbedded with shale.
Permian	undifferentiated Permian rocks	105	Interbedded reddish shale, siltstone, and sandstone.
	Cedar Hills Sandstone	60	Reddish shale, siltstone, and sandstone.
	Salt Plain Formation	90	Reddish-brown sandy siltstone and sandstone.
	Harper Sandstone	75	Brownish-red siltstone and silty shale.
	Stone Coral Formation	5	White and light-gray anhydrite and dolomite.
	Ninnescah Shale	120	Red and grayish-green shale, siltstone, and silty sandstone.

FIGURE 4—OUTCROP OF SILTY SAND IN A DUNE FIELD in the NW, NW, sec. 24, T. 26 S., R. 15 W. (modified from Arbogast and Johnson, 1998).

poorly drained soils in interdunes where seasonal water tables are relatively high. Pratt soils are well drained and occupy the lowest, least erodable slopes on dunes. Tivoli soils are also well drained but are found on dune crests where eolian erosion is most likely to occur. As a result, these soils have the poorest development of any series in the region.

Soils that have formed in younger, fluvial landscapes are classified as Fluvaquentic Haplustolls (Plevna Series) and Leptic and Typic Natrustolls (Natrustolls). Natrustolls developed in loamy, calcareous alluvium that contains layers of sand or clay in places. They are somewhat poorly drained and often contain high concentrations of salt. Seasonal water tables are relatively high. Plevna soils are often heavily gleyed and typically have developed in slight depressions on floodplains and on chaotic, channeled floodplains. Parent material is usually fine, sandy loam at the surface that is underlain by sandy and clayey alluvium (Dodge et al., 1978).

Climate

Several stations (e.g., Great Bend, Hutchinson, Greensburg, Hudson, Larned) near or on the Great Bend Sand Prairie have maintained temperature and precipitation records since early in this century. Data indicate that the present climate of the region is semiarid to subhumid and strongly continental, characterized by extreme diurnal and annual variations in temperature. Winter usually lasts from December to February with an average low temperature of about 0°C. The summer growing season is usually from April to October, during which time the average high temperature is about 25.5°C. Average annual precipitation in the study area reflects the position of the Great Bend Sand Prairie on the boundary between the dry portion of western Kansas that is influenced by the rain shadow of the Rocky Mountains and the eastern regions over which moisture-laden air from the Gulf of Mexico flows. Although yearly precipitation may vary widely depending upon the precise position of the boundary, mean annual precipitation on the western border (57 cm; 22 in) is significantly less than average yearly rainfall on the eastern margin of the study area (80 cm; 32 in) (Fader and Stullken, 1978). Most of the total annual precipitation comes from convective storms in the late spring and summer, with approximately 75% of the yearly amount occurring in the growing season (table 2).

Vegetation

Vegetation on the Great Bend Sand Prairie can be separated into that which was native to the area and that which was imported by European settlers (Kuchler, 1974). Land-use data for Stafford County serve as an estimate for this division between native and cultivated vegetation on the Great Bend Sand Prairie. In 1978, approximately 375,000 acres, or about 75% of the total land area in Stafford County, was under cultivation; 96,000 acres, or about 19% of the total land area was in pasture. The remaining 6% of the land, such as along the Rattlesnake Creek floodplain, was regarded as unsuited for agriculture. Pasture in Stafford County is generally located on the very well drained, dune topography of high relief, while cultivated land is found nearly everywhere.

Grassland, in the form of tall- and mixed-grass prairie, dominates the study area, although wooded areas exist along streams. Common native prairie grasses in high-relief dune fields include sand bluestem (*Andropogon hallii*), little bluestem (*Andropogon scoparius*), sand lovegrass (*Eragrostis trichodes*), big sandreed (*Calamovilfa gigantea*), switch grass (*Panicum virgatum*), indian grass (*Sorghastrum nutans*), sand dropseed (*Sporobolus cryptandrus*), and Texas bluegrass (*Poa arachnifera*). In areas of less relief, where better-developed, heavier soils are found, big bluestem (*Andropogon gerardii*), western wheatgrass (*Agropyron smithii*), blue grama (*Agropyron hallii*), side-oats gramma (*Bouteloua curtipendula*), tall dropseed (*Sporobolus giganteus*), and buffalo grass (*Buchloe dactyloides*) are common. Riparian trees in the Great Bend Sand Prairie include eastern cottonwood (*Populus deltoides*), American elm (*Ulmus rubra*), common hackberry (*Celtis occidentalis*), sycamore (*Platanus occidentalis*), black hickory (*Carya texana*), and black walnut (*Juglans nigra*) (Dodge et al., 1978).

Cultural History

Archaeological evidence recovered from the Great Bend Sand Prairie indicates human inhabitation during the past several thousand years. Eighteen prehistoric sites have been identified thus far, dating back to the Archaic, and Middle and Late Ceramic. Tool assemblages, consisting largely of projectile points and flakes, suggest the region served as a hunting ground. The Great Bend Sand Prairie was intermittently inhabited by nomadic bands of Native American Wichita, Cheyenne, Comanche, Kiowa, and

TABLE 2—MEAN MONTHLY TEMPERATURE AND PRECIPITATION AT HUDSON. Data from Dodge et al. (1978, p. 56).

	Temperature (°C)		Precipitation (mm)
	Mean Daily Max.	Mean Daily Min.	
J	6.0	-7.0	1.4
F	9.0	-4.4	2.3
M	13.0	-1.2	3.2
A	20.5	5.9	5.8
M	25.9	12.0	9.2
J	31.2	17.1	10.8
J	34.0	18.9	9.3
A	33.3	18.9	6.5
S	28.3	14.0	5.6
O	22.2	7.7	5.3
N	13.3	0.0	2.0
D	6.8	-4.5	2.2

Kiowa Apache during the protohistoric (Logan et al., 1993).

Large-scale European influence in the region began during the 1820's with establishment of the Santa Fe Trail along the northern boundary of the study area. In order to protect the trail, Fort Larned was established in 1859 near the site of present-day Larned, Kansas. By the late 1860's, homesteaders from the eastern United States and Europe began to settle the area during the large westward migration. Migration was fueled by the establishment of the Santa Fe Railroad, which reached the town of Great Bend in 1872 (Dodge and Roth, 1978).

By 1900 the area was fully settled by European settlers. Pratt, Reno, Barton, and Edwards counties were established in 1872, whereas Edwards and Stafford counties were organized in 1874 and 1879, respectively (Andreas, 1883). Kiowa County was established in 1886 (Hoffman et al., 1986). According to the 1990 census, the largest cities in the region are Great Bend (pop. 15,427), Pratt (pop. 6,687), Larned (pop. 4,445), Greensburg (pop. 1,792), St. John (pop. 1,274), and Stafford (pop. 1,268). Although some oil and gas speculation occurs in the area, the regional economy is largely based on farming, ranching, and related enterprises (Kansas Statistical Abstracts, 1992).

Previous Research

Numerous studies have been conducted about paleoenvironmental change in the central Great Plains, including the history of eolian sand erosion, transport, and deposition. The following discussion reviews significant research that is particularly germane to this study, beginning with the chronology and nature of paleoenvironmental change in the region. Within that context, the known geomorphic history of dune fields in the central Great Plains is subsequently examined in the Arkansas River valley of Kansas, the southern High Plains of Texas, northeastern Colorado, and the Nebraska Sand Hills.

Late Quaternary Environmental Change in the Central Great Plains

Paleoenvironmental studies in the central Great Plains indicate that climatic conditions have varied dramatically over the chronological subdivisions of the past 25,000 years (fig. 5). Despite the numerous studies that have been conducted, conflicting evidence exists regarding the character of late Wisconsinan environments in the central Great Plains. It is generally agreed that during the Woodfordian temperatures were cooler conditions and there was less seasonality than at present. Climatic models (e.g., CLIMAP Members, 1981; Kutzbach, 1987; COHMAP Members, 1988) depicting atmospheric circulation during the glacial maximum simulate a westerly jet stream that divided into northern and southern branches west of the Laurentide ice sheet. Given this split, the mean position of the southern branch of the polar front, which is presently in southern Canada, was around 34°N about 18,000 yr B.P. (Delcourt , 1979; Delcourt and Delcourt, 1983). As a result, mean annual surface temperatures in the central Great Plains were probably 2° to 4° cooler than today (Kutzbach, 1987). Wells (1983) used the orientation of late-glacial features to argue that prevailing winds were northwesterly. In addition, they were perhaps 20% to 50% stronger than at present (Crowley and North, 1991). Analysis of Pleistocene fauna further imply decreased seasonality, indicating more complex and diverse biological communities (Martin, 1984; Martin and Hoffman, 1987; Martin and Martin, 1987).

Although surface temperatures for the Woodfordian are relatively well understood, there is uncertainty regarding the levels of effective moisture. The majority of evidence suggests that the climate was relatively moist, at least compared to the Holocene. In particular, the floral record implies that levels of effective moisture were comparably high. Grass phytoliths derived from Peoria loess, exposed at the Eustis ash pit in south-central Nebraska, indicates deposition on a well-vegetated surface (Fredlund et al., 1985). According to Fredlund (1995), the pollen record at Cheyenne Bottoms, located in central Kansas, suggests a spruce parkland around 20,000 yr B.P. In the Arkansas River floodplain, near Wichita, remains (e.g., spruce wood and needles, cone fragments) of coniferous species were recovered in peat that dated to about 19,000 yr B.P. (Fredlund and Jaumann, 1987). Data from Harlan County Lake in south-central Nebraska, as

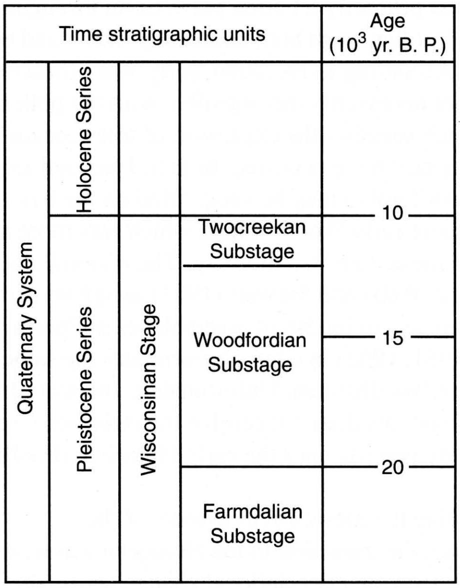


FIGURE 5—LATE QUATERNARY CHRONOLOGY USED IN THIS STUDY. Pleistocene-Holocene boundary after Hopkins (1975). Farmdalian, Woodfordian, and Twocreekan substages after Frye et al. (1965).

well as Sanders's well in northeastern Kansas, indicate that uplands were populated by *Populus* (aspen) between 24,000 and 12,800 yr B.P. (Fredlund and Jaumann, 1987; Fredlund, 1989). Moreover, Wells and Stewart (1987) reported that *Pinus flexilis* (limber pine) and *Picea* cf. *glauca* (spruce) charcoal-dated to about 14,500 yr B.P. in south-central Nebraska.

Although some evidence indicates relatively high levels of effective moisture during the Woodfordian, other data point to increased aridity. In particular, Holliday (1987) argued that pollen data are misleading because spruce pollen is more resistant to weathering than other pollen taxa. Moreover, Holliday contended that buried soils of Woodfordian age should exhibit evidence of podzolization if conifers truly dominated. According to Feng (1991), no evidence for podzolization (e.g., spodic horizon) exists in the Peoria loess.

In addition to pollen and soils, some sedimentary evidence suggests that the glacial maximum was a period of increased aridity. During the Farmdalian, for example, uplands were very stable, with extended soil development in the Gilman Canyon Formation until around 22,000 yr B.P. (Fredlund et al., 1985; Johnson, 1993; Johnson et al., 1993). Subsequently, sedimentation of Peoria loess began, perhaps due to less effective moisture in the region (e.g., Johnson, 1993; Johnson et al., 1993). According to Johnson (1993), this hypothesis is supported by modelling of environmental conditions at the glacial maximum (e.g., Kutzbach and Wright, 1985, COHMAP Members, 1988; Crowley and North, 1991).

Evidence also points to increased late Wisconsinan aridity in south-central Kansas. Fredlund (1995) noted a major unconformity at Cheyenne Bottoms, one that spans the entire Woodfordian. He hypothesized that Cheyenne Bottoms was primarily a barren playa basin throughout the early Woodfordian, with both eolian deposition and erosion occurring. According to Fredlund, early Woodfordian aridity is not necessarily incompatible with the pollen record, which suggests the expansion of trees because limited taxa may have persisted. In fact, Fredlund argues that the Woodfordian may be subdivided on the basis of climate into an early Woodfordian, which was more xeric, and a more mesic late Woodfordian. The regional data are inconclusive. Wells and Stewart (1987) assign an early Woodfordian age to landsnail assemblages in Peoria loess. Leonard (1951, 1952) in contrast, correlates the same fauna with the late Woodfordian. Unfortunately, the record from Cheyenne Bottoms does not resolve the issue because the unconformity persists until the early Holocene (Fredlund, 1995).

Following the mesic environments of the Woodfordian, the transition to the Holocene was a period of major climatic and vegetational change across the central Great Plains. The driving mechanism behind this change were adjustments in the geometric relationship of the earth to the sun. Perhelion was in July, and this, in conjunction with a decrease in the earth's tilt (obliquity 24.23°), resulted in summer solar radiation at the top of the atmosphere that was about 8% greater than at present (Kutzbach, 1981, 1985, 1987). Given the increased radiation, the Laurentide ice sheet rapidly began to disintegrate (Andrews, 1987), promoting generally drier, zonal atmospheric flow (Knox, 1983). In south-central Kansas, these new atmospheric patterns caused a dramatic increase in Chenopodeaceae-Amarathanceae (Cheno-Am) populations soon after 11,000 yr B.P. in the basin at Cheyenne Bottoms (Fredlund, 1995). According to Fredlund, this increase reflects sharp fluctuations in water levels as the climate became more variable. At Muscotah Marsh in northeastern Kansas, *Picea* sharply declined around 12,000 yr B.P., concurrent with an increase in populations of *Quercus, Ulmus, Fraxinus,* and *Salix.* By the very end of the Woodfordian (~10,500 yr B.P.), *Picea* had completely been replaced by deciduous forest, which, in turn, was completely displaced by grassland by 9,000 yr B.P. (Gruger, 1973).

As the Laurentide ice sheet continued to waste during the early Holocene, the steep north-south temperature gradient which had been present during the Woodfordian further weakened, promoting further zonal flow (Knox, 1983). As a result, seasonal temperature extremes began to increase (COHMAP Members, 1988). Ultimately, these combined factors triggered the generally warm and dry conditions of the Altithermal (Antevs, 1955), which occurred from about 8,000 to 5,000 yr B.P. in central North America (Knox, 1983; Kutzbach, 1985, 1987; COHMAP Members, 1988; Crowley and North, 1991). By approximately 6,000 yr B.P., mean summer temperatures in the region were probably 2° to 4° higher than at present (COHMAP Members, 1988; Crowley and North, 1991), and annual precipitation was perhaps 25% less (Bartlein et al., 1984; Kutzbach, 1987). In Cheyenne Bottoms, Fredlund (1995) reported that water levels stabilized but were probably lower, depressing Cheno-Am populations.

Following the Altithermal, the climate of the central Great Plains was apparently more moist in the early part of the late Holocene. Evidence for this shift was reported at Muscotah Marsh in northeastern Kansas, where deciduous forest (e.g., *Quercus, Carya*) briefly repopulated portions of the landscape after 5,000 yr B.P. (Gruger, 1973). Since that time, however, the climate has apparently fluctuated between relatively moist and arid. According to Fredlund (1995), for example, high percentages of Cheno-Am pollen in late Holocene sediments in Cheyenne Bottoms indicate that water levels have varied, resulting in periodic drying of the basin in the past few thousand years.

Evolution of Dune Fields in the Central Great Plains

Arkansas River Valley of Kansas

Prior to the 1950's, geomorphic investigations of regional sand sheets were mostly qualitative in nature. In general, the research was speculative with regard to the chronology of events and associated climatic history because reliable age-determination methods (e.g., carbon-14, thermoluminescence) were not available. According to Ahlbrandt et al. (1983, p. 379), inferences were based upon correlations with loess or terrace-fill sequences and upon assumptions concerning strong katabatic winds during glacial periods. Much of the early research was conducted in Kansas, where interest focused on the dune fields along the Arkansas River valley.

In a general reconnaissance of southwestern Kansas, Hay (1893) thought the sand in the sand dunes along the Arkansas River was from a local bedrock source (e.g., the Ogallala Formation). Haworth (1897) conducted a study of the physiography of western Kansas, noting that the dune sand resembled valley sands along the river. In an overview of the geology of Kansas, Moore (1920) argued that the dune sand along the Arkansas River was derived from the floodplain, having been blown out to the south by northwesterly winds.

The most complete, early study of the physiography and geology of south-central Kansas was conducted by Courtier (1934). After observing that the prevailing winds during the early Dust Bowl years were southwesterly, Courtier assumed similar conditions existed when the dunes originally formed. As a result, he concluded that the residual soils from the local Dakota, Ogallala, and Sanborn Formations were the sand sources and not the Arkansas River to the north. Referring to the Great Bend Sand Prairie region specifically, he suggested the name "Great Bend sand plains" for the portion of the area covered by sand.

Additional paleoenvironmental evidence of dune age in Kansas was supplied by Smith (1938) from a small study near Hutchinson, Kansas. He observed three soil zones exposed in a blowout, which suggested at least three periods of "alternating sand movement and stabilization" (p. 115). Also exposed in the section were the fossil remains of *Bison alieni,* which confirmed a Pleistocene age for a least one of the units. According to Smith, strike and dip in a reddish-brown basal dune sand indicated northeasterly winds prevailed when the dune formed. The dune appeared to have been reworked by winds of varying directions, however, with modern blowouts resulting from southerly winds.

Smith (1939, 1940) argued that vegetation plays an important role in dune morphology on the Great Plains since pervasive dune forms (e.g., parabolic) suggest that true desert conditions have never prevailed. As a result, he concluded that sand dunes on the Great Plains evolve in two phases, eolian and eluvial. According to Smith, dunes grow during the eolian phase, when deflation removes sand from one part of the dune and deposits it nearby where vegetation is thick. In contrast, dune degradation occurs during the eluvial phase, when vegetation is sufficient to stop further erosion. As the eluvial phase progresses, dune contours are rounded by soil formation, creep and slope wash. Smith further argued that the cycle may be interrupted at any time and, in fact, that multiple cycles had probably occurred, giving most dunes in Kansas what he called a chaotic appearance.

Smith (1940) also considered the source of sand dunes in southwestern Kansas and prevailing paleowind directions. Since the dunes are not migratory in nature, he concluded the source for sand must be local, probably a combination of fluvial and outcropping bedrock origins. Interestingly, he thought that fluvial sources contributed to the smaller dune fields in the region, but that the Arkansas River was not the origin of sand for the massive dune field along its length because dunes were not presently moving across the floodplain. He acknowledged that the dune belt along the south side of the Arkansas River valley suggested that northwesterly winds had prevailed (implying a floodplain source) when the dunes formed. According to Smith (1940), however, dominant modern winds are southerly because transverse ridges near Syracuse, Kansas, face north.

The last substantative work of a qualitative nature about sand dunes in Kansas was done by Simonett (1960) in a study of dune development and stratigraphy along the Arkansas River near Garden City and Syracuse. At most localities, Simonett found Peoria loess underlying the dune sand, a relationship that indicated to him a northerly source for the sand because the loess crops out farther south. Moreover, he recognized northerly dipping, low-angle backset beds in many dunes, further suggesting a northerly sand source. As a result, Simonett concluded that "Wisconsin alluvials from the Arkansas River" were the likely source for the dune sand and that most sand movement occurred during the Wisconsinan when northerly winds prevailed (1960, p. 223).

Simonett (1960) presented evidence, however, that recent dune-forming winds have been southerly. He agreed with Smith's (1940) interpretation that unvegetated transverse dunes and barchan dunes near Syracuse with steep, north-facing, slip faces indicate southerly winds. In addition, he recognized parabolic dunes nearby with arms that open upwind to the south. Simonett (1960) also discovered that the line of interdune depressions near Garden City runs southwest to northeast. Based on this combined evidence, he concluded that recent southerly winds have reworked dunes initially formed by northwesterly winds.

After a hiatus of about 30 years, new information regarding the origin and age of sand dunes in Kansas, specifically from the Great Bend Sand Prairie, has recently emerged. In a general study of late Quaternary stratigraphy of the region, Rosner (1988) concluded that surficial deposits of eolian sand overlie a poorly sorted silt-clay layer. Subsequently, Johnson (1991) referred to the silt-clay layer as the silt layer and determined that it was probably Peoria loess. Johnson obtained five radiocarbon ages, largely late Holocene in age, from the upper part of the silt layer where it was overlain by dune sand. As a result, he concluded that the ages represented intervals when the silt-layer was last exposed. In a review of the early descriptions of the Arkansas River valley by 19th-century explorers, Muhs and Holliday (1995) found evidence for both active and inactive sand in the area of the Great Bend Sand Prairie. At several localities, the presence of naked sand was noted in addition to tracts that were densely populated by sunflowers. In the western part of the region, near Pawnee County, the dunes were apparently more active, with large areas totally void of vegetation. This pattern of active and inactive dunes was described into Colorado, leading the authors to conclude that the degree of eolian sand mobilization along the Arkansas River in western Kansas varied spatially.

Despite the recent interest in the Great Bend Sand Prairie, the focus of research on Great Plains sand sheets in the past 30 years has clearly been in Texas, Colorado, and Nebraska. Moreover, geomorphic histories of dune fields in these areas are much more detailed and quantitatively based than those in Kansas. Palynological evidence, for example, has been used increasingly to determine local paleovegetation. In addition, age-determination techniques, specifically carbon-14, optical simulated luminescence (OSL), and thermoluminescence (TL), have been successfully used to establish chronologies of stability and instability within dunes.

Southern High Plains

Although data have begun to emerge regarding mobilization of eolian sand in Oklahoma (Brady, 1989; Olson et al., 1995), research south of Kansas has centered on the sand sheets and dune fields of the southern High Plains in the Texas panhandle. The earliest extensive work was conducted by Melton (1940), who classified the dune fields north of Lubbock into three series based upon their age. Reeves (1965) studied the lunettes surrounding the playas on the southern High Plains, concluding that they formed as playas deflated during dry interpluvials. Reeves obtained a radiocarbon age of approximately 19,000 yr B.P. for lacustrine strata beneath a dune on the southern margin of a playa, indicating that encroachment occurred during the Wisconsinan by northwesterly winds. He argued that dunes on the playa's northern and eastern margins had formed in the last 5,000 years due to strong, southerly winds.

The most extensive research on dune fields in the region has been conducted in the past 10 years around Lubbock Lake, Texas. Holliday (1985) excavated a lunette associated with Cone Playa and recognized seven stratigraphic units and five buried soils in the dune that yielded ages ranging from about 34,000 yr B.P. to 1,400 yr B.P. A detailed study of Holocene eolian sedimentation on the southern High Plains was reported by Holliday (1989). Using stratigraphic data from a variety of sites (e.g., Blackwater Draw, Yellowhouse Draw, Lubbock Lake), he reached the following conclusions: (1) eolian sedimentation in the region first occurred locally between about 10,000 yr B.P. and 9,000 yr B.P.; (2) eolian sedimentation was episodic but widespread from 9,000 yr B.P. to 5,500 yr B.P., with most areas affected by 6,500 yr B.P.; (3) eolian sedimentation occurred at all localities between 5,500 yr B.P. and 4,500 yr B.P.; and (4) landscapes have been stable for the past 4,500 years. Of particular importance, Holliday (1989) reported that two pulses of eolian sedimentation, separated by a brief period of soil formation, occurred during the middle Holocene: about 6,300 yr B.P. to 5,000 yr B.P. and 5,000 yr B.P. to 4,500 yr B.P. Using previous research (e.g., Wendorf and Hester, 1975; Johnson, 1986, 1987) and data on fossil gastropods collected from the Lubbock Lake site, Pierce (1987) reviewed the history of late Quaternary climate change on the southern High Plains. He argued that mean annual temperature increased from a low of 7.5°C during the late Wisconsinan to an Altithermal high of 20°C. During the late Holocene, increased precipitation, along with a decrease in mean annual temperature to about 17.5°C, temporarily re-established more mesic conditions in the region. According to Pierce (1987), climate has changed slightly to the present semi-arid conditions in the past 1,000 years.

The latest research from the southern High Plains was reported by Holliday (1995). Radiocarbon ages, soils, and archeological evidence indicate sedimentation in four phases during the past 11,000 years. In the latest Pleistocene, sand sheets accumulated in vegetated valleys and lake basins. The first widespread dune formation occurred between about 10,000 yr B.P. to 8,000 yr B.P. Evidence for middle Holocene eolian deposition exists in valley fills, but upland dunes were probably reworked in the later Holocene. The last phase of dune construction occurred in the past 3,000 years, with multiple episodes of stability characterized by buried soils with A/C or A/Bw profiles.

Northeastern Colorado

Abundant, detailed research has been conducted on the sand sheets and dune fields of northeastern Colorado, specifically the Hudson, Wray, and Fort Morgan dune fields. In a detailed study of these localities, Muhs (1985) classified dune types and determined their relative ages and source. According to Muhs, the pervasive dune type is parabolic with arms that point 30–40° west of north, indicating prevailing northwesterly winds and the South

Platte River as a sand source. Muhs estimated the approximate age of the dunes by comparing surface soil development there with localities in Nebraska studied by Ahlbrandt and Fryberger (1980) and Ahlbrandt et al. (1983). He concluded that the soils, which are mostly Typic Ustipsamments with A/AC/C profiles, are similar in development. As a result, he inferred that the Wray and Fort Morgan dune fields formed at the same time as a major period of dune development in Nebraska (3,000–1,500 yr B.P.), once thought by Ahlbrandt and Fryberger (1980) and Ahlbrandt et al. (1983) to correlate with the interstade between the Triple Lakes and Audubon glacial advances in the Front Range (Benedict, 1973).

Additional studies in northeastern Colorado have been reported by Forman and Maat (1990), Forman et al. (1992), and Madole (1994, 1995). Forman and Maat (1990) used soil morphology and age to conclude that reactivation of the Hudson dune field occurred sometime between 9,000 yr B.P. and 7,000 yr B.P. and that the dunes stabilized only in the past 3,000 years (Forman and Maat, 1990). Subsequently, Forman et al. (1992) analyzed the regional dunes through principal components analyses of Landsat Thematic Mapper (TM) imagery. In addition, the stratigraphic record from a 7-m (23-ft)-thick section of sheet sand near Hudson, Colorado, was described. Radiocarbon ages from four buried soils exposed in the section indicated that the sands at this locality have been reactivated at least four times during the Holocene: 9,500–5,500 yr B.P., 5,500–4,800 yr B.P., 4,800–1,000 yr B.P., and <1,000 yr B.P.

The most recent research from dune fields in northeastern Colorado was conducted by Madole (1994, 1995). Two buried soils were recognized in dunes at five, widely scattered localities along the South Platte River valley. Typically, the lower soil was better developed, with A/Bw/C horizonation, whereas the upper one was an A/C profile. Radiocarbon ages derived from the total humate fraction of the sola indicated that significant activation of eolian sand has occurred in the last 1,000 years, resulting in 3–4 m (10–13 ft) of sand accumulation. From the lower solum, radiocarbon ages of about 1,400, 1,200, and 900 yr B.P. were derived. In contrast, the upper soil yielded ages of approximately 1,000, 900, and 800 yr B.P. Values of $\delta^{13}C$ ranged from -18.2‰ to -15.2‰ (parts per thousand), suggesting to Madole that plants with a C_4 pathway (warm, dry-adapted) have inhabited the region in the past 1,000 years.

Nebraska Sand Hills

The most extensive, long-term study of sand dunes and sand sheets in the Great Plains has been conducted in the Sand Hills of Nebraska. The Sand Hills cover approximately 50,000 km^2 (12,000 mi^2), by far the most extensive area of eolian sand in the western hemisphere (Smith, 1965). Dunes are also much larger in the Sand Hills than elsewhere on the Great Plains. Individual barchan-ridge dunes, for example, may be as much as 40 km (130 ft) long and 150 m (490 ft) high. Average parabolic dune length and height is approximately 450 m (1,475 ft) and 20 m (65 ft), respectively (Swinehart, 1990). The Pliocene Ogallala Formation, either as in-situ deposits or reworked alluvial sediments, is thought to be the source of the dune sand (Lugn, 1935; Swinehart, 1990).

The age of the Sand Hills is controversial, with active debate about whether the dunes formed during the late Wisconsinan or Holocene. Early research suggested primary mobilization of sand during the late Wisconsinan in a periglacial environment. Watts and Wright (1966), for example, obtained a radiocarbon age of about 13,000 yr B.P. on organic sediments at the base of the core in an alluviated lowland between two large dunes at the Rosebud site. Palynological evidence from the core indicated that a boreal spruce (*Picea* cf. *glauca*) forest existed in the region about 13,000 yr B.P. that was soon replaced first by pine (*Pinus ponderosa*) and then by grassland. According to Bradbury (1980), the palynological data obtained by Watts and Wright (1966) placed several constraints on the time of eolian deposition in the Sand Hills. Bradbury argued that dunes were probably present before the boreal forest, but that dunes could not have formed during forest occupation of the region. Accordingly, he concluded that the last, major period of dune movement occurred after the forests left, most likely during the middle Holocene.

Ahlbrandt et al. (1983) suggested that ages obtained from interdunes by Watts and Wright (1966) and Bradbury (1980) are unreliable chronostratigraphic markers because their association with dune sand has not been positively established. Instead, Ahlbrandt et al. (1983) argued that the only reliable estimates of dune mobilization are those from material directly underlying eolian sand. They reported several radiocarbon ages, ranging from approximately 10,000 yr B.P. to 900 yr B.P., on organics buried by dune sand. Based on these ages, they recognized two distinct and one possible period of Holocene dune formation in the Sand Hills. Radiocarbon ages of about 7,200 yr B.P. and 5,100 yr B.P. bracket the Altithermal, suggesting the first episode of dune formation in the region. The best-documented phase of eolian activity is the period between approximately 3,000 yr B.P. and 1,500 yr B.P. Another, poorly documented, period of dune formation may have occurred in the latest Holocene, as indicated by a radiocarbon age of about 900 yr B.P. obtained from organic-rich sand underlying 8 m (26 ft) of dune sand.

In the most extensive report regarding the Sand Hills, *An Atlas of the Sand Hills,* Swinehart (1990) included a detailed map of the region that illustrates the distribution of eight dune types, based on McKee's (1979) classification system. Swinehart also reported a radiocarbon age of 13,160 ± 450 yr B.P. obtained from organics buried beneath 50 m (164 ft) of dune sand and 3 m (10 ft) of alluvial sand. In general, Swinehart (1990) concluded that the Sand Hills most likely formed between 8,000 yr B.P. and 5,000 yr B.P. Following about 2,000 years of stabilization, significant reactivation occurred between 3,500 yr B.P. and 1,500 yr B.P. when linear and parabolic dunes developed in areas previously not covered with sand and on preexisting dune topography.

Since Swinehart's (1990) study, additional data have emerged, confirming substantial, late-Holocene mobilization of the Sand Hills. Based on radiocarbon ages derived from interdune fens, Ponte et al. (1994) hypothesized two episodes of eolian activity in the Nebraska Sand Hills during the late Holocene: from 3,500 yr B.P. to 2,800 yr B.P and after 1,000 yr B.P. In addition, Muhs et al. (1995) reported 17 new radiocarbon ages that showed dune mobilization in the past 1,000 years.

Clearly, much research has been conducted in Great Plains sand sheets and dunes this century. The earliest work (1920's–1960's) focused on the dune fields of Kansas, but was qualitative in nature and lacked absolute age control. Conclusions were based upon loess or terrace-fill sequences and oversimplified glacial chronologies. Generally, dunes along the Arkansas River valley were thought to have formed during the late Wisconsinan, when strong winds theoretically blew off of the Laurentide ice sheet to the north. In fact, the primary evidence of a late Wisconsinan age for the dunes was their position on the southern side of the river which, hypothetically, would have occurred only as a result of persistent northerly winds.

Detailed chronologies and associated climatic histories have been constructed for a variety of sand sheets and dune fields on the Great Plains since the advent of the radiocarbon dating technique in the 1950's. Rather than a late Wisconsinan age for most dunes, results indicate that many dunes may have initially formed during the Holocene Altithermal (9,000–6,000 yr B.P.) but have been destabilized episodically since. Periods of stability have often been brief, resulting in soils with A/AC/C profiles. In addition, instability appears to have varied spatially in dune fields, with some places active and others not. Prevailing winds were northwesterly in the Sand Hills and in Colorado and southwesterly on the southern High Plains.

Significance of Previous Research to this Study

The review of relevant literature revealed several research questions, specifically germaine to the Great Bend Sand Prairie, which were the hypothetical foundation of this study. With regard to the climate history of the region, it was believed that evidence would show a change from cool and moist during the late Wisconsinan to relatively warm and dry in the Holocene. From a stratigraphic perspective, Rosner (1988) recognized a near surface silt-clay layer that, she claimed, underlies most dunes in the study area. Johnson (1991) referred to the unit as a silt layer that, he argued, is probably Peoria loess and therefore late Wisconsinan in age. Given the overall lack of data regarding this relatively fine-textured deposit, one goal of this study was to test the following hypotheses: (1) the deposit is indeed commonly occurring, (2) the unit provides a maximum-limiting age for dune development, and (3) the deposit is Peoria loess.

Several questions also arose that focus on the age of dunes on the Great Bend Sand Prairie, the chronology of eolian activity, and the variables that caused eolian sand to initially mobilize and subsequently remobilize. Numerous studies (e.g., Antevs, 1955; Gile, 1979; Ahlbrandt et al., 1983; Muhs, 1985; COHMAP Members, 1988; Holliday, 1989) indicate that a climate shift from cool conditions with more effective moisture to a warm, semiarid and subhumid climatic regime occurred at the Pleistocene-Holocene boundary on the Great Plains. This shift in climate is thought to have favored dune formation in northeastern Colorado (Muhs, 1985), Nebraska (Ahlbrandt et al., 1983), and in the southern High Plains (Holliday, 1985, 1989). Moreover, the data clearly indicate that dunes have episodically mobilized throughout the region. Given these findings, the ultimate goal of this study was to test the following hypotheses: (1) dune fields on the Great Bend Sand Prairie are generally Holocene landforms, (2) mobilization of eolian sand occurred because the climate shifted from cool and relatively moist in the late Wisconsinan to semiarid and subhumid during the Holocene, and (3) Holocene mobilization of sand has been episodic and patchy.

Research Methods

Data Requirements

Given the lack of geomorphic information that previously existed from the Great Bend Sand Prairie, data were collected in this study using standard field and laboratory methods. Several sources of existing data proved to be useful in the early stages of the study, including bulletins and reports of the Kansas Geological Survey and soil surveys published by the U.S. Department of Agriculture Soil Conservation Service (SCS). In addition, landscape data were derived from U.S. Geological Survey 7.5-minute topographic maps and 1:24,000-scale stereographic aerial photography supplied by the Kansas Applied Remote Sensing Program (KARS).

Field and Laboratory Methods

The field investigation began with exploration of the study area. During this phase, sand-sheet landforms were categorized and mapped according to McKee (1979) through a combination of aerial photograph analysis and field survey. A major goal of the reconnaissance was to better understand the spatial distribution of Rosner's (1988) silt-clay layer and Johnson's silt layer (1991). Accordingly, sub-dune stratigraphy was explored at 126 widely scattered and randomly selected sites throughout the region with a Giddings coring machine and hand bucket auger. Test depth varied considerably, ranging from about 2.5 m (8 ft) to around 30 cm (1 ft), depending upon the depth of the deposit. Given the inherent imprecision associated with bucket augering, stratigraphy and texture were only generally described. Basic characteristics (e.g., texture, color, structure) were noted qualitatively, with the primary goals being to determine whether the deposit was present at a given site, to ascertain the thickness of overlying wind-blown deposits, and to select sites for more intensive stratigraphic study.

Based upon the exploratory results, 24 sites were selected and investigated in more detail. Of those, five sites were quarries, roadcuts, or stream cutbanks where radiocarbon ages were derived but no sampling for chemical and physical attributes of sediments occurred. Nineteen sites were selected for detailed stratigraphic investigations where backhoe trenches were subsequently excavated, either to provide a fresh exposure or to accentuate an already existing cutbank, quarry, or drainage trench. At each of these sites, pedostratigraphic units were differentiated according to the North American Stratigraphic Code (The North American Commission on Stratigraphic Nomenclature, 1983) and pedologic horizons were described according to Soil Conservation Service standards (Soil Survey Staff, 1987).

Various laboratory procedures were employed to characterize the physical and chemical composition of pedostratigraphic units. Differences in sediment texture were quantified by the pipette method (Day, 1965). In order to graphically plot textural results, Krumbein's (1934) logarithmic transformation (phi scale) of the Udden-Wentworth (Wentworth, 1922) grade scale was used. Subsequent graphical statistics, including the mean, median, sorting, skewness, and kurtosis (as defined by Folk and Ward, 1957) were calculated with software by Prante (1989). To better characterize the depositional environments in which the silty sand and dune sand accumulated, scatterplot analyses of textural variables (e.g., Folk and Ward, 1957; Friedman, 1967) were conducted on a total of 140 samples that were collected from the twelve sites at which both the silty sand and dune sand were well expressed.

Chemical characterization of late Quaternary sediments focused on calcium carbonate and organic matter analyses. Carbonate content in sand dunes was expected to be minimal, except where buried soils were present, because eolian sands are relatively porous and very well drained. In contrast, carbonate content in the silty deposits was expected to be relatively high, owing to low infiltratibility and the accumulation of illuvial calcium at an unknown depth in a cool, relatively moist climate. Accordingly, the Chittick method (Dreimanis, 1962) was used to quantify calcium-carbonate concentrations. Similarly, the Walkley-Black method (Allison, 1965) was used to determine the organic-matter content in sediments, to better define soil A horizons, and to test for ^{14}C-dating potential.

The chronology of deposition and stability was reconstructed from radiocarbon ages obtained from the total humate fraction of bulk samples ($\geq$4 kg; 9 lb) collected from the upper and lower 5 cm (2 in) of buried A horizons. Radiocarbon ages from the lower parts of buried A horizons provide the closest minimum-limiting ages for the host deposit, whereas those from the upper parts of buried A horizons give the closest maximum-limiting ages for overlying units. Rootlets and other detrital plant material were removed by flotation, the sand fraction was separated by decantation, and carbonates were eliminated by treatment with hydrochloric acid. Samples were subsequently oven dried, pulverized, and sent to the University of Texas (Austin) for age determination. In order to provide conservative radiocarbon estimates, all ages were calculated at 2σ, corrected for isotopic (δ^{13}C) fractionation (Stuiver and Polach, 1977) and calibrated to the tree-ring curve (Stuiver and Reimer, 1993).

In an effort to fully characterize paleoenvironmental conditions in the region, faunal and floral remains were collected at three sites and were subsequently identified in The University of Kansas soil laboratory. Additional paleofloral evidence was provided from $\delta^{13}C$ values obtained from buried soils in the silty deposits and dune sand. It has been established that $\delta^{13}C$ values can be used to infer vegetation and associated paleoclimate (e.g., Krishnamurthy et al., 1982; Delaune, 1986) at the time of soil formation or deposition because many warm-climate (C_4) prairie grasses (e.g., *Andropogon gerardii, Panicum virgatum*) typically have mean $\delta^{13}C$ values of -12‰ whereas cool-climate (C_3) prairie grasses (e.g., *Elymus canadensis, Agropyron smithii*) have average $\delta^{13}C$ values of -27‰ (Deines, 1980; Krishnamurthy et al., 1982; Cerling and Quade, 1993; Nordt et al., 1994). Buried soils in silty deposits, for example, were expected to have more negative $\delta^{13}C$ values than those obtained from buried soils in dune sand, reflecting a climate shift from cool and relatively moist to warmer and semiarid to subhumid.

Results

Reconnaissance

As noted above, coring, bucket augering (or both) occurred at 126 sites in order to ascertain the distribution of the silt-clay/silt layer. Of the 126 sites tested, 95 (75.4%) contained the silt layer, indicating that it is indeed widespread (fig. 6). Over the course of the reconnaissance, however, it was determined that the terms *silt-clay layer* or *silt layer,* which imply a single stratum largely composed of silt, are misnomers. In reality, it appears that the deposit, as previously defined (Rosner, 1988; Johnson, 1991), may include several facies, differentiated by color and texture, that intertongue in some fashion. Moreover, the deposit consists largely of sand at most sites, but does contain high, though variable, percentages of silt and clay as compared to the overlying, wind-blown sand. At all localites, well-developed soils with strong structure were recognized in the deposit. Given that a generic term for the unit(s) is useful when referring to site stratigraphy, a new term, *silty sand,* is proposed to reflect the dominant composition of the deposit. Qualitative results from exploratory testing indicated that the silty sand changes considerably in depth, texture, and color in the study area. At some localites, the unit is buried by several meters of eolian sand, whereas at others, the stratum was exposed at the surface. Texture varies from silty clay loam to sandy clay loam, and color ranges from black (10YR2/1; moist) to an oxidized, strong brown (7.5YR5/6; moist). As anticipated, the silty sand is mottled or gleyed in many places due to reduction.

Mapping

Mapping of surficial deposits was conducted by analyzing aerial photography, county soil surveys, and reconnaissance. Ultimately, a map was produced that recognizes six, primary geomorphic categories in the uplands of the Great Bend Sand Prairie (fig. 6):

1) Loess plain, including flat to slightly undulating topography underlain by loess in the area north and east of Stafford.

2) Low-relief sand sheet, including landscapes of little or no relief that are mantled by sand.

3) High-relief sand sheet, consisting of slightly undulating, irregular sand-mantled topography without steep faces.

4) Compound subparabolic dunes, including some elements of category 3, but consisting largely of dune fields displaying traces or remnants of parabolic dune limbs and steep slip faces superimposed on one another.

5) Compound parabolic dunes, dune fields that consist largely of superimposed parabolic forms, but may include elements of categories 3, 4, and 6.

6) Parabolic dunes, dune fields composed of individual parabolic forms with steep slip faces and well-defined limbs.

Stratigraphic Investigations

Nineteen sites showing the variability of late Quaternary stratigraphy on the Great Bend Sand Prairie were exposed or accentuated by backhoe trenching and systematically described, sampled, and analyzed according to the methods outlined above. The results of intensive stratigraphic investigations at each of the 19 study sites are summarized below (for details see Arbogast, 1995). Although the top of silty sand serves as a stratigraphic datum because it is generally a traceable surface (fig. 7), unit designations for deposits at each site are unique and do not imply any correlation among or between sites.

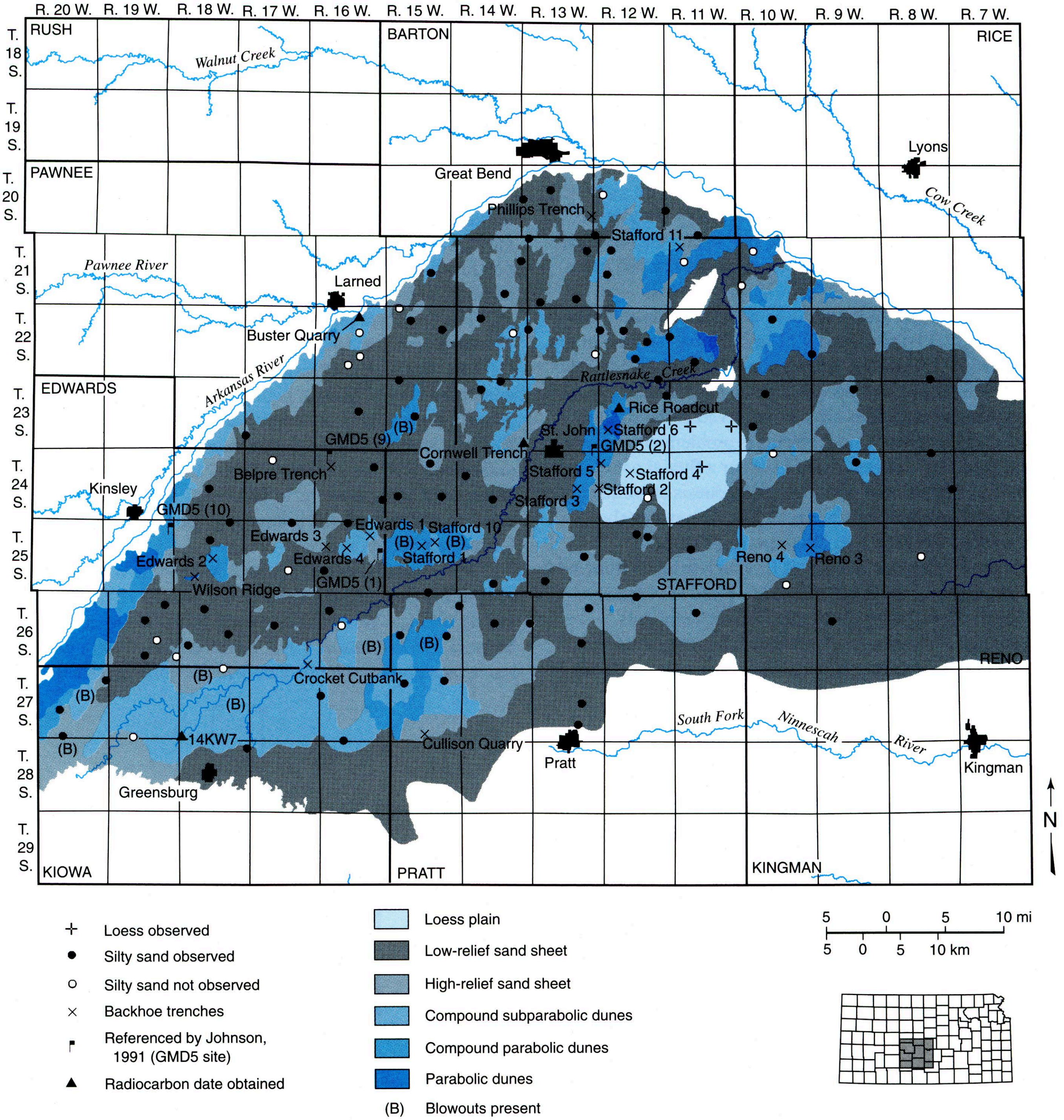


FIGURE 6—Generalized landform classes and study sites on the Great Bend Sand Prairie.

Belpre Trench

The Belpre Trench is a 4.11-m (13.5-ft)-deep section (figs. 7, 8) in a drainage pit excavated on a level sand sheet in the NE NW sec. 8, T. 24 S., R. 15 W. (fig. 6). After the trench wall was cleaned, three pedostratigraphic units were recognized in the exposure (figs. 8, 9). In general, deposits are loamy, moderately to very poorly sorted, finely to very finely skewed, and very leptokurtic to very platykurtic (Arbogast, 1995). Three radiocarbon ages were obtained from the site, indicating a geomorphic history that spans the past 20,000 years.

Unit III, the lowest exposed at the site, ranges from 3.74 m (12.3 ft) to at least 4.11 m (13.5 ft), the base of the profile. In general, the deposit consists of relatively well sorted, oxidized, and unconsolidated sand. Because sedimentary structures are not preserved in the stratum, determining a depositional facies is problematic, i.e., the unit could have accumulated in either a fluvial or eolian environment. A brief period of pedogenesis subsequently occurred, resulting in the formation of a moderately developed soil with two 3Bwb horizons (fig. 9).

Overlying Unit III is Unit II, which extends from 3.44 m (11.3 ft) to 3.48 m (11.4 ft) (figs. 8, 9). In contrast to Unit III, Unit II consists of much siltier sediments. Although sedimentary structures are not preserved, the deposit is very poorly sorted, suggesting a fluvial facies in a presumably more moist environment than at present. Following deposition, a period of pedogenesis occurred, one that promoted the formation of a well-developed soil with two 2Btgb horizons (fig. 9). The upper part of the solum dated to approximately 20,000 yr B.P., suggesting a Farmdalian, or perhaps Woodfordian age for both Units III and II. A δ^{13}C value of -22.3 implies that a mixture of plants with C_3 and C_4 pathways dominated at the site during that time. The soil is mottled, with pockets of gleying and iron oxidation present, which suggests ponding during some unknown interval of time or, alternatively, a periodically high water table.

The dominant pedostratigraphic unit at the Belpre Trench is Unit I, extending from the surface to 3.44 m (11.3 ft) (figs. 8, 9). Although the deposit is largely composed of silt (72.6%), high percentages of clay are also present, especially in the upper part (41.9% at 1.30 m; 4.3 ft) of the stratum (fig. 9). Sedimentary structures are not preserved in the unit, so depositional facies can not be determined with certainty. In fact, conflicting evidence exists regarding both the sedimentary process responsible for the unit and its age.

The color of the lower part (2.06–3.44 m; 6.8–11.3 ft) of Unit I is generally consistent (pale brown; 10YR6/3; moist) with late Wisconsinan Peoria loess (Wells and Stewart, 1987; Johnson, 1993; Johnson et al., 1993), which has been recognized immediately to the north and south of

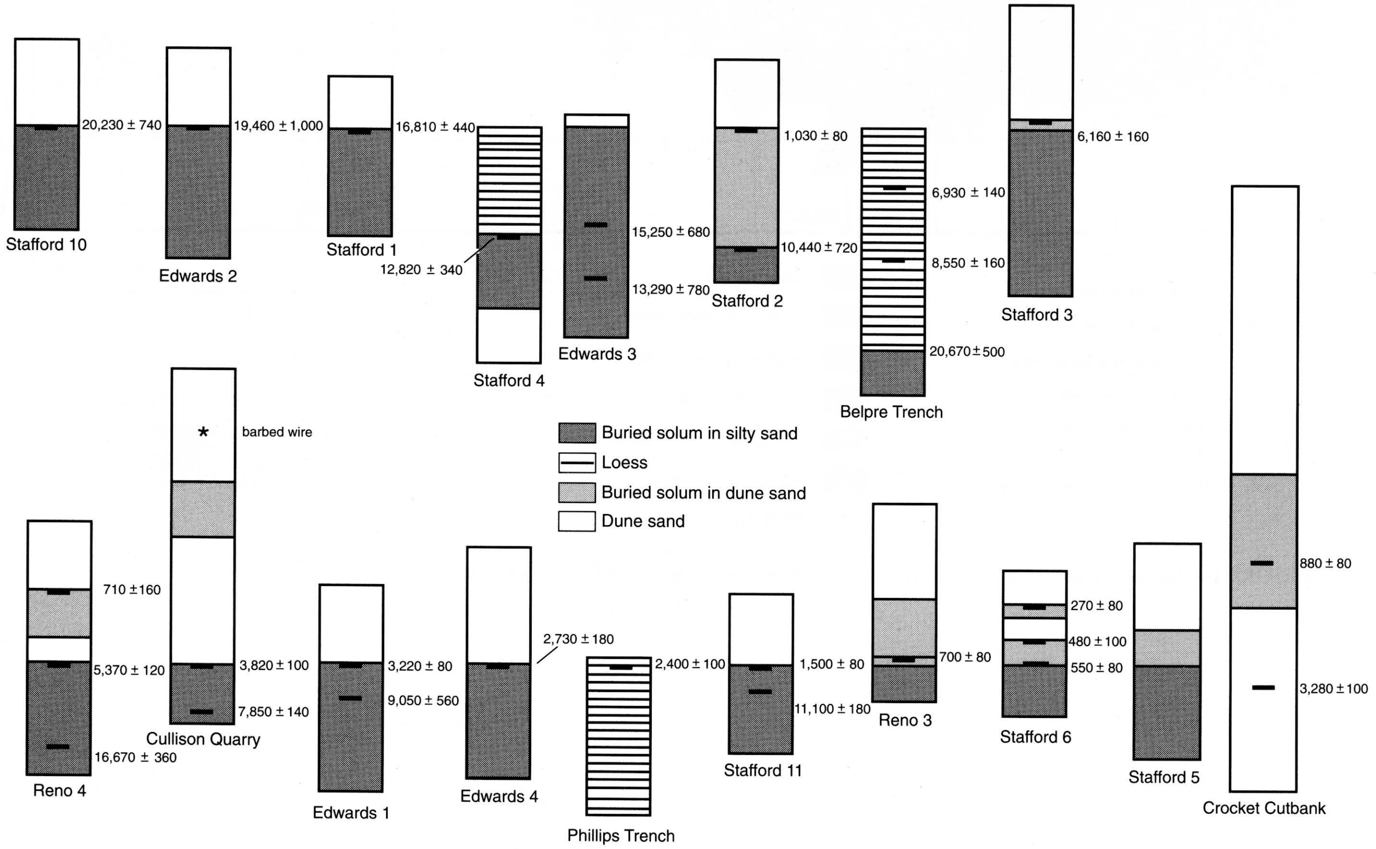


FIGURE 7—SUMMARY STRATIGRAPHIC DIAGRAM FROM THE 18 SITES where silty sand and/or loess was investigated on the Great Bend Sand Prairie (modified from Arbogast, 1996b).

the Great Bend Sand Prairie (Feng, 1991; Feng et al., 1994). Moreover, abundant late Wisconsinan gastropods, including *Succinea avara, Lymnaea parva,* and *Helicodiscus singleyanus* (fig. 10) were recovered from the base of the deposit. On first examination, therefore, a Woodfordian age was suggested for deposition of Unit I. Approximately 20 cm (7.9 ft) above the snail zone, however, a radiocarbon age of approximately 8,500 yr B.P. was obtained, implying a very early Holocene age for the deposit. Apparently, the sample that was dated must have been contaminated in some way, possibly by Holocene organic carbon illuviated during the extensive soil development which has occurred. Given the age (about 20,000 yr B.P.) derived from the top of Unit II, the presence of gastropods, and the similarity with Peoria loess, a Woodfordian age for the lower part of Unit I is more logical.

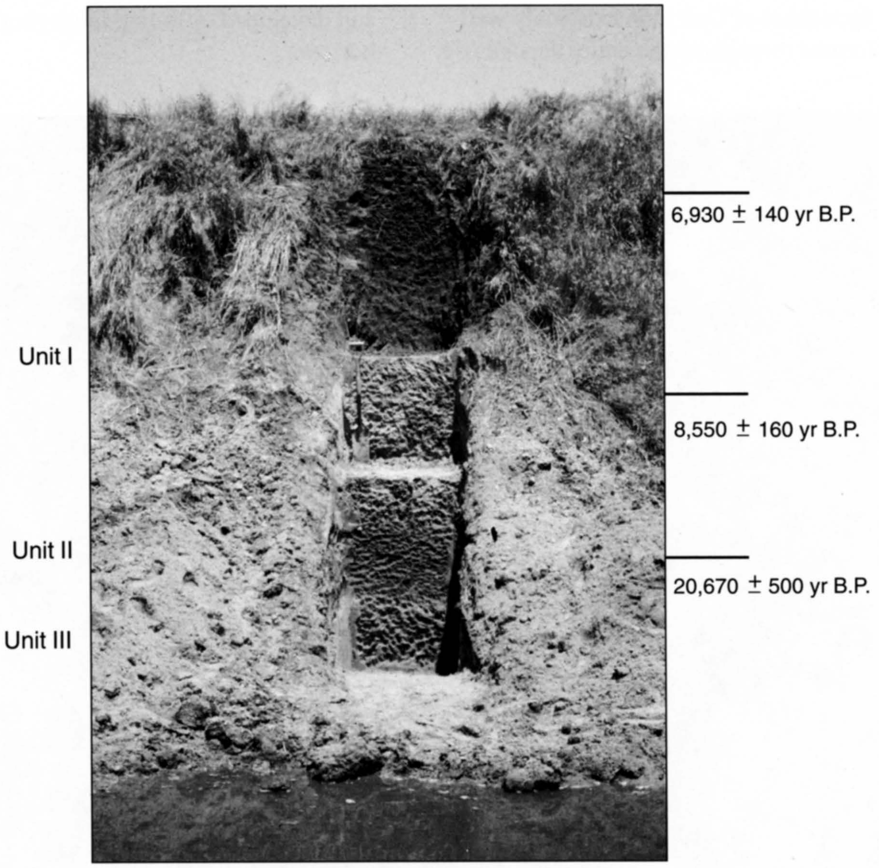


FIGURE 8—The 4.11-m (13.5-ft)-deep Belpre Trench, showing the position of pedostratigraphic units and radiocarbon ages.

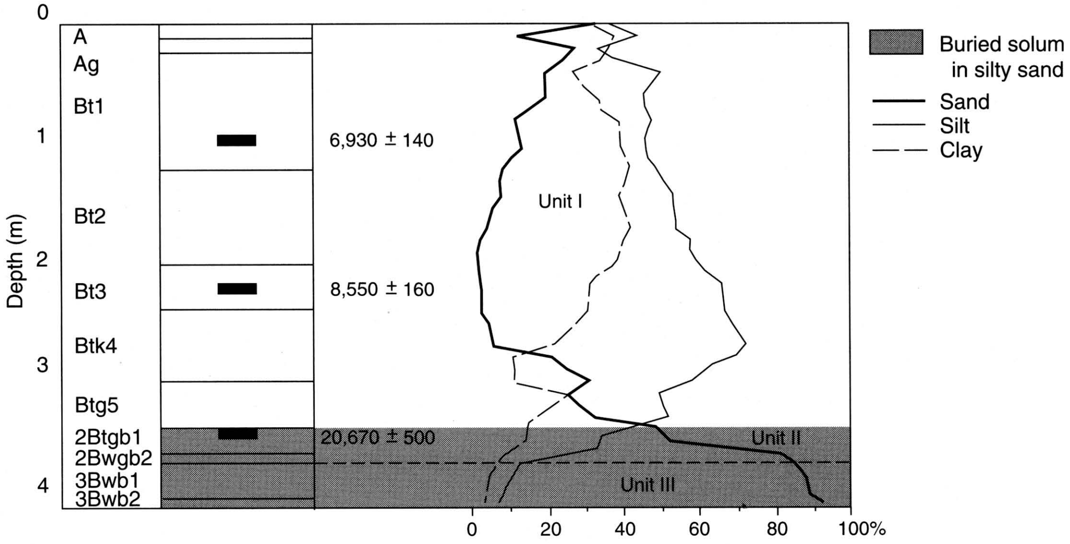


FIGURE 9—Pedostratigraphy, soil horizonation, and texture at the Belpre Trench.

In addition, an age of about 17,000 yr B.P. was obtained by Johnson (1991) from spruce (*Picea* cf. *glauca*) charcoal in a similar stratigraphic position at GMD 5, site 9, approximately 0.5 km (0.2 mi) to the northwest, further suggesting a Woodfordian age for Unit I. A $\delta^{13}C$ value of -21.0‰ derived from the deposit, coupled with the oxidized nature and presence of aquatic gastropods in the lower part of the unit, indicates relatively moist conditions during sedimentation.

Following deposition of Unit I, an extremely well developed soil formed throughout the entire deposit (fig. 9). Structure is very strong in the solum, especially in the upper 1.50 m (4.9 ft), where it is coarse prismatic, parting to moderate blocky. Overall, the character of the soil suggests pedogenesis over an extended interval of time, perhaps throughout the late Holocene. In fact, the upper part of the Bt1 horizon, approximately 1.0 m (3 ft) below the surface, dated to approximately 6,000 yr B.P. A $\delta^{13}C$ value of -20.2‰ suggests plants with a C_3 pathway were present at that time. Apparently, water has ponded periodically at the site as the soil developed, resulting in gleying of the lower part of the A horizon.

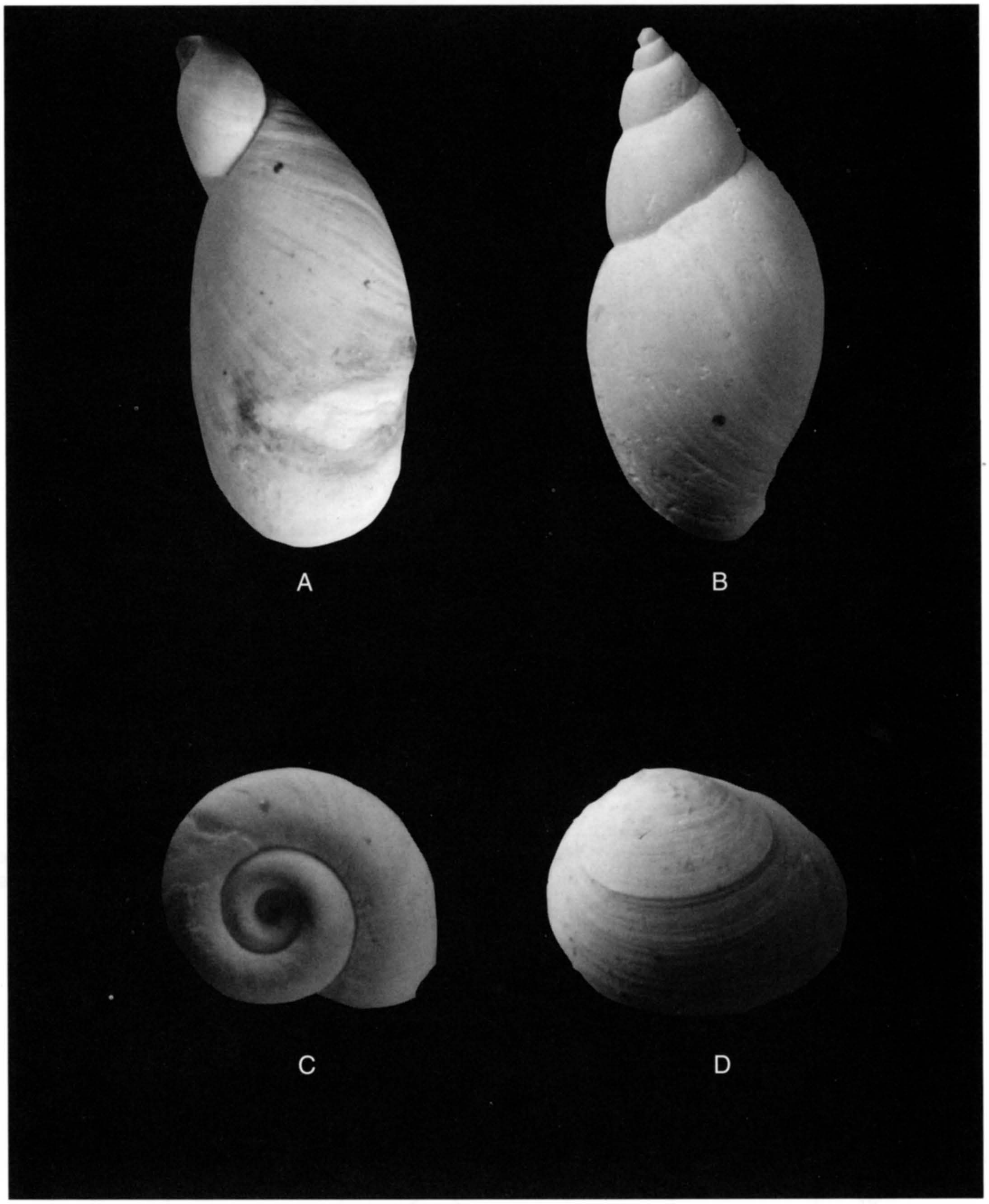


FIGURE 10—Mollusks recovered from the Belpre Trench (magnification ×10): (A) *Succinea avara,* (B) *Lymnea parva,* (C) *Helicodiscus syngleyanus,* (D) unidentified bivalve.

Crocket Cutbank

The Crocket Cutbank is an 8-m (26-ft)-high section (figs. 7, 11) located in a compound subparabolic dune field along a cutbank of Rattlesnake Creek in the SE SE sec. 35, T. 26 S., R. 17 W. (fig. 6). Through backhoe trenching, the overall height of the exposure was increased to 9.8 m (32.2 ft). Four pedostratigraphic units were described in the profile, with two weakly developed and narrowly separated buried soils featured in the exposure (figs. 11, 12). Sediments at the site are generally noncalcareous, sandy (with a mean particle size of fine to very fine sand), poorly to moderately sorted, finely to very finely skewed, and mesokurtic to extremely leptokurtic (Arbogast, 1995).

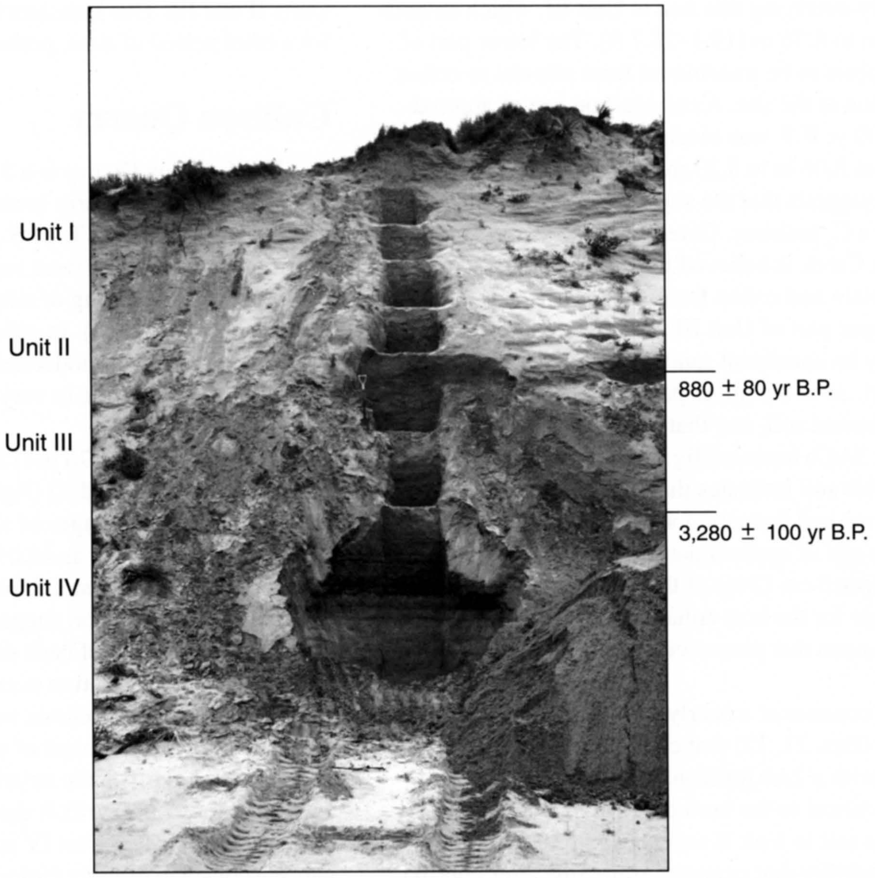


FIGURE 11—The 9.8-m (32.2-ft)-high Crocket Cutbank showing the position of pedostratigraphic units and radiocarbon ages.

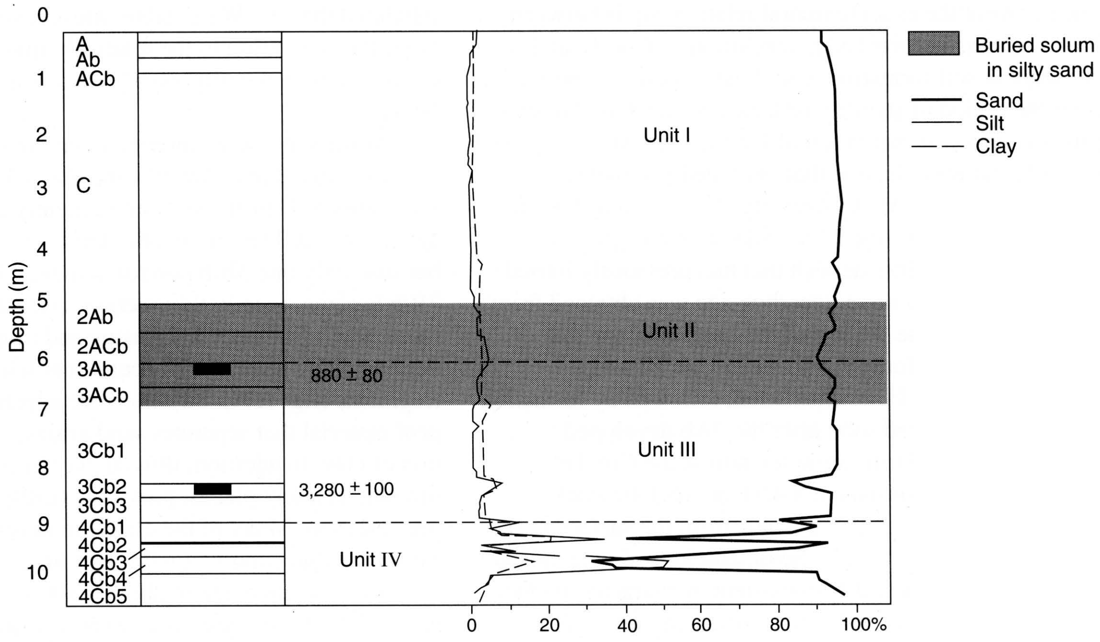


FIGURE 12—Pedostratigraphy, soil horizonation, and texture at the Crocket Cutbank.

The lowermost deposit is Unit IV; it extends from 8.76 m (28.7 ft) to the base of the profile and is a stratified alluvial deposit probably associated with Rattlesnake Creek. Although no radicarbon ages were obtained from the unit, the lack of cementation and the floodplain position relative to Rattlesnake Creek suggest a late Holocene age.

Directly overlying this unit is Unit III, which extends from 5.88 m to 8.76 m (19.3–28.7 ft). The lower part of Unit III appears to be transitional from alluvial to eolian sedimentation at the site. A radiocarbon age of approximately 3,300 yr B.P. was obtained from a silty deposit that extends from 8.06 m to 8.23 m (26.4–27.0 ft). A $\delta^{13}C$ value of -17.7‰ suggests that the site was largely populated by plants with a C_4 pathway. Given its close proximity, Rattlesnake Creek is believed to be the primary source for both floodplain and eolian facies in the unit.

The upper part of Unit III is entirely eolian, as indicated by its consistent sandy texture and moderate sorting. Formed in the upper part of the unit is a weakly developed buried soil, one that consists of a 3Ab horizon overlying a 3ACb horizon (fig. 12). The degree of development in this soil indicates that landscape stability and accompanying soil formation was relatively brief. A radiocarbon age of approximately 900 yr B.P. on humates from the upper 5 cm (2 in) of the soil provides a late Holocene age for the host eolian sand. A $\delta^{13}C$ value of -16.6‰ suggests that plants with a C_4 pathway dominated at the site.

Unit II consists of a nearly 1-m (3-ft)-thick deposit of eolian sand (figs. 11, 12) that contains a weakly developed buried soil with a 2Ab horizon and a 2ACb horizon. Virtually identical to the buried soil at the top of Unit III (fig. 12), the soil in Unit II reflects another, brief period of landscape stability that occurred sometime after approximately 900 yr B.P. Despite their close proximity, it is unclear what the exact temporal relationship is between the two soils. Theoretically, deposition of Unit II, and subsequent soil formation, could have occurred shortly after the period of stability reflected by the 3Ab. Given this scenario, sedimentation of Unit II suggests a very brief depositional episode, one that exceeded pedogenic thresholds in the 3Ab. Conversely, deposition of Unit II and development of the 2Ab could have transpired after truncation of an older deposit that had previously buried the 3Ab. Although no radiocarbon age was obtained from the 2Ab to estimate the age of the host sand, the lack of any visible unconformities between Unit III and II suggests that Unit II sedimentation, followed by formation of the 2Ab, occurred soon after the 3Ab developed.

The uppermost stratigraphic unit at the Crocket Cutbank, Unit I, consists of a 4.91-m (16.1-ft)-thick deposit of eolian sand that extends from the surface to the top of Unit II (figs. 11, 12). Sedimentological data indicate that the unit is remarkably consistent in character, including deposits that are moderately sorted and as much as 97% sand (fig. 12). This evidence, coupled with the lack of any visibile unconformities, suggests that eolian sedimentation at the site has been relatively consistent in the past few hundred years. Given the availability of fluvial sediments, Rattlesnake Creek may have been the primary source for eolian sands at the site. Recently, a poorly developed surface soil (A/AC/C horizonation) has formed, one that is consistent in its development with the buried soils recognized in Units II and III. This indicates that the dune had been stable for a brief period of time, perhaps less than 500 years.

Cullison Quarry

The Cullison Quarry is a 5.95 m (19.5 ft) section (figs. 7, 13) exposed in a quarry located in a high-relief sand sheet in the NE SW sec. 34., T. 27 S., R. 15 W. (fig. 6). Four pedostratigraphic units were recognized in the profile, with the lower pair consisting of silty sand and the upper two of dune sand (figs. 13, 14). In general, units are loamy to sandy, very poorly to moderately sorted, very finely to finely skewed, and platykurtic to very leptokurtic (Arbogast, 1995).

Extending from 4.70 m (15.4 ft) to the base of the profile are Units IV and III (figs. 13, 14), which contain relatively high percentages of silt and clay. Unit IV, in particular, has as much as 46.0% silt and 26.5% clay. Although sedimentary structures were not observed in either unit, sorting is very poor, suggesting a fluvial facies. Following deposition of each deposit, a period of landscape stability and soil formation occurred, resulting in well-developed soils. Pedogenesis was especially intense in Unit IV, promoting development of two 4Btgssb horizons (fig. 14) with strong, prismatic structures. A radiocarbon age of approximately 7,900 yr B.P. derived from the 4Btgssb1 horizon suggests that Unit IV accumulated between the very late Wisconsinan or early Holocene. A $\delta^{13}C$ value of -20.0‰ indicates a mixture of plants with C_3 and C_4 pathways inhabited the site. Water tables must have been periodically high, for in addition to the relatively high percentage (5.2%) of carbonate in the 4Btkgssb2 horizon, all of Unit IV is heavily gleyed.

At some point, an interval of erosion apparently occurred, one that removed a theorized 4Ab. The sandier sediments of Unit III were subsequently deposited and a soil again formed. The original thickness of Unit III is unknown, because only one 3Btb horizon remains. The matrix of the 3Btb horizon, seen in thin section, consists of a skeleton fabric and a fine-textured plasma and mineralogy that is dominated by quartz, with occasional feldspars and rock fragments (fig. 15A). Striations are apparent in the birefringent material that separates sand grains, indicating illuviation of clay. In addition, illuvial clay lines the internal wall of voids in many places, providing further evidence of clay translocation. Many of the clays are twisted, suggesting their expansion and contraction through time.

A radiocarbon age of about 3,800 yr B.P. from the upper part of Unit III suggests final exposure during the late Holocene when higher temperatures and probably less

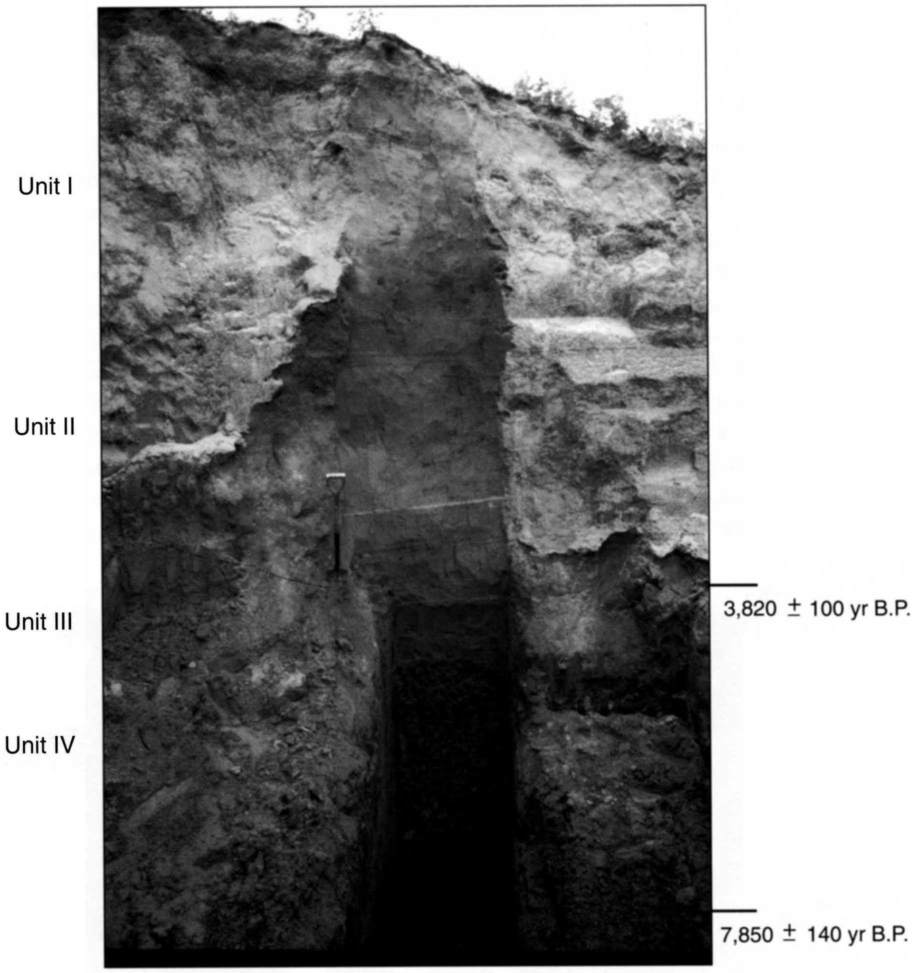


FIGURE 13—THE 5.95-M (19.5-FT)-HIGH CULLISON QUARRY, showing the position of pedostratigraphic units and radiocarbon ages.

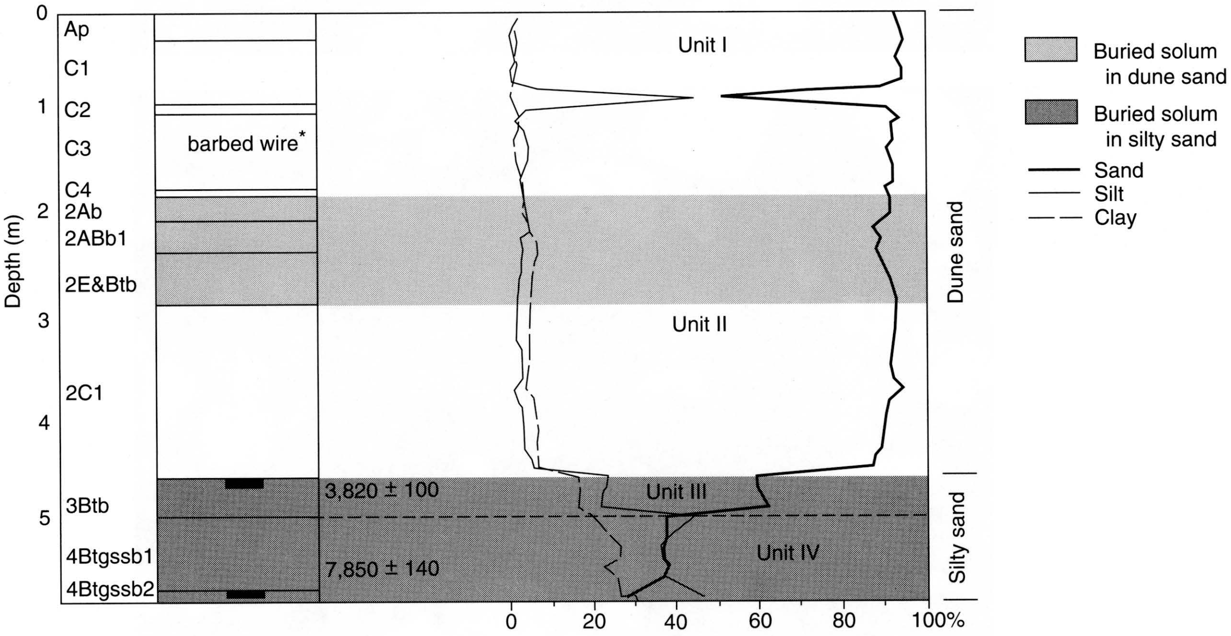


FIGURE 14—PEDOSTRATIGRAPHY, SOIL HORIZONATION, AND TEXTURE at the Cullison Quarry.

effective moisture prevailed, as indicated by a $\delta^{13}C$ value of -14.8‰. In short, the character of Units IV and III, coupled with the ages derived from them, are generally consistent with other deposits (e.g., Edwards 1–3, Stafford 3, Reno 4, for which descriptions follow) categorized as silty sand on the Great Bend Sand Prairie.

Unconformably overlying the silty sand are a pair of units, Units II and I, that are consistent in texture, structure, color, and topographic position and expression with deposits categorized as dune sand in the region. Each unit represents a period of sand accumulation, presumably during a more arid climate interval, followed by relative stability and soil formation when more effective moisture was present. A radiocarbon age of approximately 3,800 yr B.P. from the top of Unit III (figs. 13, 14) suggests that Units II and I are late Holocene deposits. Although the 2Ab was not dated, its position and degree of development (fig. 14) is consistent with buried sola in the region (e.g., Crocket Cutbank, Reno 3, Reno 4), from which ages less than 1,000 years have been derived. The matrix of the soil, seen in thin section, further indicates very little development because it consists largely of skeleton grains that lack plasma. Mineralogy is dominated by quartz with scattered grains of feldspar and miscellaneous rock fragments (fig. 15B). In contrast to the thin section of the 3Btb1 horizon (fig. 15A), the 2Ab shows very little indication of post-depositional alteration. Some sand grains are coated with birefringent material, probably clay, that may reflect illuviation, but this material may have also been present on the grain as it was transported. Either way, very little clay translocation has occurred relative to the 3Btb1 (fig. 15B).

Overlying Unit II is Unit I (figs. 13, 14), which, as indicated by a small piece of barbed wire recovered from the center of the deposit, is probably less than 200 years

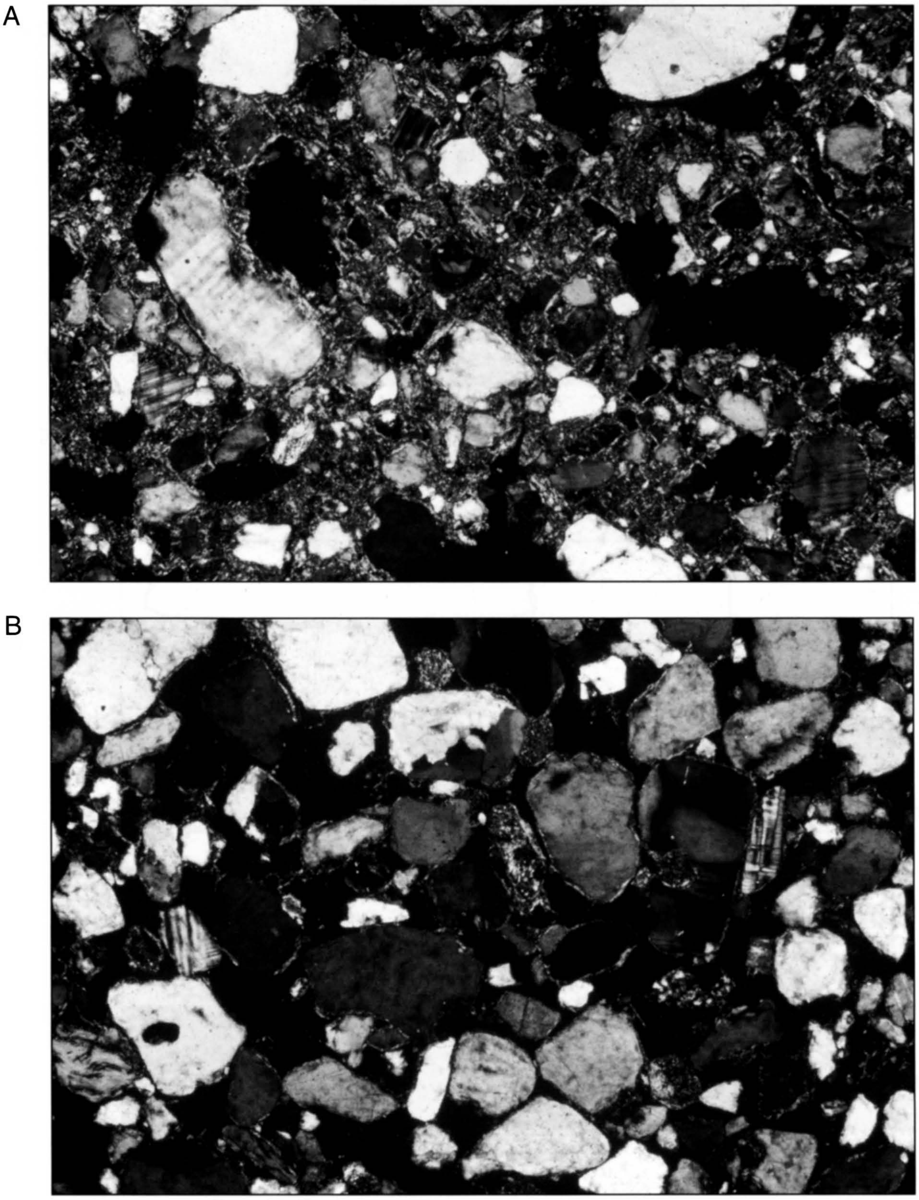


FIGURE 15—THIN SECTIONS (FOV = 1.85 × 2.69 MM) of the (A) 3Btb1 and (B) 2Ab horizons at the Cullison Quarry (from Arbogast and Johnson, in press).

old. A weakly developed surface soil, consisting of an Ap horizon and C horizons, coupled with the barbed wire found beneath it (fig. 14), strongly implies that the dune has been stable for a very brief period of time, probably around 200 years.

Edwards 1

Edwards 1 is a 3.30-m (10.8-ft)-deep section (figs. 7, 16), exposed by backhoe trenching, located in a compound subparabolic dune field in the SW NE sec. 11, T. 25 S., R. 16 W. (fig. 6). Three pedostratigraphic units—consisting of a basal silty sand, an intermediate stratum of sandier sediments, and a surficial eolian deposit—were described (figs. 16, 17). In each of the lower two units, a well-developed buried soil was recognized, but none was observed in the overlying eolian sand (fig. 17). Sediments in the silty sand are noncalcareous, loamy, very poorly sorted, very finely to finely skewed, and mesokurtic to very leptokurtic. Overlying eolian deposits are noncalcareous, very sandy, poorly sorted, very finely to finely skewed, and very leptokurtic to leptokurtic (Arbogast, 1995).

At the base of the profile is Unit III, which contains a consistently high percentage of silt (about 60%) and clay (about 25%) and is very poorly sorted, suggesting that it accumulated in a low-energy environment, possibly fluvial or lacustrine. Contained within the unit is an extremely well developed buried soil, consisting of stacked Btb horizons (fig. 17) indicating that the surface was exposed for long periods of time. The lowermost horizon is gleyed, resulting from periodically high water tables or long-term infiltration of water from the surface. A radicarbon age of about 9,000 yr B.P. from humates in the upper 5 cm (2 in) of the 3Btb1 suggests that the underlying deposits are late Wisconsinan or very early Holocene in age. A δ^{13}C value of -22.2‰ indicates the prevalence of C_3 plants, further suggesting a relatively cool environment with probably more effective moisture.

Unit II, which is approximately 50 cm (20 in) thick, is the intermediate deposit at Edwards 1 (figs. 16, 17). The contact between Units III and II is a major unconformity, as suggested by the lack of an Ab horizon in Unit III and a late Holocene age of about 3,200 yr B.P. taken from the top of Unit II. Unit II is much coarser than the underlying deposit, indicating that some change in depositional environment occurred at the site. The increase in sand in Unit II, coupled with an apparent late Holocene age, favors eolian deposition. No change in sorting can be discerned from the underlying strata, however, as relatively high percentages of silt and clay (about 20% and 15%, respectively) persist (fig. 17). Theoretically, Unit II could have accumulated when the surface occupied an interdune position on the landscape, and a combination of eolian and alluvial processes might have contributed to sedimentation. The unit must have been exposed periodically and for some time, as indicated by the strong soil that developed throughout the unit. A radiocarbon age of about 3,200 yrs B.P. (figs. 7, 16, 17) suggests that the last exposure of Unit II was during the late Holocene. There were apparently less effective moisture and warmer temperatures than during the Early Holocene, as implied by a δ^{13}C value of -13.2‰ from the upper part of the stratum.

The uppermost pedostratigraphic unit at Edwards 1 is Unit I (figs. 16, 17). Illustrated on fig. 17 by a sharp textural contact, the deposit consists of an approximately 1.10-m (3.6-ft)-thick deposit of sand that is consistent in texture, color, and topographic position and expression with other deposits in the region identified as dune sand. Following a period of erosion that removed the Ab from the upper part of Unit II, Unit I apparently accumulated during the late Holocene. Although it is unclear when that occurred, two pieces of evidence suggest that the surface of the site has been stable for a relatively long period of time: the degree of surface soil development (including a Bw and two Bt horizons with lamelli) and oxidation in the Bt2 horizon. Given this evidence, especially the extent of surface soil development, it is estimated that Edwards 1 has been relatively stable for at least 1,000 years.

Edwards 2

Edwards 2 is a 3.15-m (10.3-ft)-deep section (figs. 7, 18), exposed by backhoe trenching, located in a compound parabolic dune field in the SE NW sec. 22, T. 26 S., R. 18 W. (fig. 6). Three pedostratigraphic units were described, with two consisting of silty sand and one of surficial, eolian sand (figs. 18, 19). In general, sediments are noncalcareous, loamy, very poorly to poorly sorted, finely to very finely skewed, very leptokurtic to platykurtic, and have a mean particle size of fine to coarse silt (Arbogast, 1995).

Unit III extends from 1.78 m (5.8 ft) to the base of the profile (figs. 18, 19). Texture fines sharply from loamy fine sand at the base to loam at the top (fig. 19). Deposits in the lower part are consistent in texture and distribution with surficial wind-blown sands, suggesting an eolian facies. The heavy texture and very poor sorting of the upper portion of Unit III, however, indicates a shift in depositional environment to some form of low-energy fluvial process. Mottling and gleying in the unit indicate periodically high water tables or long-term ponding.

Unit II is characterized by its color (7.5YR4/4; moist), which suggests oxidation, and by a coarsening in texture toward the top of the unit (figs. 18, 19). At the base of Unit II, the sediment is loam, but the percentage of sand increases sharply (77%) at the top of the unit (fig. 19). Sorting is very poor from bottom to top. A radiocarbon age of about 19,000 yr B.P. and a δ^{13}C value of -19.8‰, taken from the upper part of the unit, indicate that underlying deposits, including all of Units III and II, accumulated by the early late Wisconsinan at a time when C_3 plants were abundant at the site.

Although sedimentological evidence in Units III and II generally points to alluvial sedimentation, the degree of soil formation observed in each unit indicates that stable surfaces existed for some time at the site. During intervals of soil formation, intense illuviation occurred that leached carbonate from the lower part of the deposit, and may have contributed to mottling and gleying of Unit III. The lack of buried A horizons suggests that, following stability, erosion occurred, truncating the upper part of the soil. In many places, clay skins can be traced from the top of Unit II to the base of Unit III, indicating that the soils are now welded.

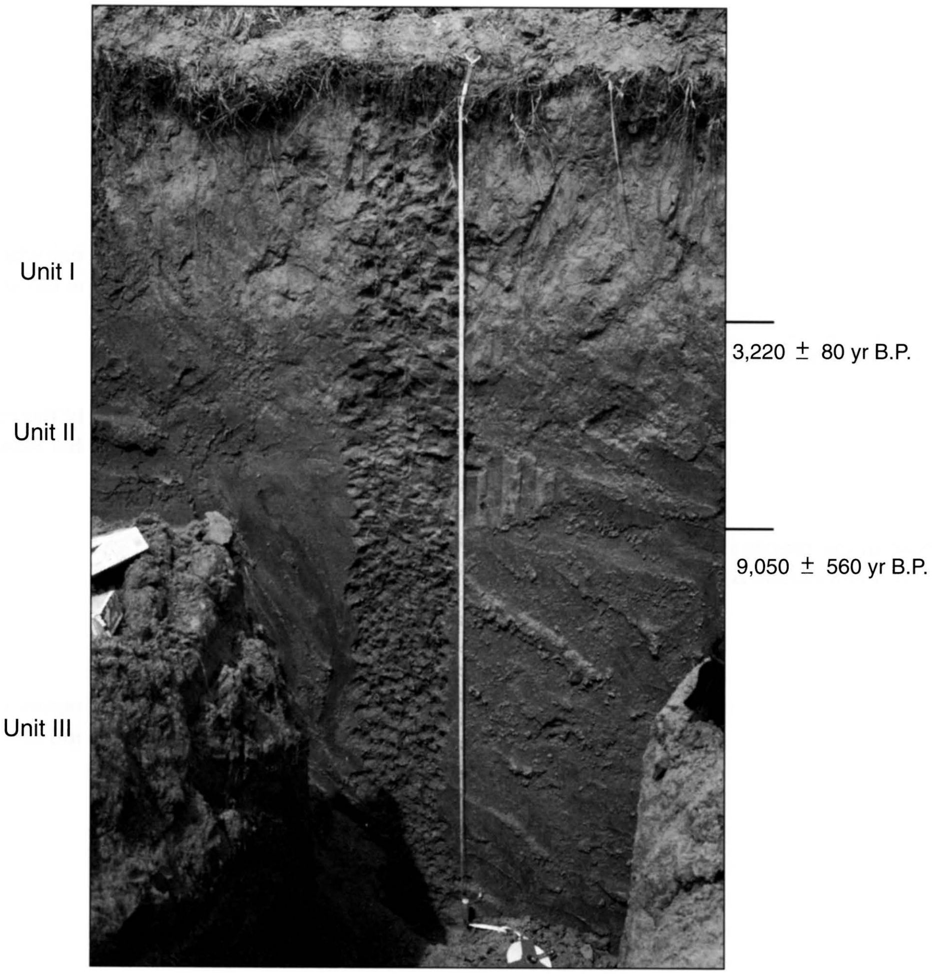


FIGURE 16—The 3.3-m (10.8-ft)-high exposure at Edwards 1, showing the position of pedostratigraphic units and radiocarbon ages.

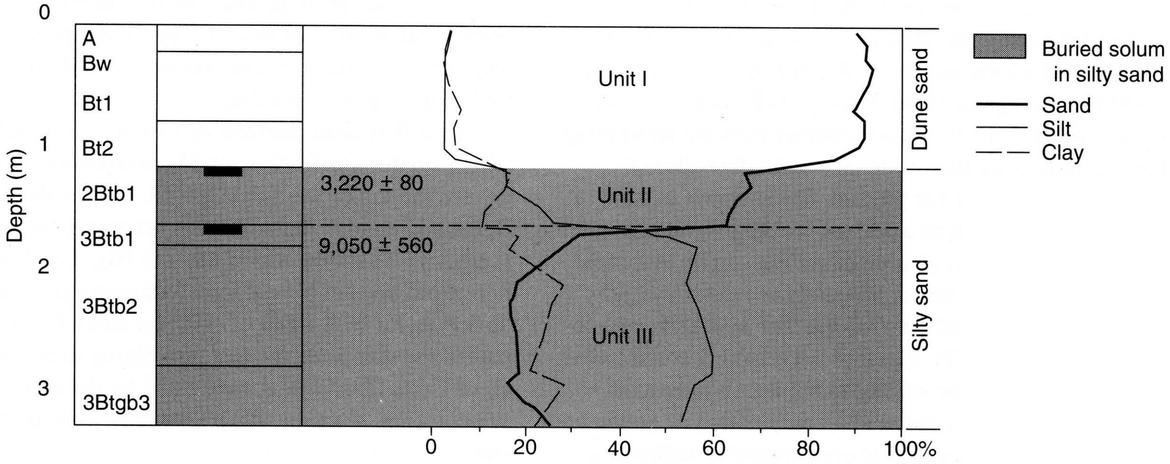


FIGURE 17—Pedostratigraphy, soil horizonation, and texture at Edwards 1.

A sharp contact at 1.23 m (4.0 ft) separates the underlying deposits of Units III and II from the surficial, sandy sediments of Unit I (figs. 18, 19). Containing as much as 88% sand (fig. 19), Unit I is consistent in character (e.g., texture, structure, color, topographic position and expression) with other eolian units observed on the Great Bend Sand Prairie that accumulated in a warm, relatively dry environment. Uncharacteristically, the deposit is very poor to poorly sorted, owing to the relatively large percentage of silt (17%) and clay (11%). Theoretically,

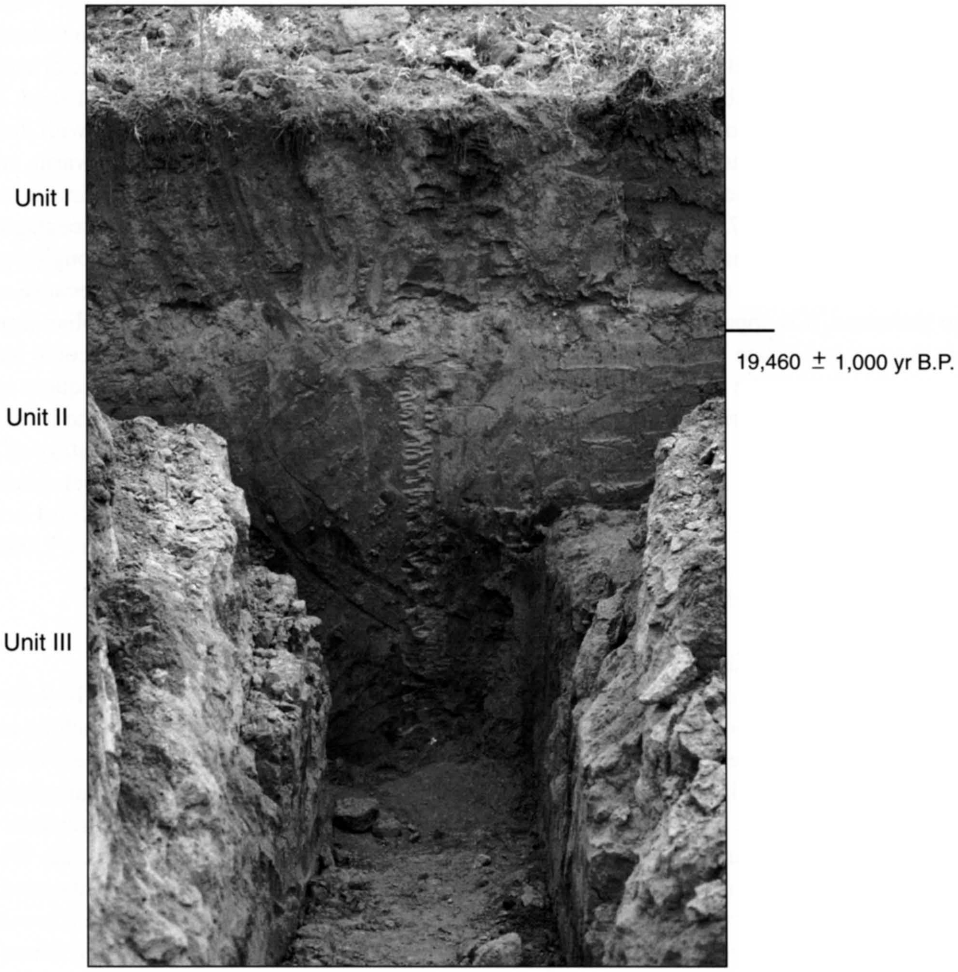


FIGURE 18—The 3.15-m (10.3-ft)-high exposure at Edwards 2, showing the position of pedostratigraphic units and radiocarbon ages.

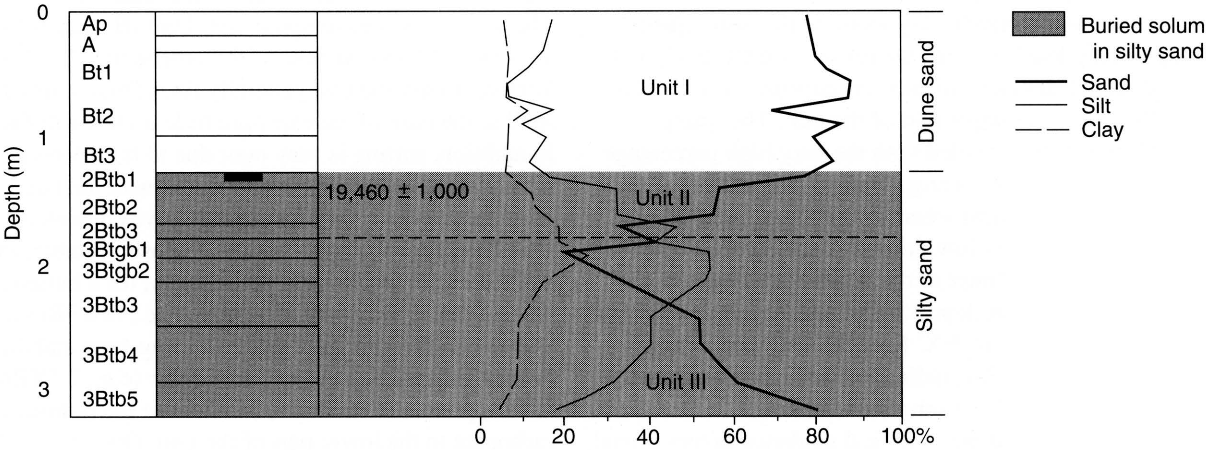


FIGURE 19—Pedostratigraphy, soil horizonation, and texture at Edwards 2.

Unit I may represent both fluvial and eolian facies. The sharp decrease in sorting at 82 cm (32 in), for example, may reflect alluvial deposition in an interdunal environment, with subsequent burial by eolian sand. Alternatively, the local source for Unit I may have contained abundant silt and clay as well as sand.

Evidence indicates that the surface soil has been stable for a relatively long period of time, at least for a deposit of eolian sand on the Great Bend Sand Prairie. Three Bt horizons (fig. 19) were described on the basis of clay films that lined root traces, indicating an extended period of soil formation for a dune in the study area. Most importantly, a radiocarbon age of approximately 19,000 yr B.P. obtained from the top of Unit II (figs. 7, 18, 19) provides a maximum-limiting, late Wisconsinan age for the base of Unit I. Although truncation of a part of Unit I may have occurred during the Holocene, it is apparent that the basal, eolian deposits in the unit are of late Wisconsinan age. Overall, the evidence suggests that the dune at Edwards 2 has been stable for a relatively long period of time.

Edwards 3

Edwards 3 is a 3.3-m (10.8-ft)-deep section (figs. 7, 20) exposed by backhoe trenching on a level sand sheet in the NW NW sec. 17, T. 25 S., R. 16 W. (fig. 6). Four pedostratigraphic units were described, with three consisting of silty sand and one of surficial, eolian sand (figs. 20, 21). In general, deposits are noncalcareous, loamy, with a mean particle size of very fine to coarse silt, are very poorly sorted, finely to very finely skewed, and platykurtic to very leptokurtic (Arbogast, 1995). The sedimentological difference between the silty sand and the overlying sand probably represents a change in depositional environment from low-energy fluvial or lacustrine to eolian.

From 30 cm (12 in) to the base of the profile, the deposits consist of silty sand. Three, distinguishable fining-upward sequences were observed, with each comprising a single pedostratigraphic unit. At the base of the exposure, Unit IV fines from loam at the base to clay-loam at the top of the unit. In the lower part of Unit III, the texture coarsens to sandy clay-loam, with a subsequent fining to clay-loam towards the top of Unit III. In Unit II, the sediment fines dramatically, culminating in 42% clay and 7% sand in the upper part of the unit. The fining-upward sequences, coupled with the very high percentage of clay and very poor sorting, strongly suggest deposition in a fluvial environment where energy fluctuated from relatively high to very low. Although radiocarbon ages from the middle and base of the deposit are inverted, they generally suggest that deposits accumulated during the late Wisconsinan. Values of $\delta^{13}C$ from the silty sands range from -25.6‰ to -24.7‰, indicating that a proportionately high number of C_3 plants inhabited the site.

Although the sedimentological evidence favors fluvial deposition, the formation of three, well-developed buried soils in the silty sand indicates that stable surfaces existed for long periods of time. Each soil consists of stacked Btb horizons with no Ab (fig. 21), suggesting that stability was followed by an erosional episode that truncated the upper part of the solum. In many instances, clay skins can be traced from the top of Unit II to the bottom of the profile, indicating that the soils are now welded.

A sharp contact at 30 cm (12 in) separates the underlying silty sands from the surficial sediments of Unit I (figs. 20, 21). Composed largely of sand, Unit I is consistent in texture, structure, and color with dune sand, deposits that accumulated in a relatively warm, arid environment. Uncharacteristically, the sediment in Unit I is very poorly sorted, owing to the high percentages of silt (ca. 10%) and clay (ca. 10%; fig. 21). Although a precise determination of facies can not be made, because of conflicting evidence, it may have resulted from eolian input of different intensities. Theoretically, the sand could have accumulated initially, followed by a subsequent and infiltrating input of silt and clay. Alternatively, the local source for Unit I may have contained a high percentage of silt and clay that was reworked. The moderate development of the surface soil suggests that deposition of Unit I occurred during the late Holocene, probably in the past 1,000 years.

Edwards 4

Edwards 4 is a 3.31-m (10.9-ft)-deep section (figs. 7, 22) exposed by backhoe trenching in a compound subparabolic dune field in the SW NE sec. 16, T. 25 S., R. 16 W. (fig. 6). Three pedostratigraphic units, consisting of one in silty sand and two in surficial eolian sand, were described in the profile (figs. 22, 23). The sedimentological difference between the silty sand and the overlying sand may record a shift from a low-energy fluvial or lacustrine environment to one where eolian processes dominated in a time of increased aridity. In general, sediments at the site are noncalcerous, loamy, have a mean particle size of medium to coarse silt, are very poorly to poorly sorted, very finely skewed, and leptokurtic to very leptokurtic in their distribution (Arbogast, 1995).

The lowermost stratigraphic unit, Unit III, ranges from 1.58 m (5.2 ft) to the base of the profile (fig. 22, 23). Texture within the unit generally fines from loamy fine sand at the base of the exposure to loam towards the top. In addition, sorting is very poor due to high percentages of silt and clay, suggesting a mixed sediment load and accumulation in a very low energy environment. Although Unit III appears to be an alluvial deposit, a stable, exposed surface must have existed periodically, for a buried soil is present that is very well developed (i.e., two 3Btb horizons and one 3Btkb3 horizon; fig. 23). Long-term stability is further indicated by the oxidized color (e.g., 7.5YR4/4; moist), prismatic structure, and intense illuviation of carbonate to the lower part of the unit. Overall, the deposit is consistent with silty sand elsewhere in the region (e.g., Edwards 1 and 2, Stafford 1 and 2), suggesting accumulation during the Woodfordian. The surface of the deposit appears to have been severly truncated, probably by fluvial processes. An age of about 2,300 yr B.P. (figs. 7, 22, 23) from the uppermost 5 cm (2 in) of the unit implies last exposure during the late Holocene. A $\delta^{13}C$ value of -17.3‰ suggests the site was inhabited by relatively high percentages of C_4 plants.

Overlying Unit III is Unit II, extending from 79 cm to 1.58 m (2.6–5.2 ft) (figs. 22, 23). Classified as loamy fine sand, loam, and fine sandy loam (fig. 23), the deposits in Unit II are consistent in texture, structure, color, and

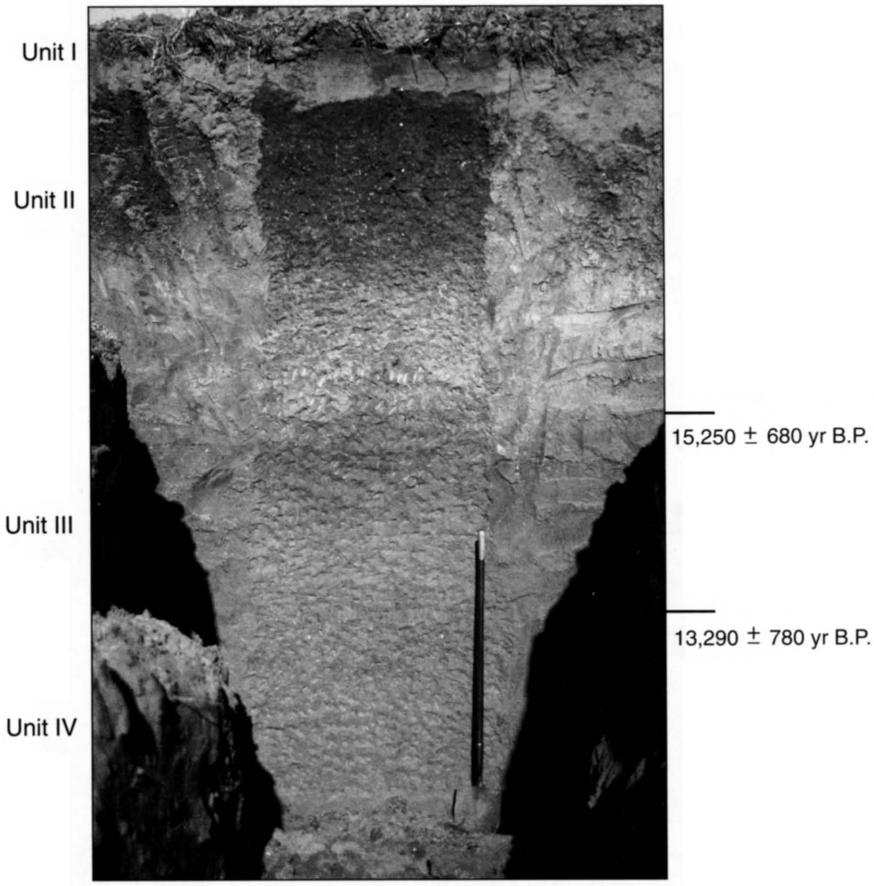


FIGURE 20—The 3.3-m (10.8-ft)-high exposure at Edwards 3, showing the position of pedostratigraphic units and radiocarbon ages.

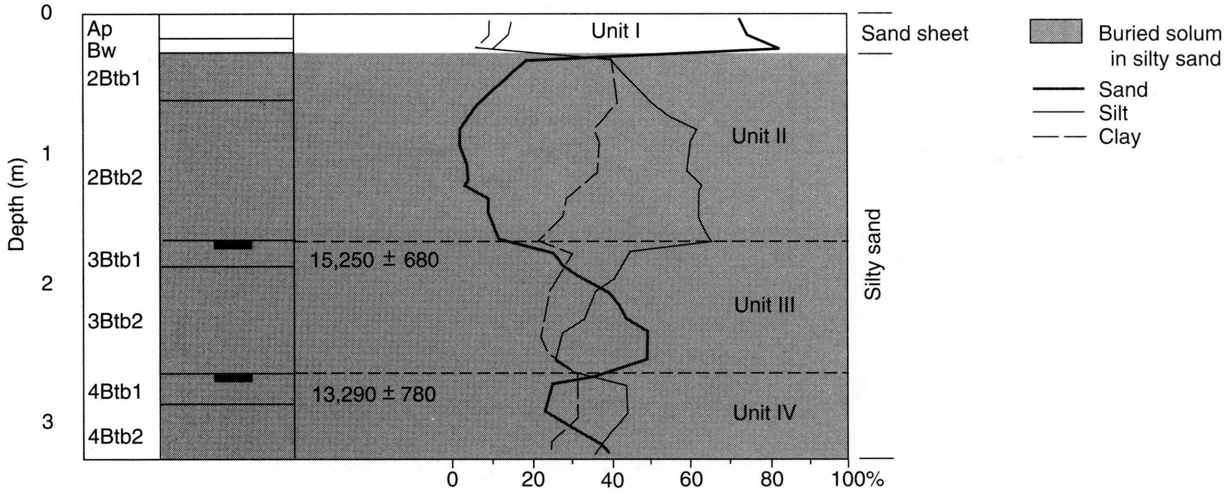


FIGURE 21—Pedostratigraphy, soil horizonation, and texture at Edwards 3. The radiocarbon ages are inverted, which may have resulted from slight contamination of one or both of the samples. Taken together, however, they verify a late Wisconsinan age for the silty sand at this site.

topographic position and expression with other units of eolian sand on the Great Bend Sand Prairie, those that accumulated in a relatively warm, probably more arid environment. The sediments are very poorly sorted, however, suggesting variable wind intensity or a nearby, poorly sorted source. Potentially, sediments in Unit II originated from exposures of the silty sand that were deflated during periods of aridity and high wind. Formed

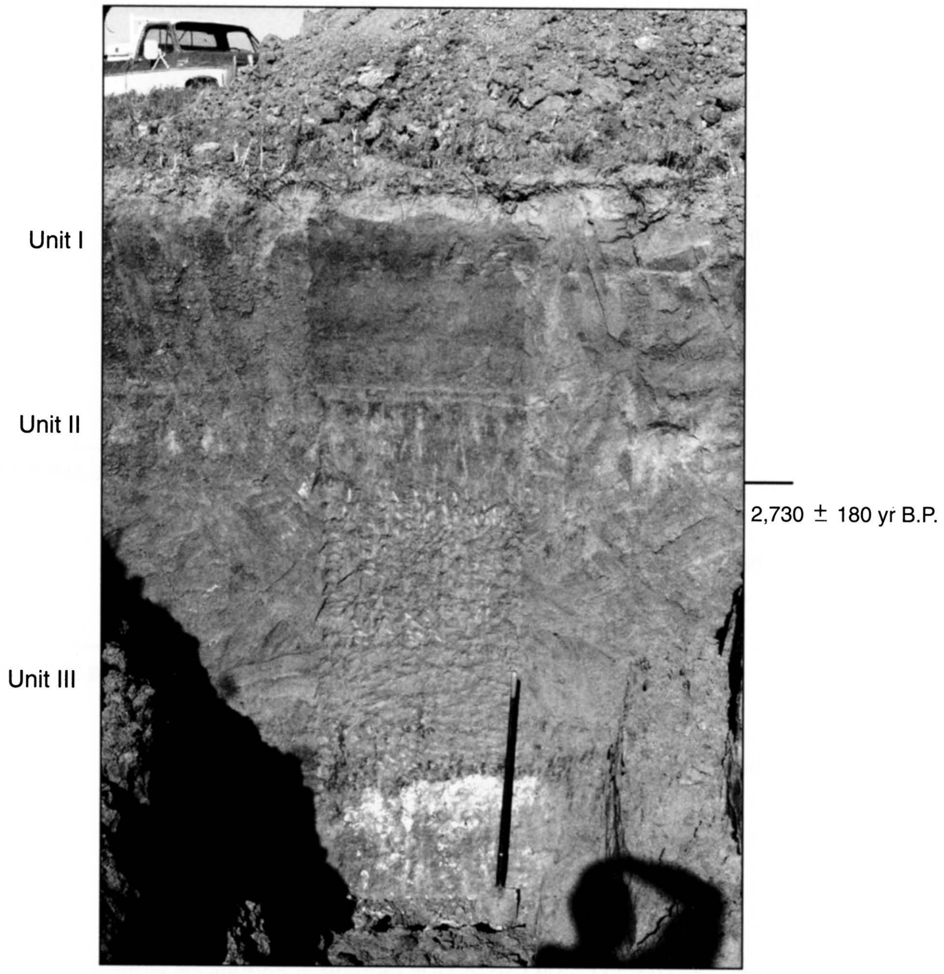


FIGURE 22—The 3.31-m (10.9-ft)-high exposure at Edwards 4 showing the position of pedostratigraphic units and radiocarbon age. Pedostratigraphy, soil horizonation, and texture at Edwards 4.

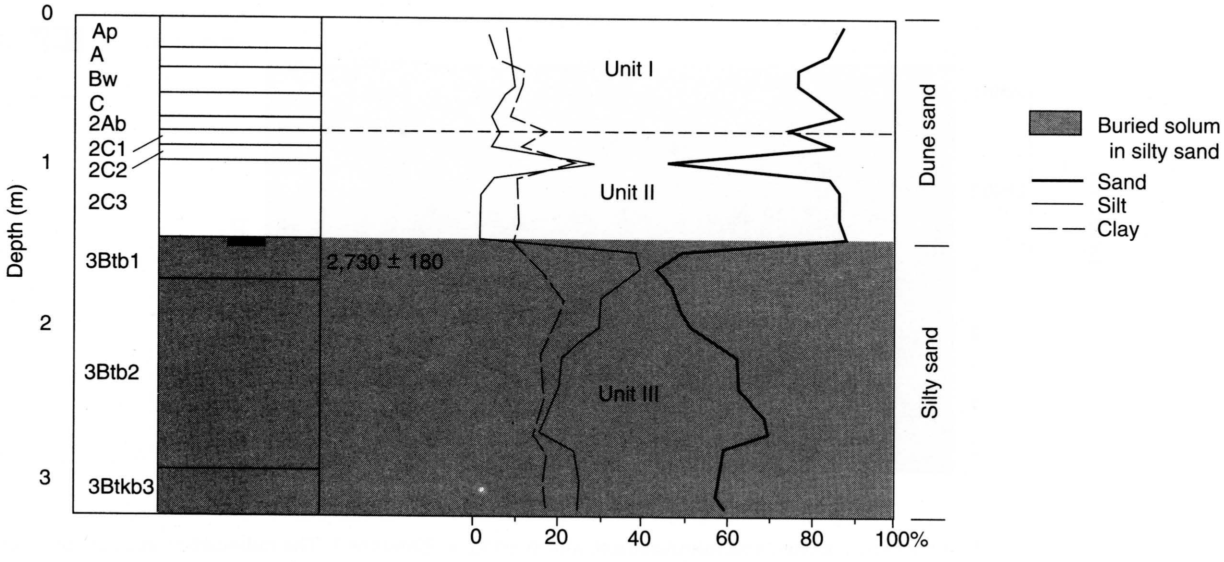


FIGURE 23—Pedostratigraphy, soil horizonation, and texture at Edwards 4.

within Unit II is a poorly developed buried soil, consisting of an 11-cm (4.3-in)-thick 2Ab horizon overlying three 2C horizons. The overall lack of structure (e.g., single grain, granular), as compared to that in the underlying silty sand (e.g., prismatic), in the soil further reflects eolian origins. Evidence suggests that Unit II accumulated during the late Holocene. Although a radiocarbon age was not obtained from the 2Ab, the age of the upper 3Btb1 (fig. 23) provides a maximum-limiting age for the deposit. The slightly oxidized color (e.g., 7.5YR3/4; moist) of the 2C horizons suggests a minimum-limiting age of greater than 1,000 years. Apparently, the period of stability was less than 1,000 years, as indicated by the overall lack of development.

Unit I extends from the surface to 79 cm (2.6 ft) (figs. 22, 23). In its sediments (fine sandy loam and sand) and structure (single grain to granular), Unit I is consistent with other deposits of eolian sand observed in the region. Sorting ranges from very poor to poor in the unit, suggesting winds of variable intensity or a nearby, poorly sorted source (e.g. silty sand). Contained within Unit I is a moderately to weakly developed surface soil consisting of two A horizons, a Bw and a C horizon (fig. 23). The character of the soil, coupled with its unoxidized color (10YR3/4; moist), suggests that the site has been stable for 500 to 1,000 years.

Phillips Trench

The Phillips Trench is a 2.7-m (8.9-ft)-deep section (figs. 7, 24) in a drainage pit located in a level sand sheet in the NE NE sec. 25, T. 13 S., R. 20 W. (fig. 6). Three pedostratigraphic units were described, with the lower pair in silty sand and the upper in sand (figs. 24, 25). These units indicate that at least three intervals of rapid sedimentation, interrupted by landscape stability and soil formation, occurred at the site during the late Quaternary. In general, deposits are loamy, very poorly sorted, finely to very finely skewed, and very platykurtic to leptokurtic (Arbogast, 1995).

Unit III is the basal deposit, ranging from about 2.35 m to 2.7 m (7.7–8.9 ft) (figs. 24, 25) and consisting of very poorly sorted, sandy sediments that are rich in silt. The deposit is slightly oxidized, indicating that some wetting and drying occurred following sedimentation. A well-developed soil subsequently formed, consisting of at least two 3Btb horizons (fig. 25). Although a radiocarbon age was not obtained from the soil, the solum is very similar in color, texture, and stratigraphic position with a buried soil recognized at the Belpre Trench that was dated to about 21,000 yr B.P. (figs. 8, 9). As a result, Unit III is assigned a late Farmdalian age.

Unit II is a 1.98-cm (6.5-in)-thick deposit (figs. 24, 25), mostly composed of calcareous silt and clay (fig. 25). The lower 58 cm (23 in) of Unit II is slightly oxidized, suggesting periodically moist conditions. Abundant gastropods, the majority of which were fragmented, were also observed at the base of the deposit. A few intact specimens were recovered and identified. The presence of *Discus cronkhitei, Helicodiscus singleyanus, Succinea avara,* and *Vertigo tridentata* (fig. 26) provides additional evidence for a cooler environment with probably more effective moisture, possibly during the early Woodfordian. Above this 58-cm (23-in) deposit is a 1.12-m (3.7-ft)-thick deposit of silt and clay that is slightly calcareous. In general, the stratum is similar in character (e.g., color, texture) with Peoria loess (Wells and Stewart, 1987; Johnson, 1993; Johnson et al., 1993), which has been recognized to the north and south of the Great Bend Sand Prairie (Feng, 1991; Feng et al., 1994). Given the late Farmdalian age assigned to the top of Unit III, a Woodfordian age is logical for Unit II. Although no sedimentary structures were observed, the very poorly sorted nature of Unit II, coupled with its loesslike nature, implies deposition by eolian processes. A long period of landscape stability must have subsequently occurred, promoting formation of a very well developed soil. A radiocarbon age of about 2,400 yr B.P. was obtained from the upper 5 cm (2 in) of Unit II, suggesting exposure during the late Holocene.

Overlying Unit II, with an extremely sharp stratigraphic contact, is Unit I (figs 24, 25). Whereas Unit II consists largely of silt and clay, Unit I consists predominantly of sand (fig. 25). Unit I is very poorly sorted, suggesting a nearby source (e.g., silty sand) that had a wide range of textures available for transport. A maximum-limiting age of approximately 2,400 yr B.P., derived from the top of Unit II, indicates that Unit I is a late Holocene deposit. Moreover, a $\delta^{13}C$ value of -15.8‰ implies that the climate was warmer and probably drier than when gastropods inhabited the site during the Woodfordian. The surface has been stable for a relatively brief period of time, as indicated by the weak development of the surface soil.

Reno 3

Reno 3 is a 3.24-m (10.6-ft)-deep section (figs. 7, 27) exposed by backhoe trenching in a compound parabolic dune field in the SE SE sec. 12, T. 25 S., R. 10 W. (figs. 6, 27). Four pedostratigraphic units were recognized, one in silty sand and three in sediments predominantly composed of sand (figs. 27, 28). The sedimentological differences between the silty sand and the overlying sand suggest a shift in depositional environment from low-energy alluvial or lacustrine to one where eolian processes dominated in a relatively warm, possibly more arid climate. In general, sediments at Reno 3 are noncalcareous, loamy, very poorly to moderately sorted, very finely to finely skewed, platykurtic to very leptokurtic, and have a mean particle size of medium silt to very fine sand (Arbogast, 1995).

The lowermost deposit is Unit IV, extending from 2.70 m (8.9 ft) to the base of the profile (figs. 27, 28). Unit IV includes very poorly sorted sediments, sandy clay-loam in texture, that are consistent with other deposits in the region generically categorized as silty sand. Evidence (e.g., very poor sorting, fining-upward sequences) obtained elsewhere (e.g., Edwards 2–4) suggests that these deposits accumulated during the late Wisconsinan or very early Holocene in a relatively moist, low-energy environment where fluvial or lacustrine processes probably dominated. Periodically, Unit IV must have been exposed and stable, as evidenced by a well-developed buried soil, consisting of two Btb horizons with prismatic structure, that formed in the deposit (fig. 28). Gleying in the lowermost horizon suggests long-term ponding of surface water or high ground-water tables from time to time.

Unit III ranges from 2.57 m to 2.70 m (8.4–8.9 ft) (figs. 27, 28). In general, this unit is consistent in texture, struc-

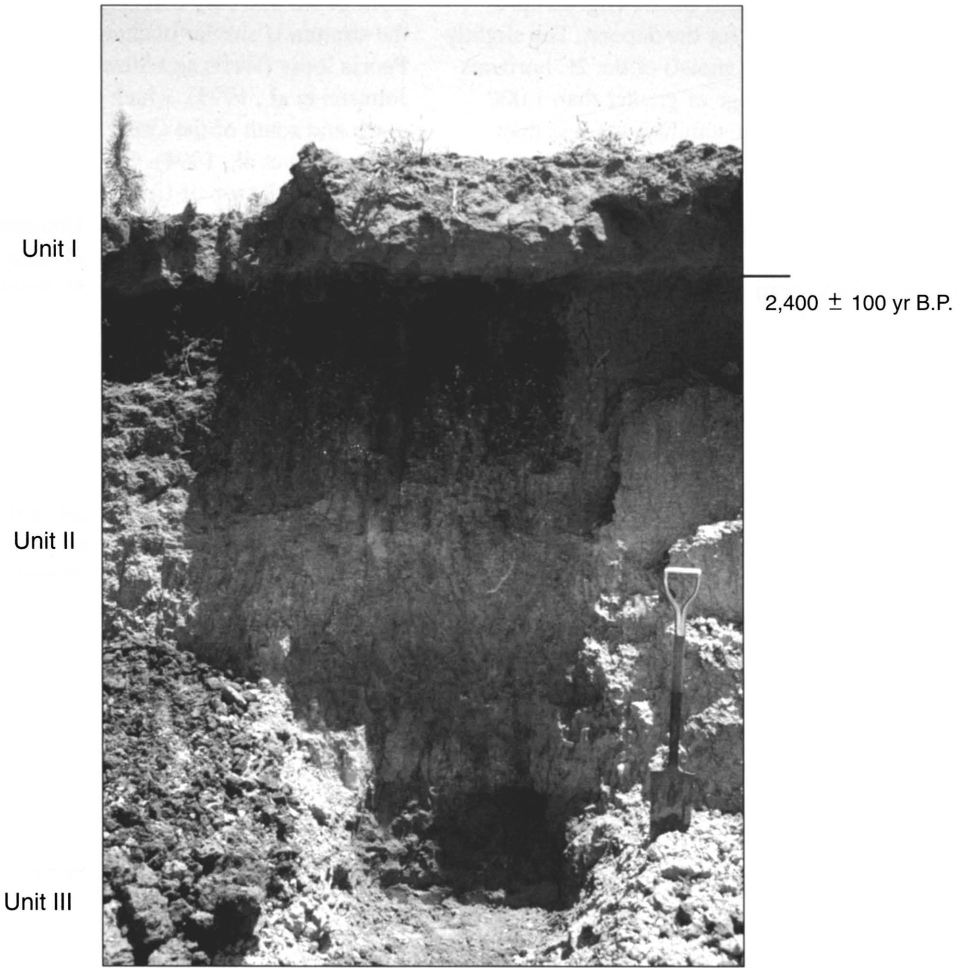


FIGURE 24—The 2.7-m (8.9-ft)-deep Phillips Trench, showing the position of pedostratigraphic units and radiocarbon age.

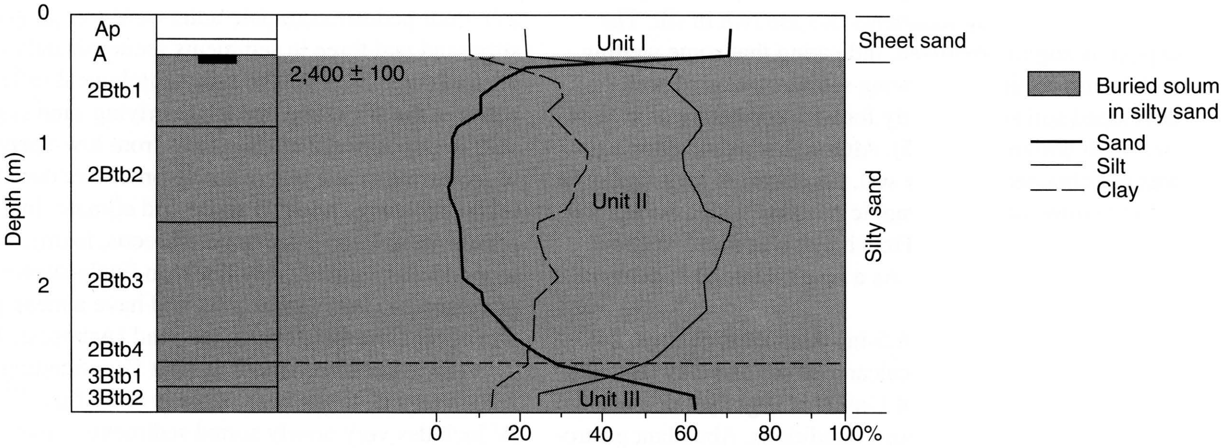


FIGURE 25—Pedostratigraphy, soil horizonation, and texture at the Phillips Trench.

ture, and color with sediments categorized as eolian sand in the region, deposits that accumulated in a relatively warm, arid environment. Texture, for example, shows an abrupt increase in sand from the underlying, presumably alluvial Unit IV. Although sorting is improved in Unit III, it is still very poor, suggesting a nearby, very poorly sorted source area such as the underlying silty sand. Alternatively, mobilizing winds could have varied in intensity, resulting in accumulation of various textures. Essentially, Unit III consists of a 3Ab soil horizon (fig. 28) with weak, subangular blocky structure, parting to single grain, that is also characteristic of a soil buried in eolian sand. A radiocarbon age of about 700 yr B.P. (figs. 27, 28) suggests accumulation during the very late Holocene when the climate was apparently warm with less effective moisture, one that promoted inhabitation by C_4 plants, as implied by a $\delta^{13}C$ value of -14.4‰.

The surficial deposits at Reno 3 consist of two stratigraphic units, Units II and I. Similar in character (figs. 27, 28), the sediments clearly represent a period of eolian sand deposition, one that was punctuated by a brief period of soil formation at the end of Unit II sedimenta-

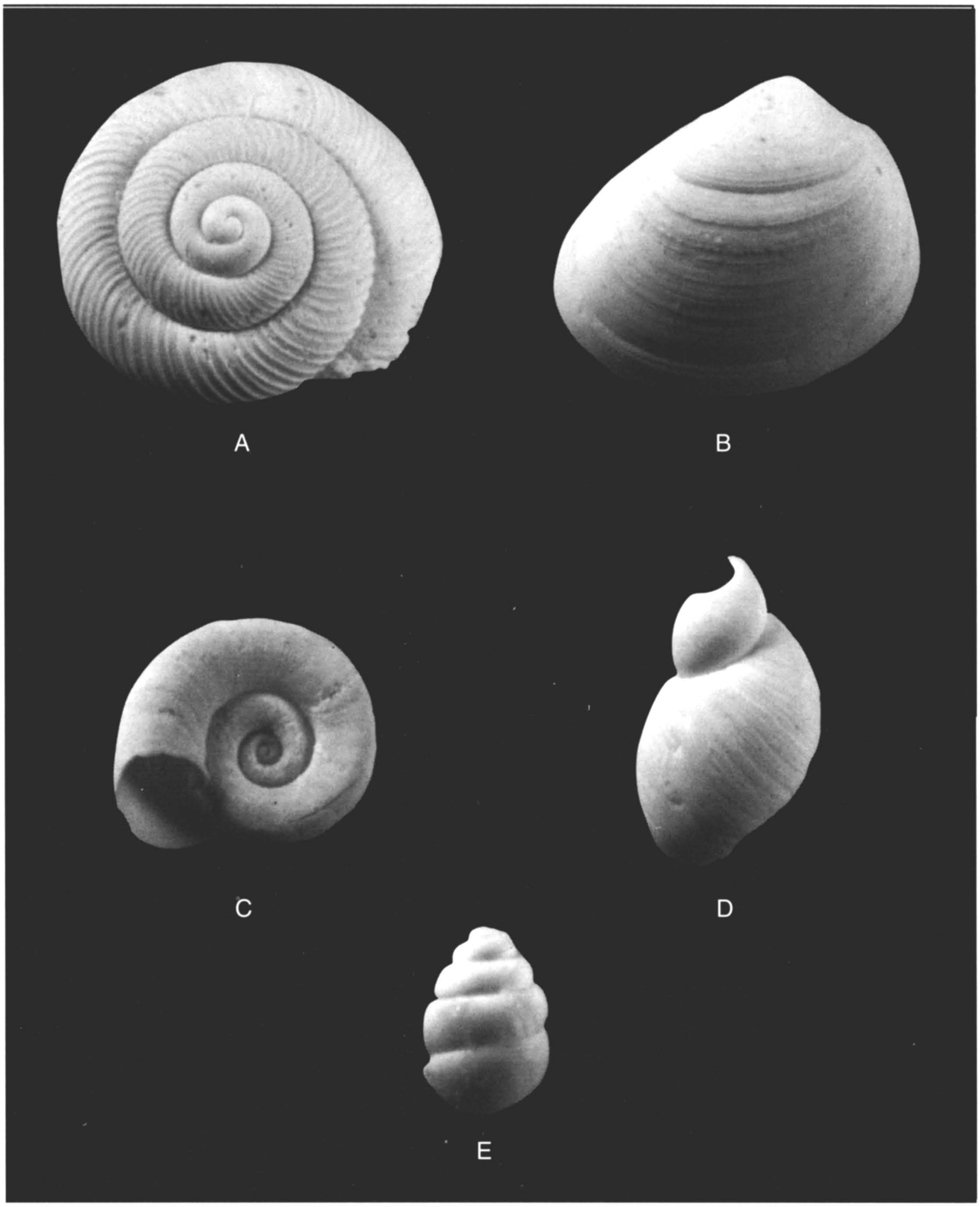


FIGURE 26—Mollusks recovered from the Phillips Trench (magnification ×10): (A) *Discus cronkhitei,* (B) unidentified bivalve, (C) *Helicodiscus syngleyanus*, (D) *Succinea avara*, (E) *Vertigo tridentata.*

tion. The deposits in both units are classified as sand, with as much as 97.4% sand present (e.g., 1.0 m; 3.3 ft). Compared to the underlying deposits, in Units II and I sorting improves considerably, becoming moderate; this suggests that eolian processes were dominant and the source was relatively well sorted. Each unit contains a

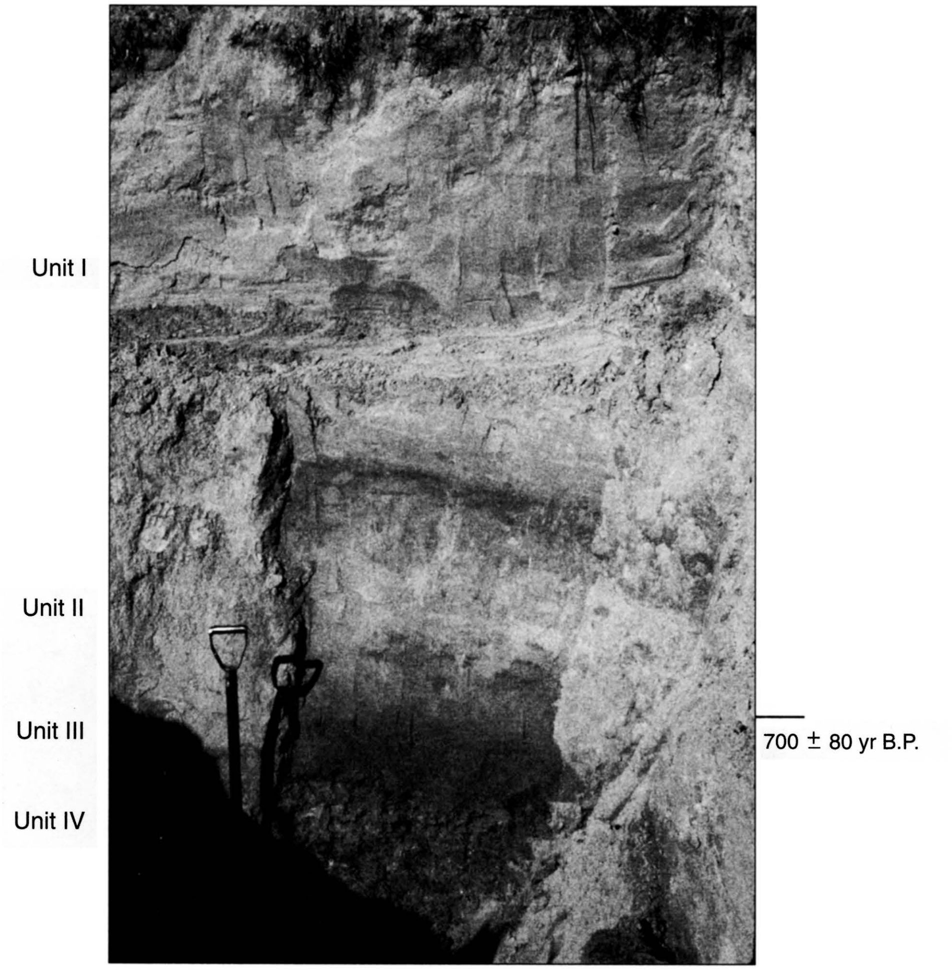


FIGURE 27—THE 3.24-M (10.6-FT)-HIGH EXPOSURE AT RENO 3, showing the position of pedostratigraphic units and the radiocarbon age.

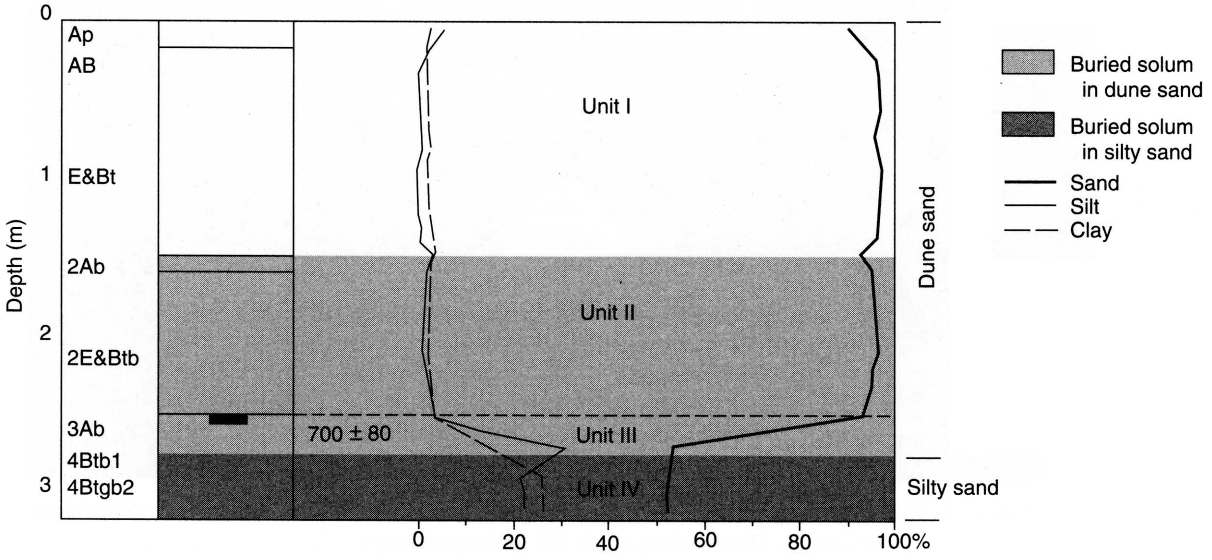


FIGURE 28—PEDOSTRATIGRAPHY, SOIL HORIZONATION, AND TEXTURE at Reno 3.

weakly developed soil virtually identical in character, consisting of thin A horizons overlying thick E&Bt horizons. Although the development of Bt horizons (fig. 28) suggests long-term stability, they are distinguished only by the presence of thin, weakly developed silt and clay lamellae that could have formed in a relatively brief period of time. In addition, the structure of the E&Bt horizons is single grain, further indicating short-term pedogenesis. Overall, both Units II and I appear to have accumulated sometime in the very late Holocene, as suggested by a maximum-limiting age of about 700 yr B.P. from the underlying 3Ab (figs. 27, 28). The last millennium at Reno 3, therefore, can be characterized as a period of episodic, eolian sand deposition. The lack of a well-developed surface soil suggests that the site has been stable for only a brief period of time.

Reno 4

Reno 4 is a 4-m (13-ft)-deep section (figs. 7, 29), exposed by backhoe trenching, located in a compound subparabolic dune field in the SE SW sec. 10, T. 25 S., R. 10 W. (fig. 6). Three pedostratigraphic units, consisting of silty sand overlain by eolian sand, were described (figs. 29, 30). The sedimentological and pedogenic differences among the units appear to reflect diachronous depositional environments at the site in the past 20,000 years. Sediments are generally noncalcareous, loamy, poorly to very poorly sorted, finely skewed, mesokurtic to very leptokurtic, and have a mean particle size ranging from fine silt to fine sand (Arbogast, 1995).

The lowest unit, Unit III, ranges from 2.27 m (7.4 ft) to the base of the profile (figs. 29, 30) and consists largely of gleyed silt, with significant amounts of sand and clay (fig. 30). Essentially, Unit III is very poorly sorted but is extremely poorly sorted in some places. A radiocarbon age of about 17,000 yr B.P. was derived from the lower part of the stratum (figs. 29, 30), suggesting sedimentation during the Woodfordian. Unfortunately, diagnostic sedimentary structures are not preserved, making a precise facies determination problematic. The very poorly sorted nature of the deposit, coupled with a distinct fining-upward sequence (fig. 30), however, suggests a low-energy fluvial or lacustrine environment. Cool conditions, with possibly more effective moisture, apparently existed during sedimentation, as implied by a δ^{13}C value of -19.9‰ near the base of the unit. Overall, the character of Unit III is entirely consistent with other deposits (Edwards 1–4, Reno 3) generally categorized as silty sand on the Great Bend Sand Prairie.

Subsequent to the deposition of Unit III, an extremely well developed soil formed. Whether the soil formed in one interval or several is impossible to determine. The solum is slightly gleyed and contains very little carbonate, suggesting intensive illuviation or high ground-water tables. A radiocarbon age of about 5,400 yr B.P. was obtained from the top of the soil (figs. 7, 29, 30), implying exposure during the late Holocene.

Overlying, with a very sharp contact, are Units II and I (figs. 29, 30). Although Units II and I represent different pedostratigraphic intervals, they can be grouped together because they are so similar in nature. In stark contrast to the underlying silty sand, Units II and I consist of very sandy deposits that are much better sorted (fig. 30). Units II and I are virtually identical in character (e.g., texture, color, structure, stratigraphic position, topographic expression) to deposits recognized throughout the Great Bend Sand Praire as wind-blown, dune sand. Such an eolian source implies a relatively warm environment with less effective moisture than that in which the silty sand was deposited.

A maximum-limiting age of approximately 5,300 yr B.P., taken from the top of Unit III, suggests that dune sand accumulated during the late Holocene at Reno 4. Sedimentation of eolian sand has been episodic, as indicated by the buried soil recognized at the top of Unit II. A radiocarbon age of about 700 yr B.P. from the upper part of the soil (figs. 7, 29, 30) suggests stability occurred between 1,000 and 500 years ago. Soil formation evidently lasted for some time, as indicated by the Btb horizons and E&Btb horizon (fig. 30) observed in the solum. Apparently, the soil was rapidly buried by an influx of relatively coarse, better-sorted eolian sand. Approximately 1.10 m (3.6 ft) of sand has accumulated at Reno 4 in the past few hundred years. The degree of surface soil development (i.e., two A horizons, E&Bt horizon; fig. 30) indicates that the dune has been stable for perhaps several hundred years.

Stafford 1

Stafford 1 is a 2.50-m (8.2-ft)-deep section (figs. 7, 31), exposed by backhoe trenching, in a compound parabolic dune field in the NW NW sec. 15, T. 25 S., R. 15 W. (fig. 6). Two pedostratigraphic units were described in the profile, one in silty sand and another in dune sand (figs. 31, 32). In general, the deposits at Stafford 1 are noncalcareous, loamy, have a mean particle size ranging from medium to coarse silt, are very poorly to poorly sorted, and are very finely skewed and leptokurtic to very leptokurtic (Arbogast, 1995). The sedimentology of Unit II suggests a low-energy fluvial or lacustrine depositional environment. That of Unit I suggests a predominantly eolian environment.

Unit II is the thickest deposit, extending from 98 cm (39 in) to the base of the profile. Deposits are generally sandy loam, containing as much as 46.4% silt and 16.6% clay (fig. 32), and are very poorly sorted. Overall, Unit II is consistent with other silty-sand deposits in the region. The degree of sorting, coupled with a subtle but steady fining-upward sequence, suggests accumulation in a low-energy alluvial or lacustrine environment during a time of increased effective moisture. Periods of exposure and relatively long-term stability must have occurred, however, as indicated by the thick, well-developed soil with prismatic structure that formed within the unit. Alternating periods of wetting and drying are suggested by soil characteristics. Towards the base, the solum is gleyed, mottled, and relatively rich in calcium carbonate, suggesting ponded water or high ground-water tables. In the middle,

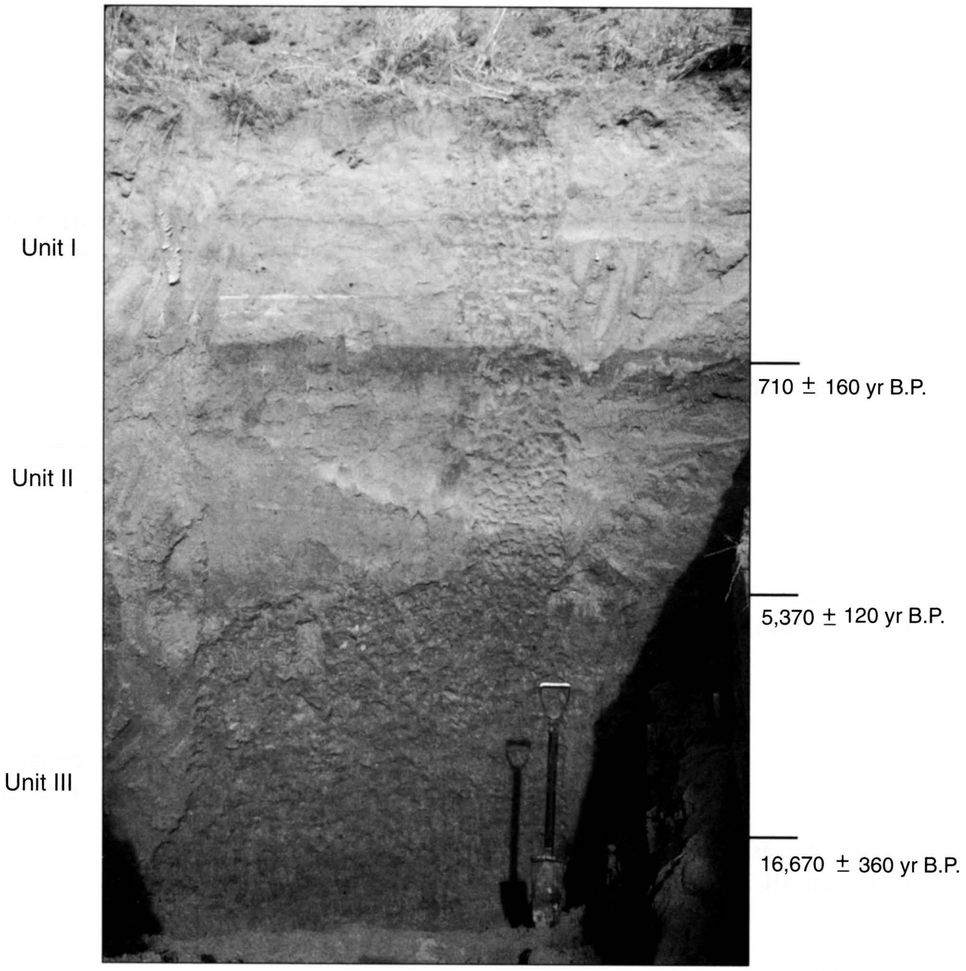


FIGURE 29—The 4-m (13-ft)-high exposure at Reno 4, showing the position of pedostratigraphic units and radiocarbon ages.

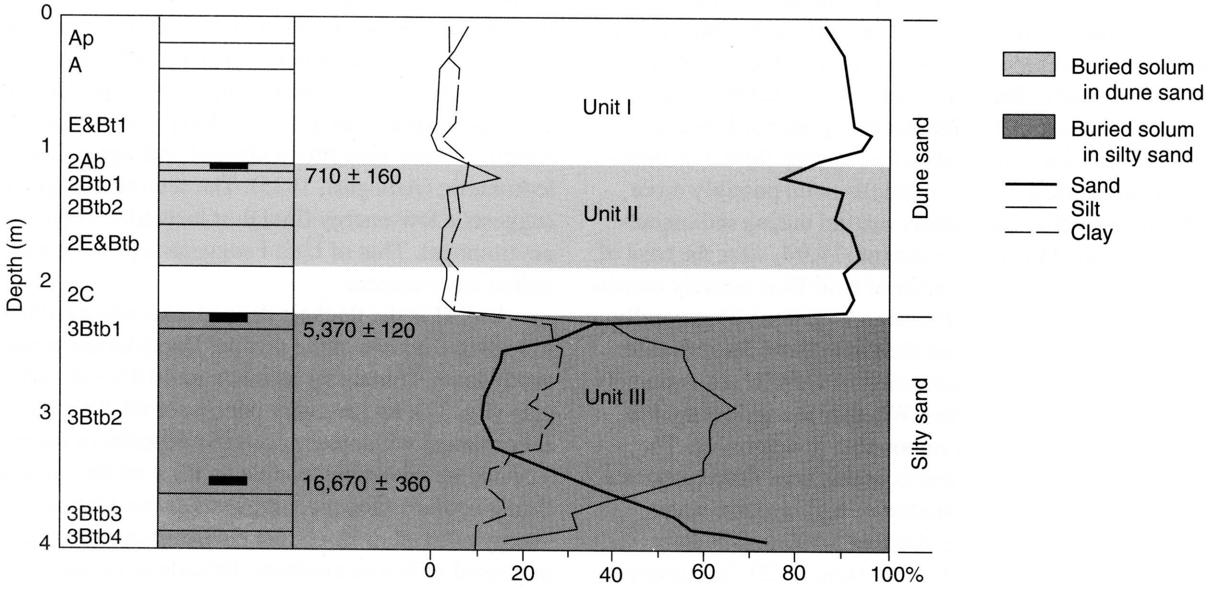


FIGURE 30—Pedostratigraphy, soil horizonation, and texture at Reno 4.

however, the sediment is slightly oxidized, indicating weathering in a more arid environment. In the upper part of the soil, the sediment is once again gleyed, implying a return to moist conditions. Ultimately, the soil must have been exposed again, and for some time, resulting in complete truncation of the A horizon. A radiocarbon age of

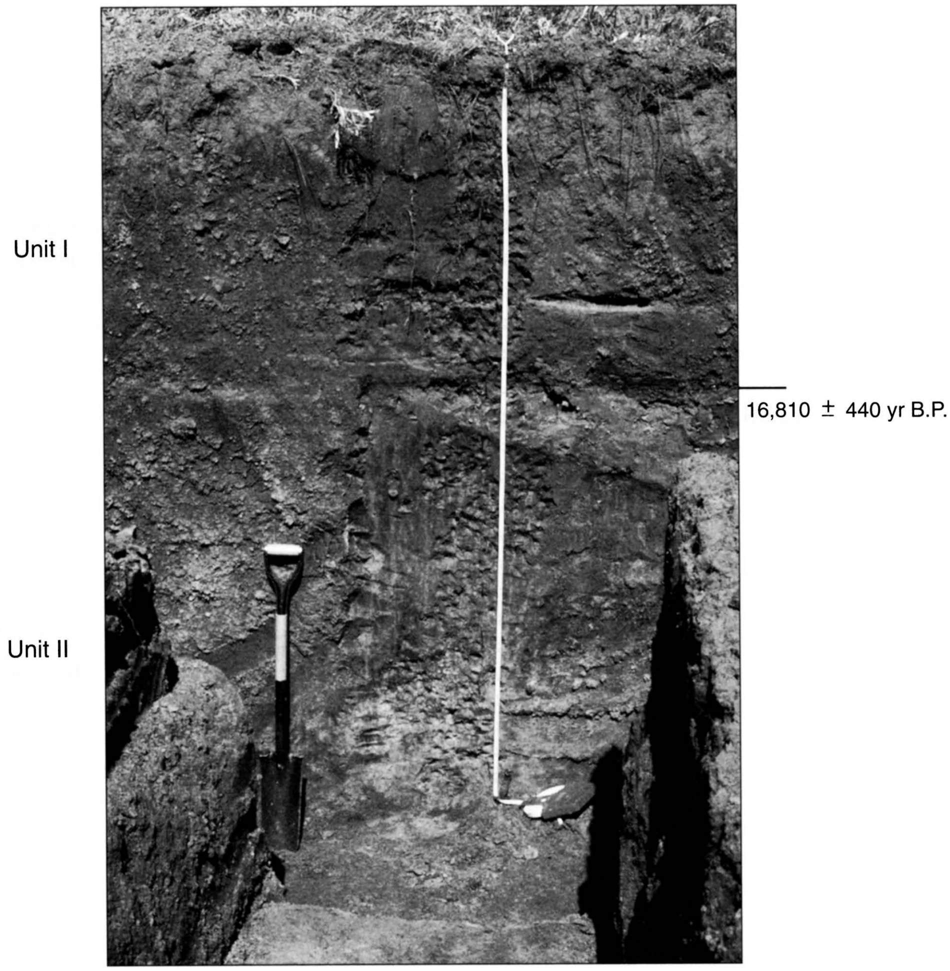


FIGURE 31—THE 2.50-M (8.2-FT)-HIGH EXPOSURE AT STAFFORD 1, showing the position of pedostratigraphic units and radiocarbon ages.

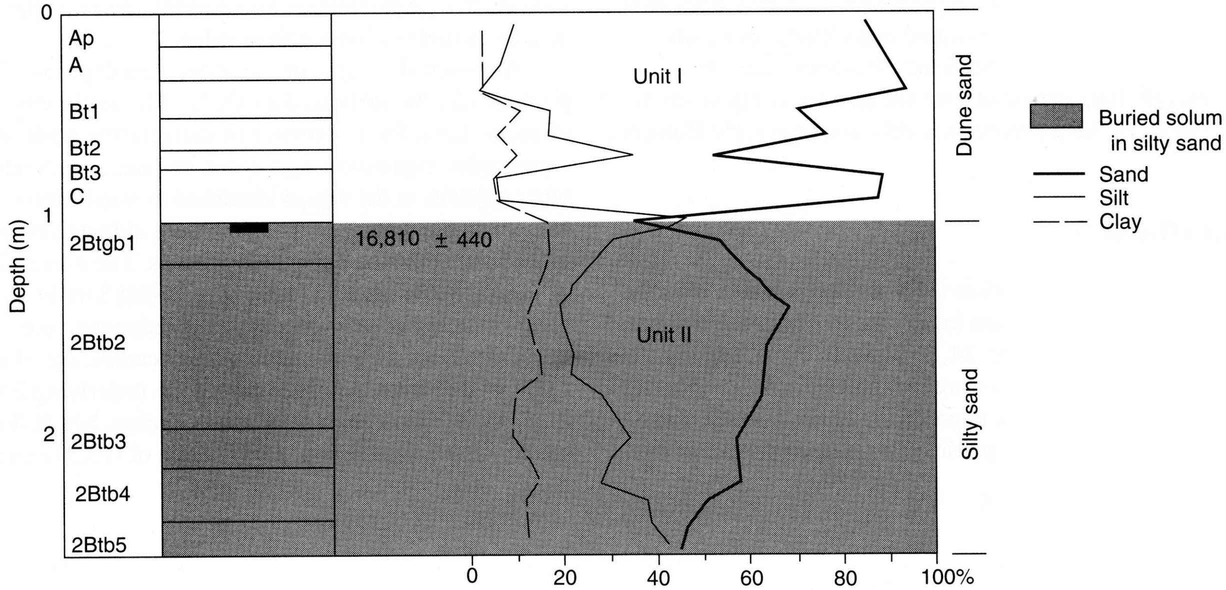


FIGURE 32—PEDOSTRATIGRAPHY, SOIL HORIZONATION, AND TEXTURE at Stafford 1.

about 17,000 yr B.P. from the upper part of the soil indicates that Unit II is probably a late Wisconsinan deposit. A $\delta^{13}C$ value of -21.9‰ suggests proportionately high percentages of C_3 plants inhabited the site, further implying a relatively cool climate with possibly more effective moisture.

The surficial deposit, Unit I, consists largely of single-grained sand between the surface and 98 cm (39 in) (figs. 31, 32). At the base of the unit, the deposit contains approximately 90% sand, but texture fines sharply in the middle of the stratum, with about 36% silt and 10% clay present. Towards the top of the unit, the sand content increases sharply again to around 90% (fig. 32).

Overall, the character (e.g., texture, structure, color, topographic position and expression) of Unit I is consistent with other deposits generally classified as eolian sand in the region. A sharp increase in sorting in the lowermost stratum of Unit I, as compared to the underlying deposits, is indicative of eolian sedimentation. Apparently, this change in facies occurred sometime during the late Wisconsinan, as suggested by an approximate age of 17,000 yr B.P. from the uppermost part of Unit II. Varied textures in the middle of Unit I, coupled with a sharp decrease in sorting, imply mixed alluvial and eolian deposition, possibly when the site occupied an interdunal position. Near the surface, a sharp improvement in sorting, coupled with percentages of sand of about 90%, suggests a return to eolian sedimentation.

It is difficult to estimate the age of Unit I. Evidence suggests that the unit is relatively old, possibly early Holocene. Based on the late Wisconsinan age of the underlying Unit II, the lowermost part of Unit I appears to be a late Wisconsinan deposit. Assuming no truncation, the entire thickness of Unit I would then be of a similar age. In contrast, the development of the surface soil, although excellent for one in eolian sand, does not imply great antiquity. Accordingly, it appears that an unconformity exists between the upper and lower parts of Unit I and that two periods of eolian sedimentation are represented. The development of the surface soil, with illuviated clays lining root casts, strongly suggests that Stafford 1 has been stable for a relatively long period of time for an eolian deposit on the Great Bend Sand Prairie, possibly since the early Holocene.

Stafford 2

Stafford 2 is a 3.65-m (11.9-ft)-deep section (figs. 7, 33), exposed by backhoe trenching, in a high-relief sand sheet in the NE NE sec. 24, T. 24 S., R. 13 W. (fig. 6). Four pedostratigraphic units were recognized in the profile (figs. 33, 34). In general, the deposits are noncalcareous, loamy, fine sand with a mean particle size ranging from fine to coarse silt. The sediments are poorly to very poorly sorted, consistently very finely skewed, and very leptokuric to mesokurtic (Arbogast, 1995).

The deposits at Stafford 2 can be considered as two units: Unit I, a surficial deposit of unoxidized dune sand, and Units II–IV, an underlying sequence of oxidized, siltier sediments. The origin of the upper unit can be determined with certainty. Conflicting evidence exists, however, as to the depositional environment in which Units II through IV accumulated.

Three variables suggest that Units II–IV have a fluvial origin. First, the deposit is very poorly sorted, suggesting a mixed sediment load that may occur in a low-energy fluvial or lacustrine environment. Second, a $\delta^{13}C$ value of -23.6‰, derived from the top of the deposit, suggests that C_3 plants were present in high numbers at the site, evidence of a cool climate with possibly more effective moisture. Third, the deposit contains two, well-defined fining-upward sequences, further indicative of alluvial sedimentation.

Other evidence, however, favors an eolian source for Units II–IV. Sand is a much larger component in the stratum at Stafford 2 than in other silty sands, making up as much as 85.1% (fig. 34). The radiocarbon age of approximately 10,400 yr B.P., obtained from the base of the deposit (figs. 7, 33, 34), also points to an eolian source. At other localities where the lower part of the silty sand was dated (e.g., Edwards 2 and 3, Reno 4), late Wisconsinan ages greater than 13,000 yr B.P. were derived. The age obtained from the base of Stafford 2, however, strongly suggests that the remainder of the section accumulated during the Holocene, a time of demonstrably warmer, more arid conditions. Overall, Units II through IV appear to be predominantly eolian in nature, especially in the lower part, but probably include some deposits of a very low-energy (e.g, interdune) fluvial origin. The presence of three, well-developed buried soils indicate that sedimentation was episodic and was punctuated by periods of long-term stability.

Although the origin of the underlying deposits is problematic, the surficial deposit, Unit I, clearly consists of eolian sand, for it is similar in stratigraphic position, topographic expression, age, color, texture, and structure to other deposits in the region identified as wind-blown sand. Moreover, the sorting of the upper unit, although poor, is much better than the deposits it overlies. The overall lack of sorting in the sand, in fact, suggests that Unit II, or a nearby outcrop of very poorly sorted sediments (e.g., silty sand) was the sediment source. A radiocarbon age of about 1,000 yr B.P. from the upper part of the underlying 2Ab (figs. 33, 34) provides a maximum-limiting, late Holocene age for Unit I. In addition, a $\delta^{13}C$ value of -13.3‰ clearly implies that the associated climate was relatively warm with probably less effective moisture. The degree of surface soil development (i.e., A/AC/C horizonation) suggests that, in general, the dune field immediately surrounding Stafford 2 has been stable for a relatively short period of time, probably less than 500 years.

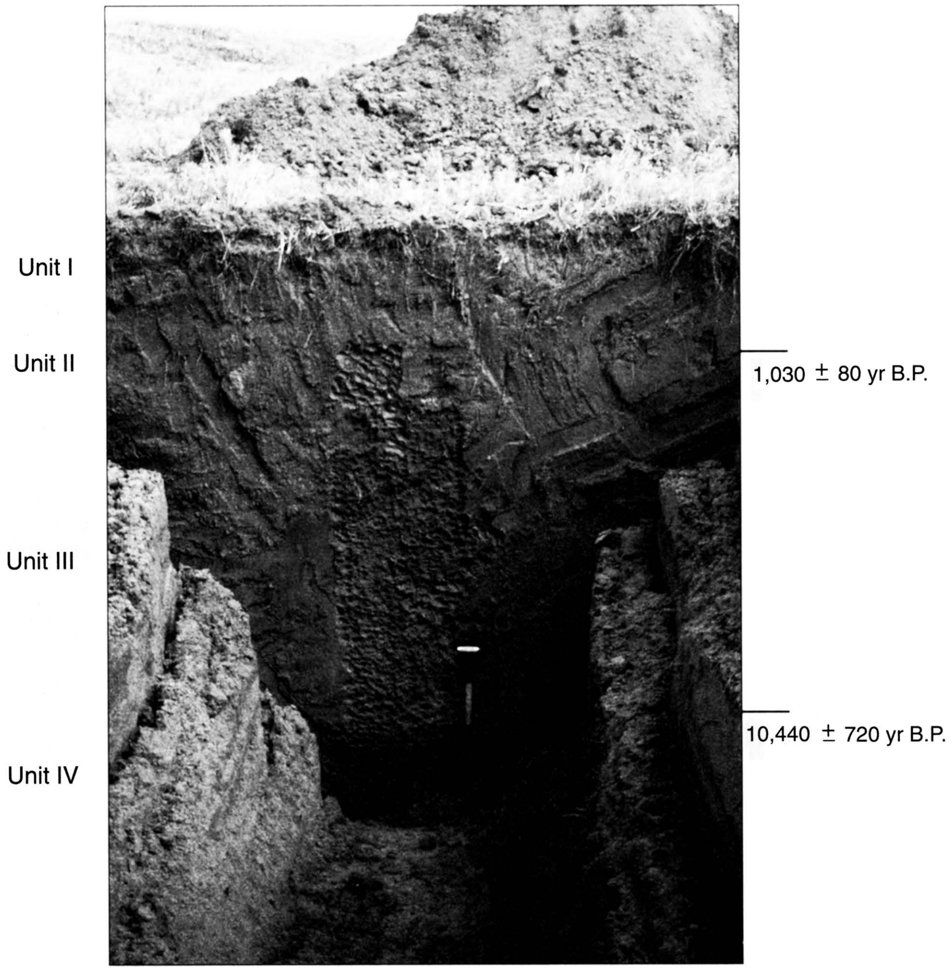


FIGURE 33—The 3.65-m (11.9-ft)-high exposure at Stafford 2, showing the position of pedostratigraphic units and radiocarbon ages.

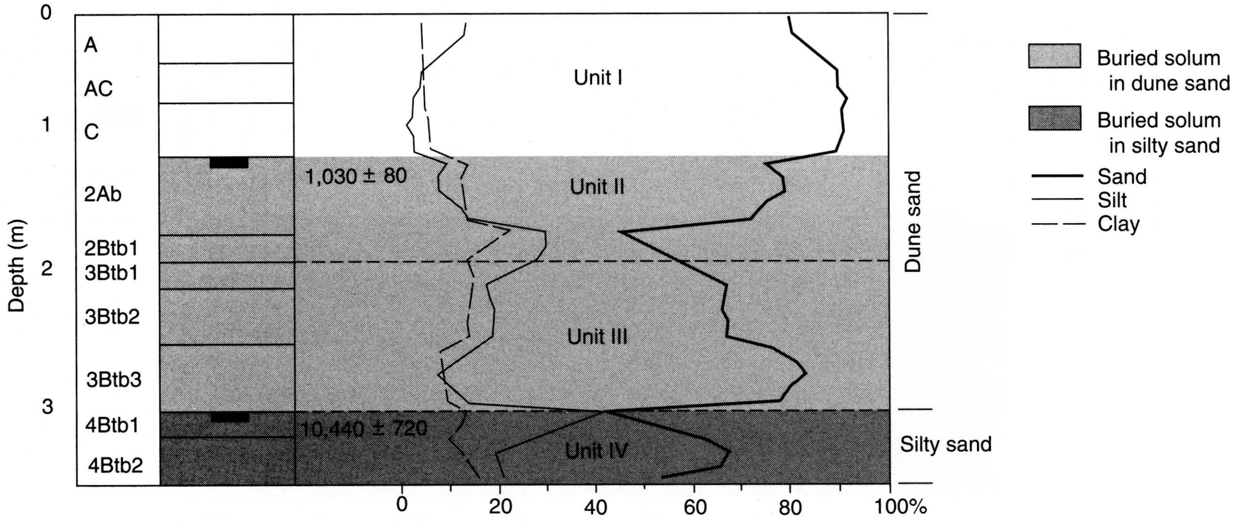


FIGURE 34—Pedostratigraphy, soil horizonation, and texture at Stafford 2.

Stafford 3

Stafford 3 is a 4.58-m (15.0-ft)-deep section (figs. 7, 35), exposed by backhoe trenching, in a compound parabolic dune field, in the NW NW sec. 23, T. 24 S., R. 13 W. (fig. 6). Three pedostratgraphic units were recognized, Unit III in silty sand and Units I and II in eolian sand (figs. 35, 36). Skewness (finely skewed) and kurtosis

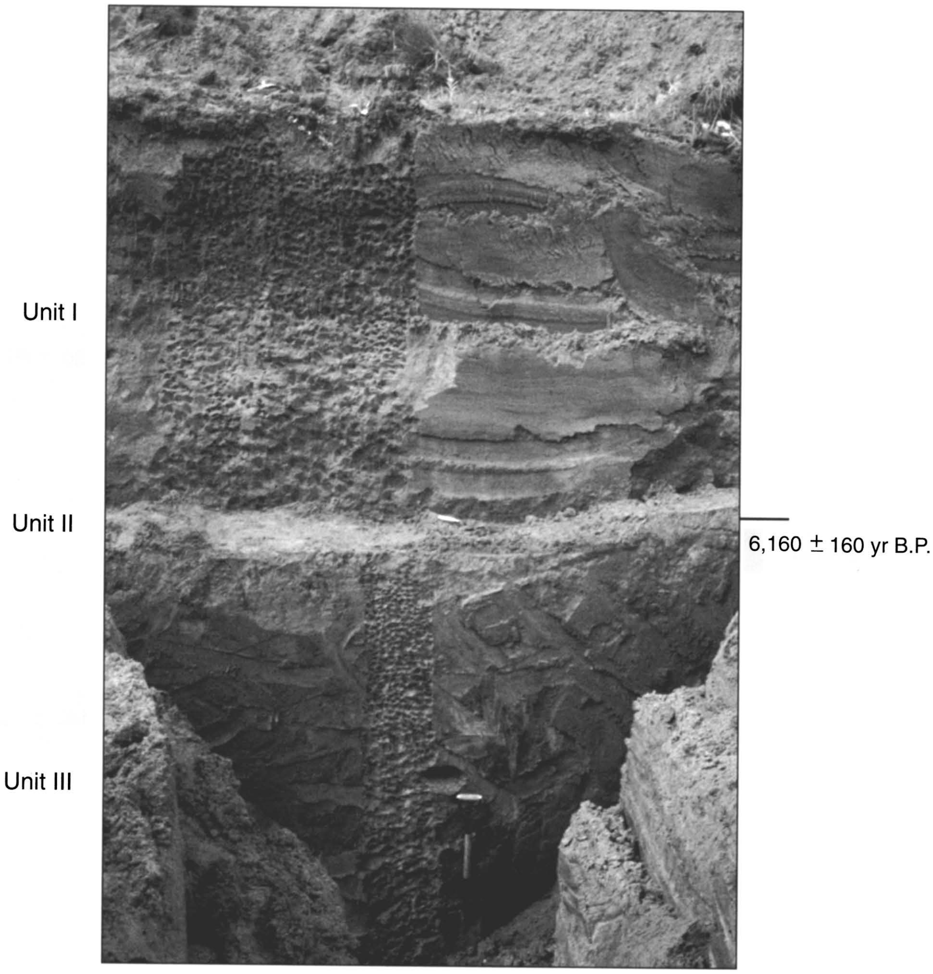


FIGURE 35—THE 4.58-M (15.0-FT)-HIGH EXPOSURE AT STAFFORD 3, showing the position of pedostratigraphic unts and radiocarbon ages.

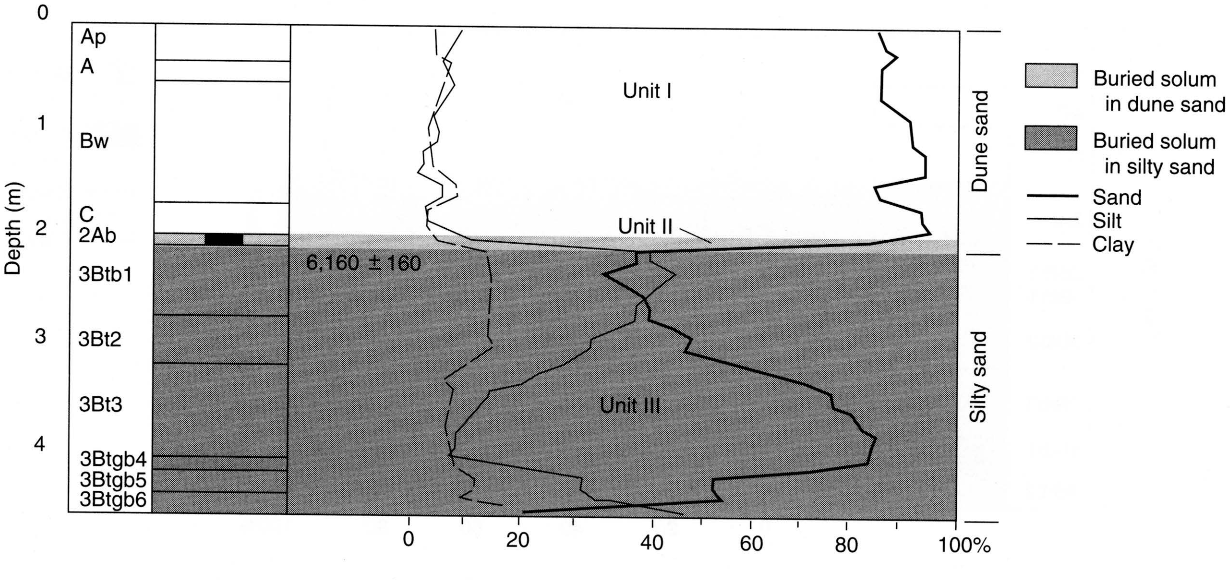


FIGURE 36—PEDOSTRATIGRAPHY, SOIL HORIZONATION, AND TEXTURE at Stafford 3.

(very leptokurtic to leptokurtic) are consistent in these deposits, all of which are basically noncalcareous. Unit III is much more poorly sorted, however, than the overlying deposits and contains abundant silt and clay. In contrast, Units II and I are much sandier and better sorted (Arbogast, 1995). The sedimentological differences between the lower (Unit III) and upper (Units II and I) units reflect diachronous depositional environments. Theoretically, Unit III represents a predominantly low-energy fluvial or lacustrine facies that accumulated in a wetter climate, whereas Units II and I represent eolian sedimentation in a relatively arid environment.

Unit III is consistent with silty sands recognized elsewhere on the Great Bend Sand Prairie. The stratigraphic position of the deposit (i.e., beneath eolian sand), coupled with the degree of sorting and high percentages of silt and clay, favors this correlation. Unfortunately, deposits such as Unit III contain no sedimentary features, rendering facies categorization difficult. Sedimentologically, Unit III is very poorly sorted, indicative of a low-energy fluvial or lacustrine environment where a mixed sediment load existed. In addition, the deposit contains two, very distinct, fining-upward sequences (fig. 36), which futher imply fluvial sedimentation. The deposit could be the result of both fluvial and eolian sedimentation. Because the deposit contained no measurable organic carbon, its age could not be determined through radiocarbon dating. Based on ages obtained from similar units elsewhere (e.g., Edwards 2 and 3, Stafford 1, Reno 4), which were deposited sometime between 20,000 and 13,000 yr B.P., Unit III is assigned a late Wisconsinan age.

Regardless of when and how Unit III accumulated, it is clear that following sedimentation, a stable surface existed for some time, one that promoted formation of a very strongly developed soil (fig. 36). Each horizon is strongly mottled, with the lower three heavily gleyed, suggesting that ponded water at the surface or high water tables promoted chemical illuviation. At some point, a period of surficial erosion must have truncated the A horizon from the top of the soil.

The derivation of Units II and I (figs. 35, 36) can be ascertained with relative certainty. Both are consistent in stratigraphic position, texture, cohesiveness, color, and sorting with eolian sand. Unit II, which is 11-cm (4.3-in)-thick, sits unconformably on Unit III. Essentially, Unit II consists of a 2Ab horizon that formed following deposition of wind-blown sand. A radiocarbon age of about 6,200 yr B.P. from the upper part of the 2Ab (fig. 36) suggests last exposure during the middle Holocene when climate was relatively warm, and this is further indicated by a δ^{13}C value of -17.9‰. Moreover, the age provides a maximum-limiting age for the overlying, 1.96-m (6.4-ft)-thick dune deposits of Unit I. Unit I contains a moderately developed surface soil, containing one Bw horizon. Although the 6,200-yr-B.P. age obtained from Unit II implies a middle Holocene age for Unit I, the relative lack of soil formation at the surface suggests it is younger. Apparently, an unconformity exists somewhere in Unit I, one that represents a period of erosion in the later Holocene that removed the upper part of a pre-existing dune, but did not scour Unit II. Subsequently, the remainder of Unit I was deposited and the surface soil developed. Overall, the degree of soil formation suggests that the dune has been stable for less than 1,000 years.

Stafford 4

Stafford 4 is a 3.75 m (12.3 ft) section of loess (figs. 7, 37), exposed by backhoe trenching, located in the loess plain in the NW NE sec. 16, T. 24 S., R. 12 W. (fig. 6). Although the trench was excavated to 3.75 m (12.3 ft), instability precluded sampling below 2.85 m (9.4 ft). Two pedostratigraphic units were recognized at the site (figs. 37, 38). Deposits at the site are basically noncalcareous and loamy, have a mean particle size ranging from coarse to fine silt, are very poorly sorted, and are very fine to finely skewed and very leptokurtic to mesokurtic (Arbogast, 1995).

In general, the lowermost part of Unit II consists of very sandy (>83% sand), slightly laminated sediments that lack structure. The character (e.g., texture, structure, sorting) of these deposits is consistent with dune sand. In the upper part of Unit II, the sediment is much siltier (e.g., 40%) in the presence of a moderately well developed buried soil, one that consists of one 2Ab horizon, one 2ABb horizon, and two 2Btb horizons. A radiocarbon age of approximately 12,800 yr B.P. from the upper part of the soil (fig. 38) indicates that Unit II is likely a late Wisconsinan deposit. In addition, the unit is slightly oxidized (e.g., 7.5YR4/6; moist), further suggesting relative antiquity. At most localities where late Wisconsinan ages have been obtained (e.g., Edwards 1–3, Stafford 1, Reno 4), the associated deposits imply a fluvial or lacustrine facies of mixed energy. The age obtained from the 2Ab at Stafford 4 suggests that dune formation apparently occurred in some parts of the study area between 20,000 and 10,000 yr B.P. A δ^{13}C value of -14.2‰ suggests that plants specialized for sandy, well-drained landscapes dominated the site.

Unit I conformably overlies Unit II and extends from the surface to 1.66 m (5.4 ft) (figs. 37, 38). Fundamentally, Unit I is silty, very poorly sorted, and lacks defineable sedimentary structures. At first glance, the unit is similar to other strata in the study area categorized as silty sand, but in this case it is not buried. However, the radiocarbon age of approximately 12,800 yr B.P. from the underlying 2Ab (fig. 38) provides a maximum-limiting, late Wisconsinan age for the unit, a chronology recognized at several localities (e.g., Edwards 3; Stafford 11) where the silty sand has been recognized. On closer examination, however, several variables suggest that Unit I reflects a different environment from that in which the silty sand accumulated. First, the unit lies conformably over Unit II, a relationship unrecognized elsewhere in the region. Second, the deposits associated with Unit I have not been reduced or oxidized as has the silty sand at other localities. Third and most importantly, the soil horizons associated with Unit I have relatively poor structure, largely weak prismatic parting to weak subangular blocky, when compared to the moderate to strong prismatic structure observed elsewhere in

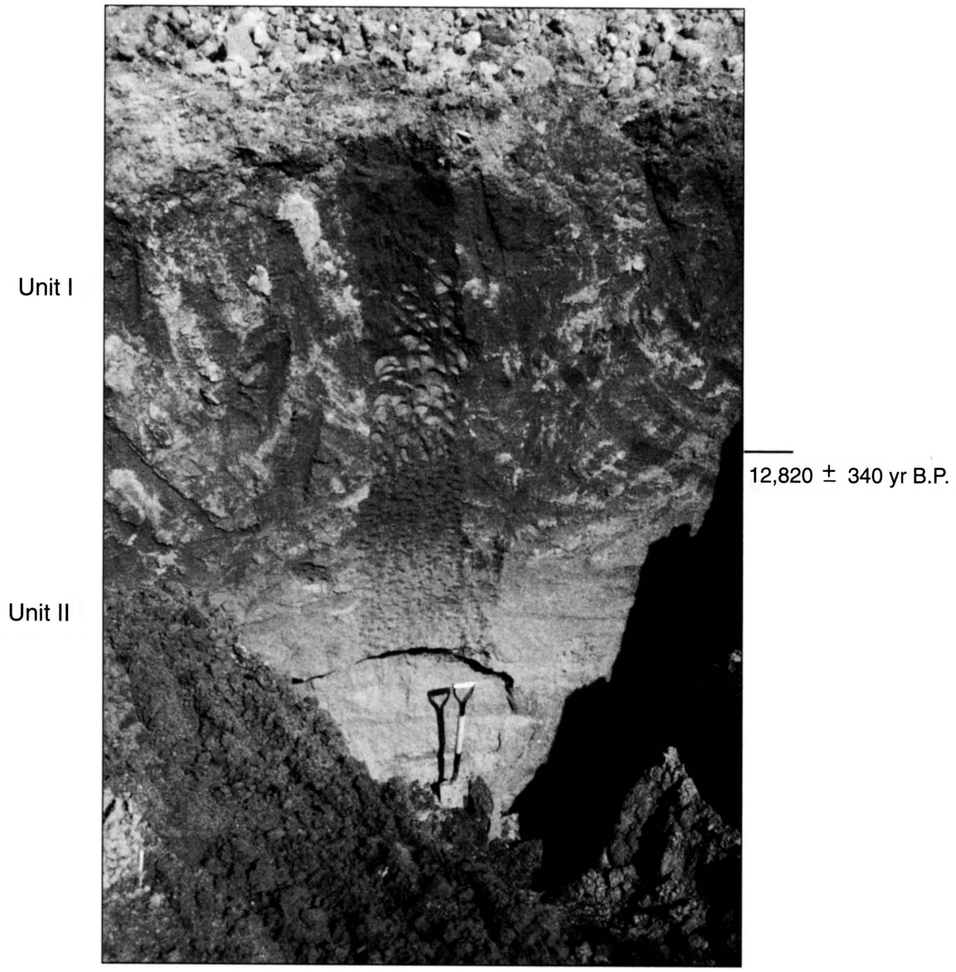


FIGURE 37—The 3.75-m (12.3-ft)-high exposure at Stafford 4, showing the position of pedostratigraphic units and radiocarbon age.

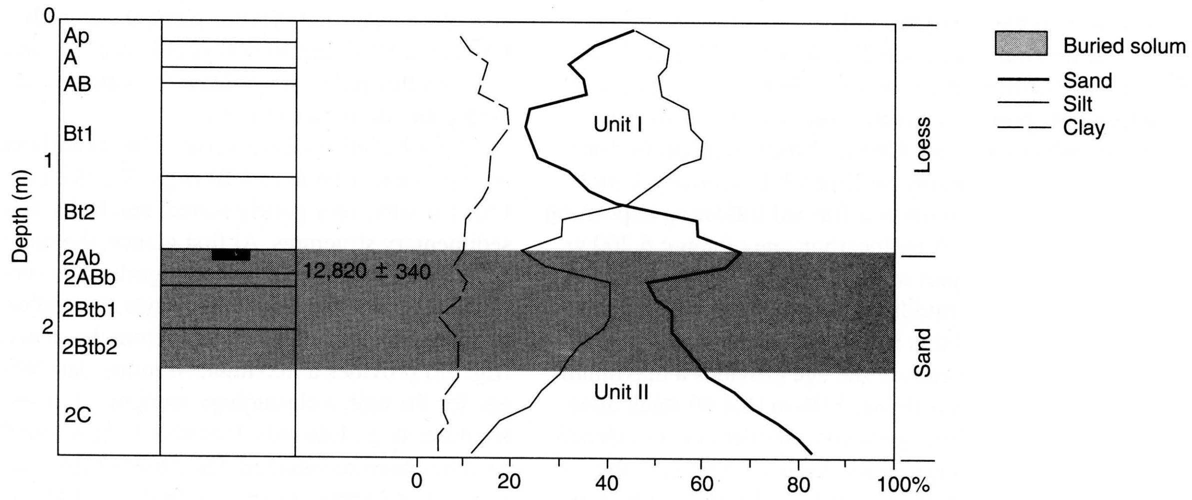


FIGURE 38—Pedostratigraphy, soil horizonation, and texture at Stafford 4.

the silty sand. The character of Unit I implies an eolian facies that consisted predominantly of silt at Stafford 4 rather than sand. As a result, Unit I is interpreted to be a Holocene wind-blown deposit. Apparently, the site has been stable for at least 1,000 years, as suggested by a well-developed surface soil that includes two Bt horizons.

Stafford 5

Stafford 5 is a 3.46-m (11.4-ft)-high section (figs. 7, 39), exposed by backhoe trenching, in a compound parabolic dune field in the SW SW sec. 6, T. 24 S., R. 13 W. (fig. 6). Three pedostratigraphic units were described (figs. 39, 40), one in very poorly sorted deposits of sand, silt, and clay, and two in overlying, relatively well sorted dune sand. In general, the contrasting sedimentology between Unit III and the overlying deposits appears to reflect changes in depositional environments at the site in the past 20,000 years. Two buried soils were described, indicating that sedimentation was punctuated by periods of stability and pedogenesis (fig. 40). The deposits at Stafford 5 are noncalcareous, loamy, very poorly to moderately sorted, very finely to finely skewed, mesokurtic to very leptokurtic, and have a mean particle size of medium silt to very fine sand (Arbogast, 1995).

Unit III ranges from 2.03 m (6.7 ft) to the base of the profile (figs. 39, 40). Although abundant in sand (e.g., 66.2%), it also contains relatively high amounts of silt (e.g., 36.5%) and clay (21.8%; fig. 40) and, as a result, is very poorly sorted. The overall lack of sorting, coupled with the presence of three fining-upward sequences in the stratum, suggest accumulation in a very low-energy fluvial or lacustrine environment. A fluvial association is tentative, however, because sedimentary structures are not preserved. Deposition must have been episodic, because the stratum contains a well-developed soil, one that consists of a heavily gleyed Btgb horizon. Implicit in the presumed sedimentary environment is a climate of more effective moisture than the present climate. In addition, the gleyed nature of the soil suggests ponded water at the surface or higher ground-water tables. Although no radiocarbon ages were obtained from Unit III, the character of the unit is consistent with other silty sands recognized elsewhere (e.g., Stafford 1, Edwards 2, Reno 4), demonstrated to be late Wisconsinan in age. As a result, Unit III at Stafford 5 is assigned a Woodfordian age.

Although the origin of the facies associated with Unit III is uncertain, the origins of Units II and I (figs. 39, 40) can be made with confidence. In general, the sediments in Units II and I consist of massive, single-grain sand that is much better sorted than the underlying deposits. The character of the deposits, coupled with their topographic position and expression, correlates very well with dune sand. Inherent in dune development is a relatively warm, arid environment, one promoting destabilization and eolian mobilization of sand. Soil development throughout Unit II, and at the surface of Unit I, indicate episodic sedimentation. Following deposition of Unit II, a brief period of pedogenesis occurred that resulted in the development of a soil with A/Bw horizonation. Although no radiocarbon ages were obtained from the 2Ab, its character (e.g., color, structure, position) suggests a middle to late Holocene age. Subsequent to development of the 2Ab, at least 1.40 m (4.6 ft) of eolian sand accumulated. The weakly developed surface soil, with A/AC/C horizonation (fig. 40), suggests that Stafford 5 has been stabilized for only a brief period of time.

Stafford 6

Stafford 6 is a 2.30-m (7.5-ft)-thick section (figs. 7, 41), exposed by backhoe trenching, in a parabolic dune field located in the SW NE sec. 29, T. 23 S., R. 12 W. (fig. 6). In general, the sediments at Stafford 6 are noncalcareous, loamy (with a mean particle size ranging from coarse silt to very fine sand), very poorly sorted to moderately sorted, very fine to finely skewed, and leptokurtic to very leptokurtic (Arbogast, 1995). Four pedostratigraphic units are recognized in the profile, with one in deposits rich in silt and three in surficial dune sand (figs. 41, 42). Essentially, these diachronous strata represent contrasting depositional regimes in the past 20,000 years that were punctuated by periods of soil formation.

Unit IV extends from 1.52 m (4.9 ft) to the base of the profile. The deposits are loamy and very poorly sorted because of high percentages of sand, silt, and clay (fig. 42). As a result, the sediments are extremely cohesive. In its character (e.g., texture, structure, color, stratigraphic position), Unit IV is similar to other silty sands (e.g., Edwards 1–4, Stafford 1 and 2) in the region. Unfortunately, unlike other localities, sedimentary structures are not preserved that would attest to the facies. The overall lack of sorting, coupled with a clear fining-upward sequence in the middle of the deposit, suggests a low-energy fluvial or lacustrine facies. A climate that was apparently more moist than the present is also implied. Although the age of Unit IV was not estimated through radiocarbon dating, the deposit is consistent with others (e.g., Edwards 1 and 2, Reno 4, Stafford 1) assigned a late Wisconsinan age and is considered to be a late Wisconsinan stratigraphic unit. Regardless of when and how the stratum accumulated, the formation of a strongly developed soil (fig. 42) indicates that the unit was exposed for some period of time. A rather intense period of erosion must have subsequently occurred, one that truncated the 4Ab that is presumed to have existed.

A sequence of relatively uncohesive, sandy sediments, extending from the surface to 1.52 m (4.9 ft), is included within Units I to III (figs. 41, 42). The overall character of these sediments—including very sandy texture, poor structure, relatively good sorting, color, and topographic position and expression—qualify them as wind-blown dune sand. Given no significant truncation, the sequence represents at least three cycles of rapid eolian sedimentation during a relatively warm, arid climate, followed by brief periods of soil formation during periods of more effective moisture. Following deposition of Unit III, a poorly developed soil formed. Radiocarbon ages of about 550 yr B.P. from the base of the 3ABb and 480 yr B.P. from the top of the 3Ab (fig. 42) suggest that sedimentation occurred in

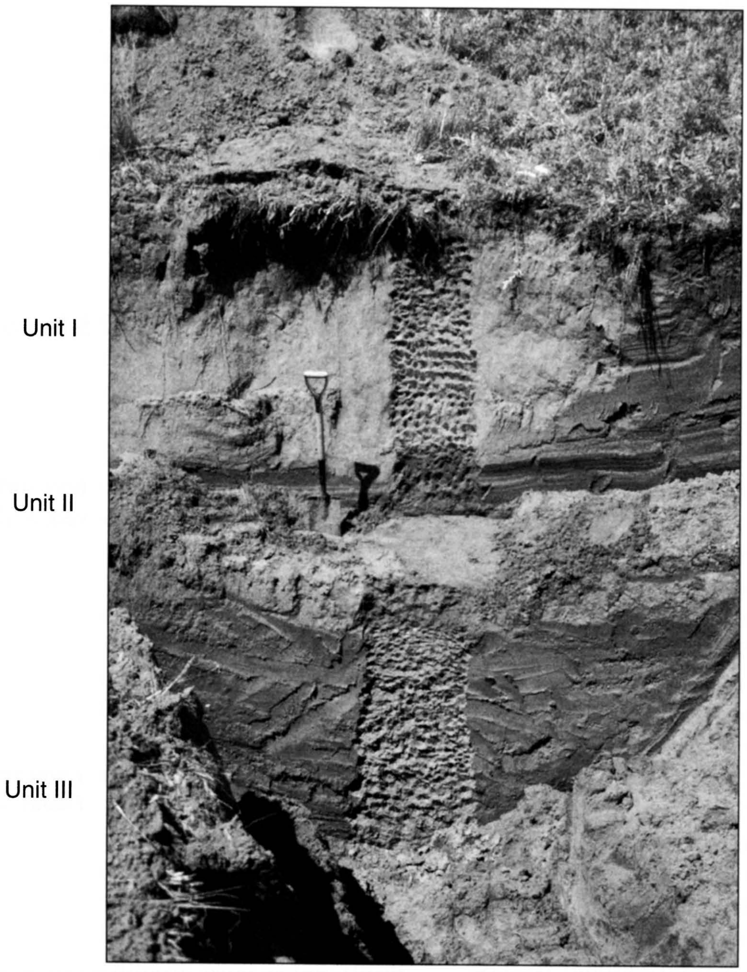


FIGURE 39—THE 3.46-M (11.4-FT)-HIGH EXPOSURE AT STAFFORD 5, showing the position of pedostratigraphic units.

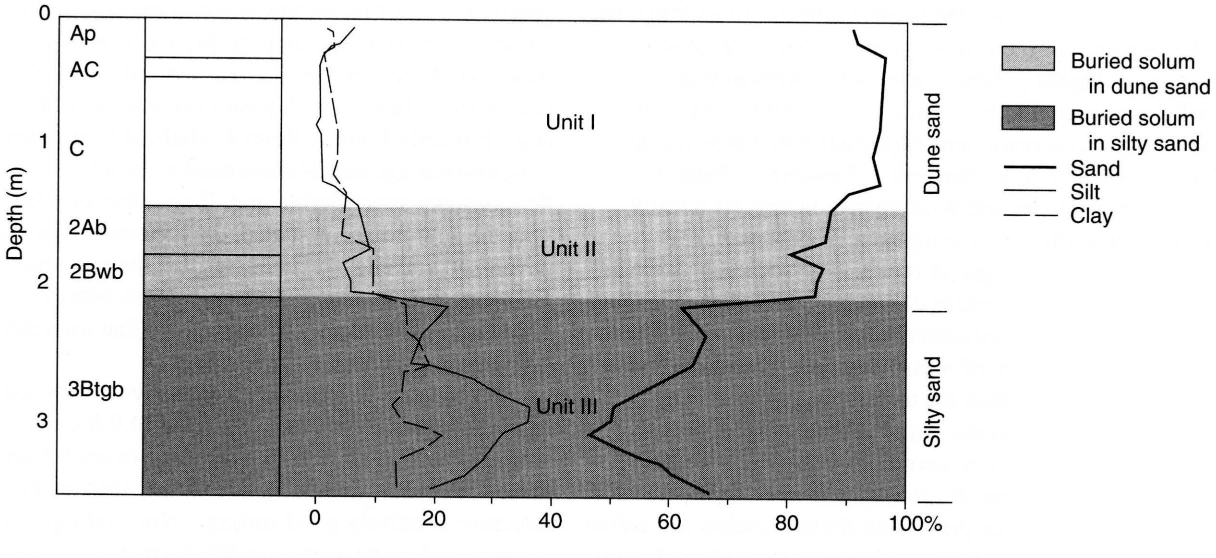


FIGURE 40—PEDOSTRATIGRAPHY, SOIL HORIZONATION, AND TEXTURE at Stafford 5.

the past 1,000 years and pedogenesis lasted for, at most, a few hundred years. Soon after this soil formed, at least 70 cm (28 in) of sediment accumulated that buried the 3Ab. Another brief period of soil formation subsequently occurred, one that altered the upper part of Unit II. A radiocarbon age of about 270 yr B.P. from the upper part of the 2Ab verifies a very late Holocene age for the overall deposit. Values of $\delta^{13}C$ are about 15.0‰ in both buried soils,

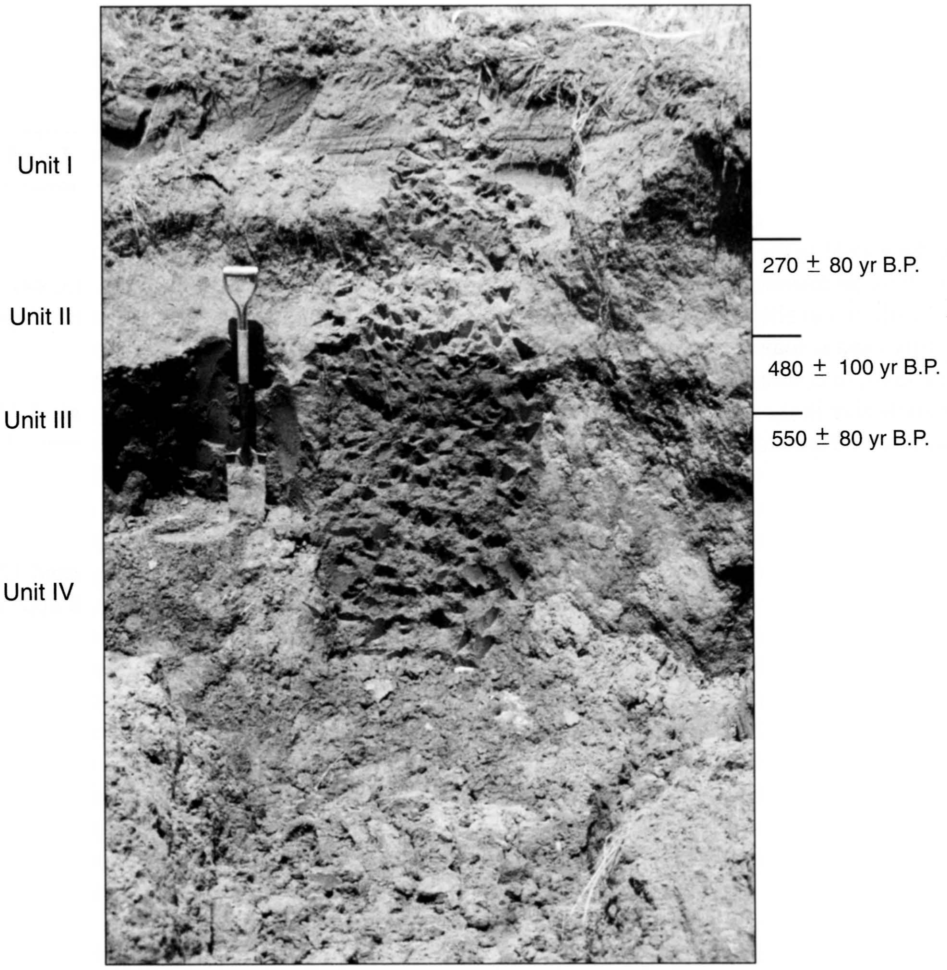


FIGURE 41—The 2.30-m (7.5-ft)-high exposure at Stafford 6, showing the position of pedostratigraphic units and radiocarbon ages.

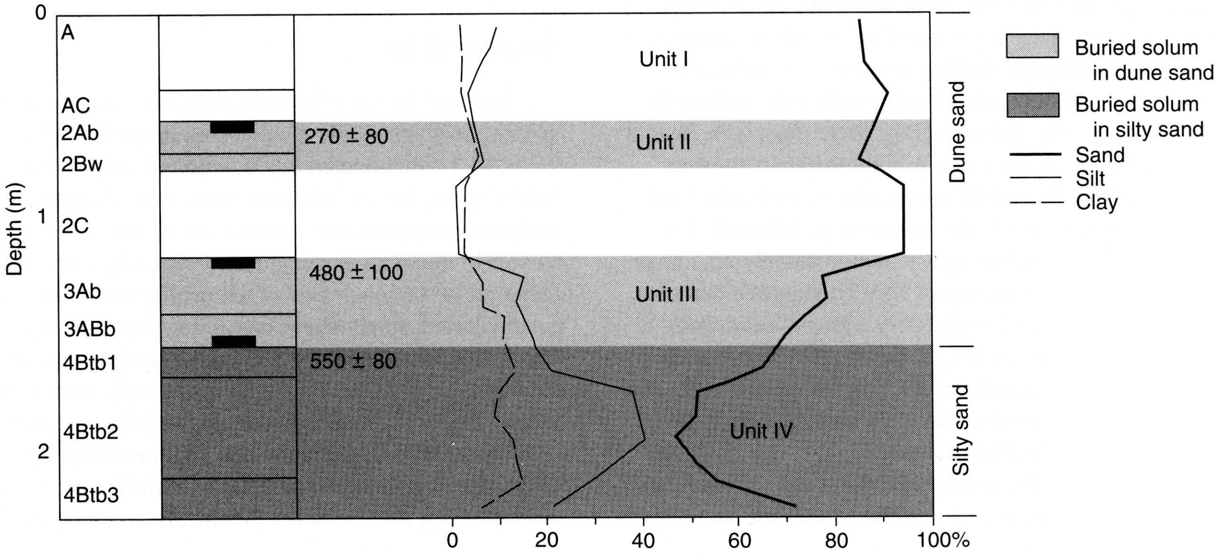


FIGURE 42—Pedostratigraphy, soil horizonation, and texture at Stafford 6.

suggesting C_4 plants dominated the site in the past 1,000 years. Subsequent to the development of the 2Ab, at least 53 cm (21 in) of eolian sand accumulated. The overall lack of surface soil development, which includes an A and AC horizon, coupled with the radiocarbon ages obtained from the underlying soils, suggests that Stafford 6 has been stable for a brief period of time.

Stafford 10

Stafford 10 is a 3.0-m (9.8-ft)-deep section (figs. 6, 43), exposed by backhoe trenching, in a compound subparabolic dune field in the SW SW sec. 11, T. 25 S., R. 15 W. (fig. 6). Four pedostratigraphic units were described in the profile, with two in silty sand and two in sand (figs. 43, 44). The silty sand is loamy, ranging from sandy loam to loam. Dune sand, in contrast, is classified as sand (fig. 44). Mean particle size in the silty sand ranges from coarse to fine silt, with two, distinct fining-upward sequences, and the deposit is very poorly sorted, fine to very finely skewed, and mesokurtic to very leptokurtic. In the dune sand, mean particle size is coarse silt to very fine sand, sorting is poor, and the deposit is very fine to finely skewed and platykurtic to very leptokurtic (Arbogast, 1995). In general, the sedimentological differences between the dune sand and silty sand appear to reflect changes in depositional environments due to a shift in climate at the site in the past 20,000 years. Both the silty sand and dune sand are basically noncalcareous.

In general, Units IV and III can be grouped together. Although each represents a unique period of deposition, they are sedimentologically similar because they are loamy, have relatively strong structure, and are very poorly sorted. The primary difference is color, with Unit IV being slightly oxidized. Both units are also gleyed to some degree, suggesting a period of sediment saturation. Although sedimentary features are not preserved, the presence of two, distinct fining-upward sequences indicates deposition in a fluvial or lacustrine environment of mixed energy. After each unit was deposited, intervals of long-term landscape stability occurred, resulting in extremely well developed buried soils with stacked Bt horizons (fig. 44). Subsequently, there must have been a period of erosion since soil A horizons are truncated. Overall, Units IV and III are consistent with silty sand described elsewhere in the region (e.g., Edwards 1–4, Reno 4). A radiocarbon age of approximately 20,230 yr B.P. obtained from the top of Unit III suggests that the underlying silty sand is probably a late Wisconsinan deposit. In addition, a $\delta^{13}C$ value of -20.1‰ suggests relatively high proportions of C_3 plants inhabited the site, providing evidence of a relatively cool climate with possibly more effective moisture.

Unconformably overlying these units is Unit II, a 12-cm (5-in)-thick deposit of stratified sand, silt, and clay (figs. 43, 44). Overall, the relationship of Units III and II suggests a mixed environment, one that promoted both erosion and deposition. The contact between Units III and II is very uneven, with small channels visible in cross section (fig. 45) that indicate flowing water of unknown energy. Following truncation of Unit III, Unit II apparently resulted from fluvial processes that culminated in deposition of a 2-cm-thick silty-clay drape toward the top of the unit. Given the late Wisconsinan age of the top of Unit III, it is theorized that truncation of Unit III and subsequent deposition of Unit II occurred sometime between approximately 20,000 and 15,000 yr B.P.

Overlying Units II to IV, ranging from the surface to 1.20 m (3.9 ft), is Unit I (figs. 43, 44). Over 90% sand (fig. 44), Unit I is consistent with other deposits in the region categorized as dune sand in topographic position and expression, color, structure, and texture of the stratum. In contrast to the underlying silty sand, Unit I probably was deposited during a period of increased aridity that promoted accumulation of wind-blown sand. Conflicting evidence exists as to the time of this deposition. Based on a maximum-limiting age of approximately 20,230 yr B.P. from the upper part of the silty sand 12 cm (5 in) below the base of the dune (figs. 43, 44), it appears that Unit I is a late Wisconsinan deposit. However, Unit II may have been a much thicker deposit that was eroded during the early or middle Holocene. Because the silty-clay drape in Unit II resisted further truncation, this erosion did not reach the top of Unit III. Subsequently, eolian processes promoted deposition of Unit I on the remnants of Unit II. The Holocene age of Unit I is supported by surface soil development. Although the soil is relatively well developed, consisting of A/Bw horizonation (fig. 44), it does not appear to reflect 20,000 years of landscape stability. In fact, equilibrium throughout the Holocene is difficult to imagine. More likely, Unit I is a middle to early late Holocene deposit that has been stable only during the late Holocene.

Stafford 11

Stafford 11 is a 2.90-m (9.5-ft)-deep section (figs. 7, 46), exposed by backhoe trenching, in the SW SE sec. 6, T. 21 S., R. 11 W. Located in a compound subparabolic dune field (fig. 6), the site contains three, noncalcareous pedostratigraphic units: a lower one in silty sand overlain by two in massive, single-grained sand (figs. 46, 47). Deposits in the lower part of the profile are loamy, very poorly sorted, finely skewed and very leptokurtic to leptokurtic. In contrast, overlying sandy deposits are classified as sand, are poorly to moderately sorted, finely to very finely skewed, and very leptokurtic to leptokurtic (Arbogast, 1995). In general, the sedimentological differences between the silty sand (Unit III) and the dune sand (Units II and I) reflect the shift from moist conditions during the late Wisconsinan to more arid conditions in the Holocene.

Unit III extends from 1.07 m to the base of the profile and is consistent in character with silty sand elsewhere (e.g., Edwards 1, 2, and 4; Reno 4) in the study area. Two radiocarbon ages were derived from the deposit. One of approximately 11,000 yr B.P. from the center of the unit suggests late Wisconsinan deposition. The other, about 1,500 yr B.P. from the top of the stratum (figs. 46, 47), implies exposure during the late Holocene. Based on $\delta^{13}C$ values of

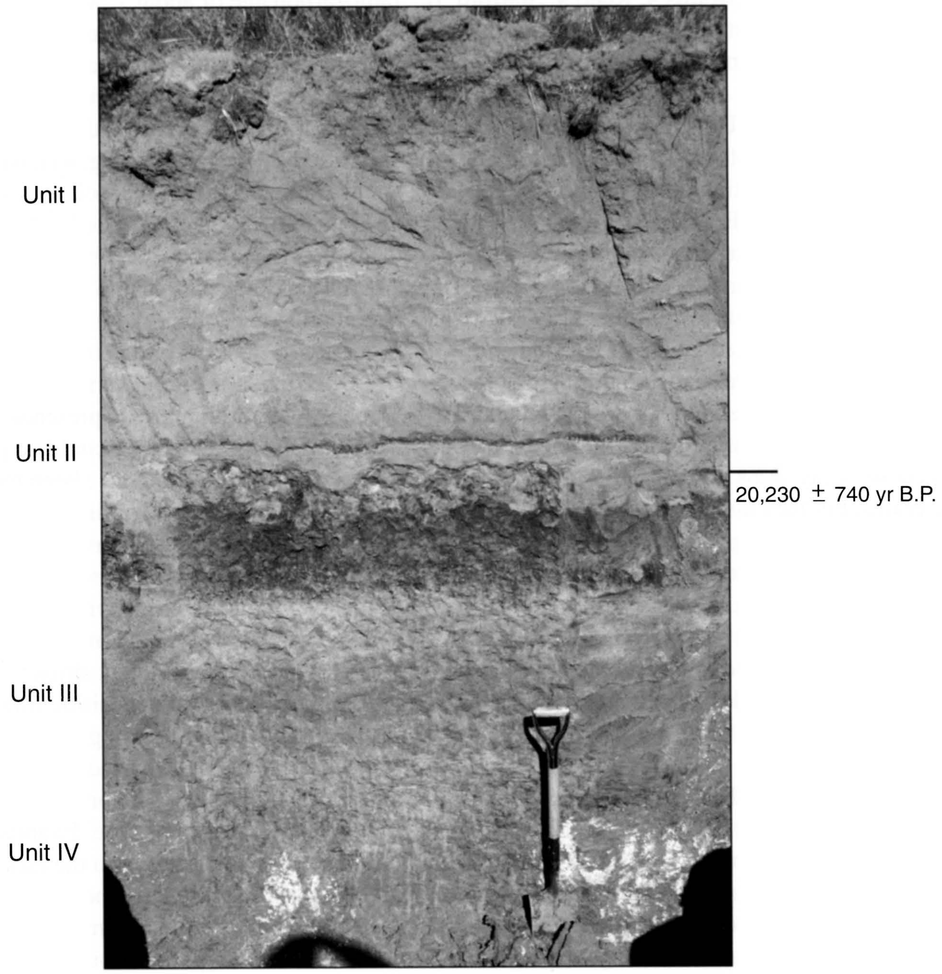


FIGURE 43—The 3.0-m (9.8-ft)-high exposure at Stafford 10, showing the position of pedostratigraphic units and radiocarbon age.

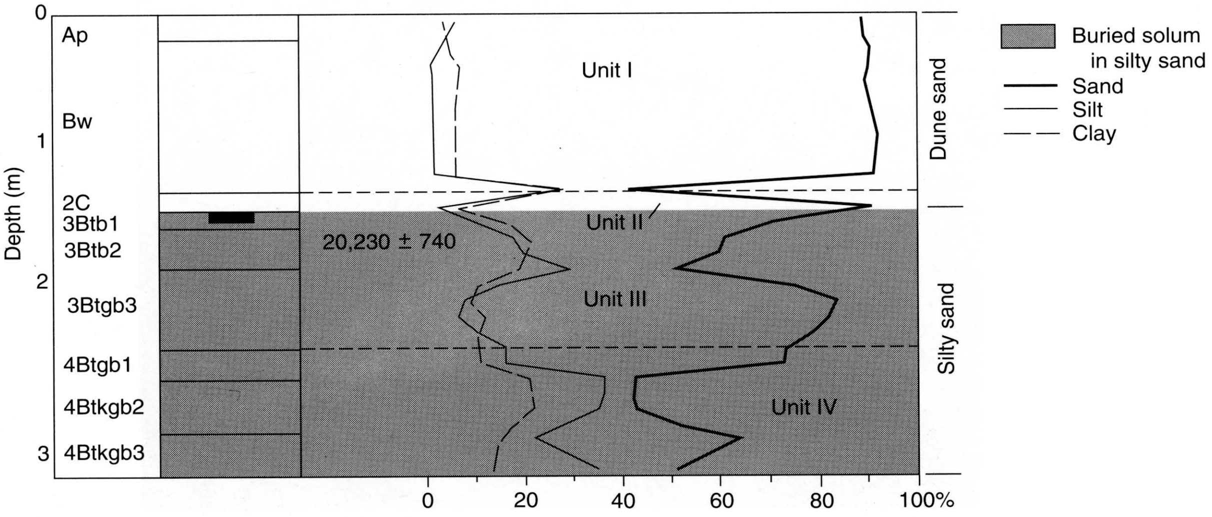


FIGURE 44—Pedostratigraphy, soil horizonation, and texture at Stafford 10.

about -16.0‰, C_4 plants have dominated the site for the past 11,000 years. Unit III is also loamy, containing high percentages of silt (ca. 57%) and clay (ca. 14.5%). Very poorly sorted, the stratum contains a well-defined fining-upward sequence similar to those described at other localities in the region (e.g., Edwards 4, Reno 4, Stafford 3). Although no sedimentary structures were discerned, the nature of the unit suggests deposition in a fluvial environment of mixed energy.

Following sedimentation, Unit III was evidently exposed and essentially stable for a very long period of time, and this promoted formation of a strongly developed buried soil consisting of five 3Btb horizons (fig. 47) with moderate to strong prismatic structure. In thin section, the matrix of the 3Btb3 horizon consists of skeleton grains of sand and silt surrounded by a clayey plasma. Mineralogy is dominated by quartz, but isolated grains of feldspar and rock fragments (e.g., biatite) are scattered throughout the section. The plasma is composed largely of striated clays, which indicate intensive illuviation of fine-grained material during pedogenesis. Clay skins are located primarily on the edge of sand grains, but they also line the inside of voids. Many of the sand grains are highly stained with iron, indicating oxidation. In addition, several sand grains are pitted, suggesting transport (fig. 48A).

Subsequent to development of the soil in Unit III, an erosional episode must have occurred, completely truncating the 3Ab horizon. Water tables must have been periodically high at this time, because the lower part of the 3Btb5 is slightly mottled.

Units II and I overlie Unit III with an apparent unconformity and sharp contact at 1.17 m (3.8 ft) (figs. 46, 47). Since they are similar in character, Units II and I can be grouped for purposes of discussion. Both, for example, contain over 90% sand that is poorly to moderately sorted (fig. 47). In addition, each contains a weakly developed soil with a mainly single-grain structure. Overall, they are consistent with dune sand in texture, sorting, structure, and topographic position and expression. In short, Units II and I reflect episodic sedimentation of eolian sand during a more arid climate than that in which Unit III was deposited. A radiocarbon age of about 1,500 yr B.P. from the top of Unit III (figs. 46, 47) provides a maximum-limiting, late Holocene age for wind-blown sand deposition. Following accumulation of Unit II, a brief period of landscape stability and soil formation occurred that resulted in development of a 2Ab and 2Bwb horizon (fig. 47). Although no radiocarbon age was obtained from this soil, it is similar in development and position with solums dated to less than 1,000 yr B.P. in the region (e.g., Crocket Cutbank, Reno 3 and 4, Stafford 2 and 6, Rice Roadcut). In thin section, the lack of pedogenic alteration in the 2Ab horizon is clear. The matrix consists primarily of skeleton grains of sand and, to a lesser extent, silt. Birefringent material surrounding the sand grains is probably clay. Although the presence of clay on sand grains implies illuviation, the clay could be present from previous weathering episodes and have been transported with the sand. Occasionally, sand grains are pitted and are lined with iron oxide. According to Ransom (personal communication, 1995), this evidence is inconsistent with the environment in which the grains are now found, further indicating transport. In addition, many of the grains are frosted (fig. 48B), which is consistent with eolian deposition.

Another period of deposition followed the development of the soil in Unit II, resulting in Unit I. Because Unit I is a very weakly developed surface soil, consisting of an Ap and C horizon (fig. 47), the site probably has been stable for a relatively brief period of time. Examination of the C horizon in thin section illustrates both the lack of post-depositional alteration of the sediments and verifies the brevity of the 2Ab. The matrix consists primarily of skeleton grains of sand and, to a lesser extent, silt. Quartz is the dominant

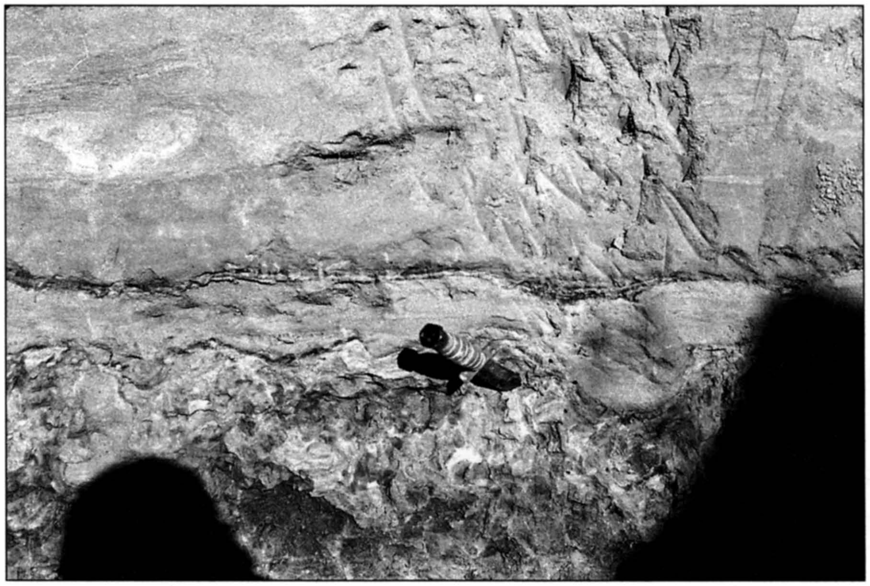


FIGURE 45—Truncated surface at the top of Unit III at Stafford 10. Note the knife for scale. The nails identify the boundaries of sedimentary beds.

mineral, although scattered grains of feldspar and rock fragments are present (fig. 48C). As in the 2Ab horizon in Unit II (fig. 48B), several of the grains are lined with clay. Given the single-grain structure of the horizon, this clearly suggests transport rather than illuviation of clay. Further evidence of transport lies in the many pitted grains covered with iron oxide and the frosting on many particles in the section (fig. 48C).

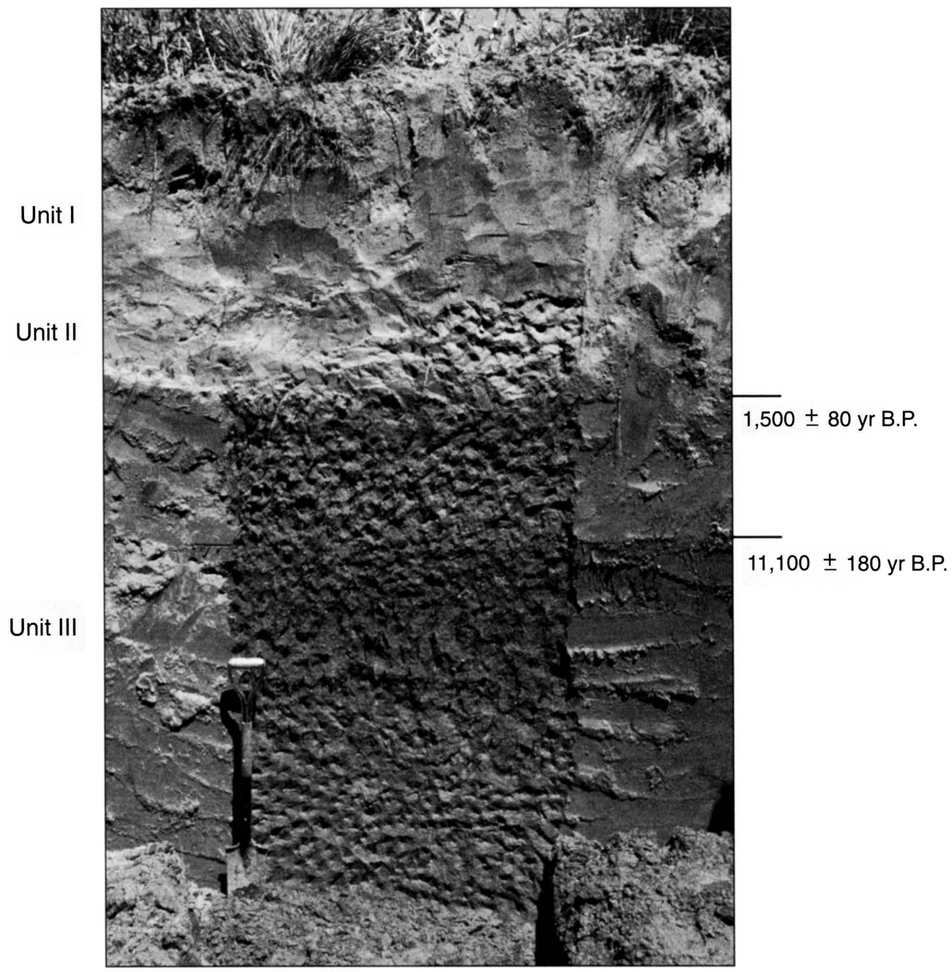


FIGURE 46—The 2.90-m (9.5-ft)-high exposure at Stafford 11, showing the position of pedostratigraphic units and radiocarbon ages.

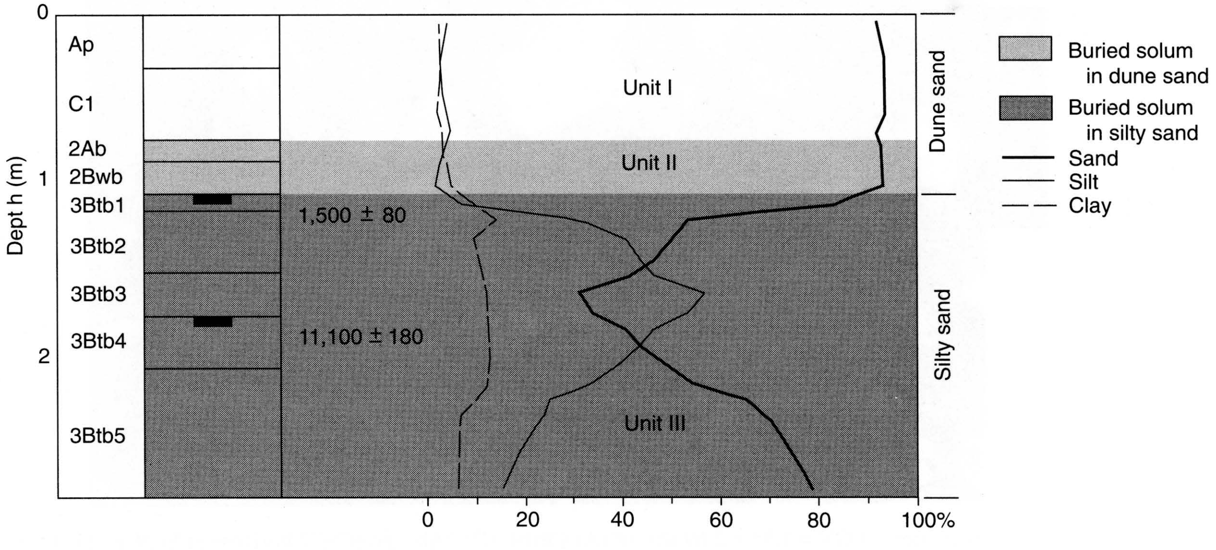


FIGURE 47—Pedostratigraphy, soil horizonation, and texture at Stafford 11.

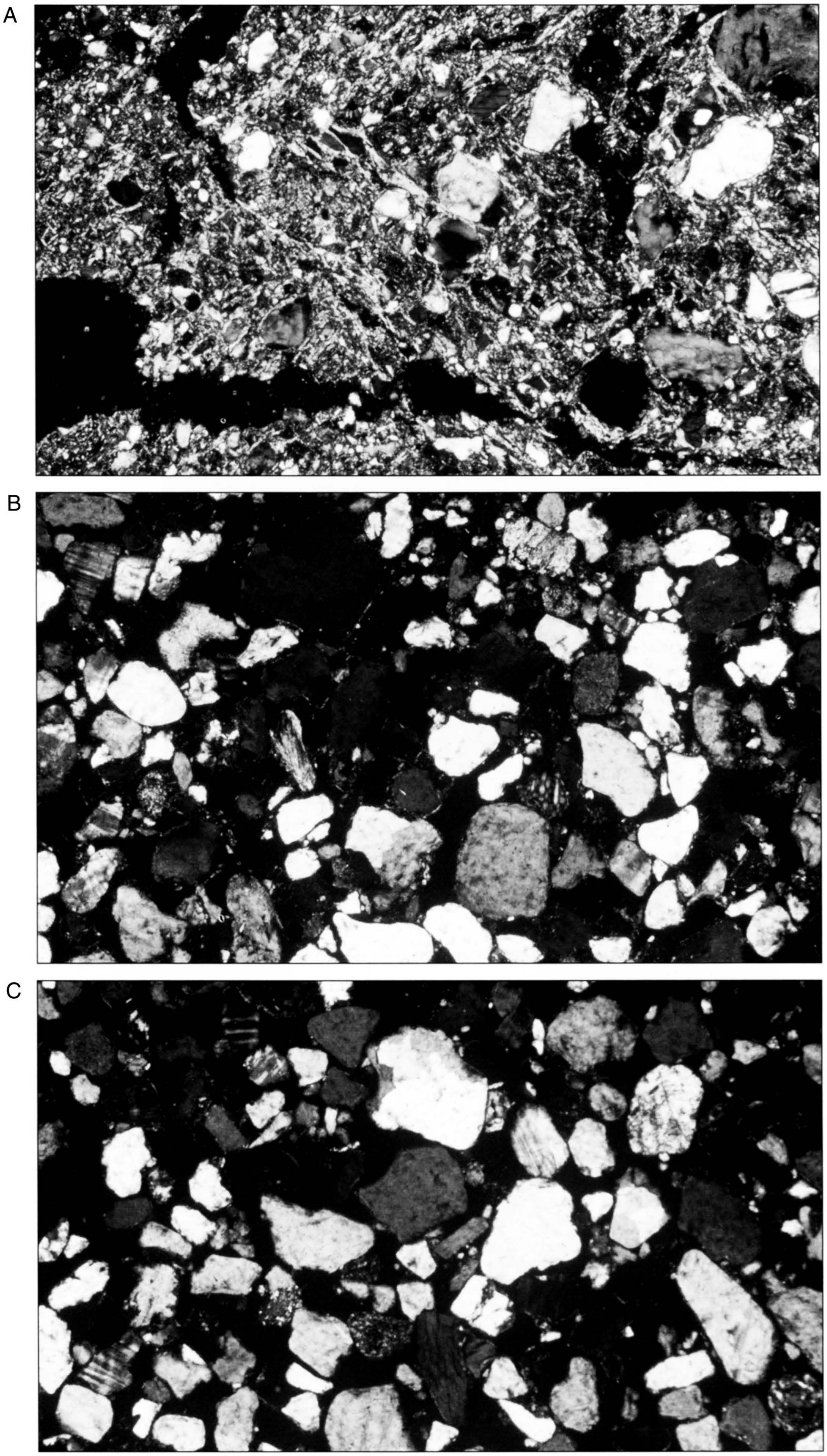


FIGURE 48—Thin sections (FOV = 1.85 × 2.69 mm) of (A) EBtb4, (B) 2Ab, and (C) C2 horizons at Stafford 11 (from Arbogast and Johnson, 1998).

Wilson Ridge

During the mapping phase of the project, a large anomalous dune was recognized in the west central part of the Great Bend Sand Prairie. The dune was unusual because of its size, relationship to other landforms, and orientation. Because of the potential for significant paleoenvironmental data that could supplement the entire study, detailed investigations were conducted and are presented below.

The site is located in the S1/2 secs. 28, 29, T. 25 S., R. 18 W. (figs. 6, 49) and was named Wilson Ridge after landowner Chad Wilson of Lewis, Kansas. In contrast to the majority of dunes in the study area, which are ≤100 m (328 ft) in length, Wilson Ridge is approximately 1.5 km (0.58 mi) long. In addition, the dune is concave in a northwesterly direction, rather than southwesterly as are other parabolic dunes in the region. When first discovered, the dune appeared to be related to a playa-lake basin immediately to the north (fig. 49). Preliminary investigations revealed a complex stratigraphy in the dune and a clear association with the playa. As a result, a hypothesis was formulated that the dune was a lunette, one that formed as sediment was deflated from the nearby lake bed during dry periods.

This lunette hypothesis was interesting because lunettes had not been formally recognized in the central Great Plains of Kansas and Nebraska. In fact, lunettes in the Great Plains had been previously identified only in the southern High Plains (Reeves, 1965, 1966; Holliday, 1985, 1989, 1995). Of further significance, however, were the chronologic and paleoenvironmental ramifications of Wilson Ridge to the study as a whole. If indeed the feature was a lunette, then it must have formed when northerly or northwesterly winds prevailed in order to scour the adjacent playa and construct the dune. Northwesterly winds are thought to have last dominated during the late Wisconsinan (COHMAP Members, 1988; Kutzbach et al., 1993). Therefore, a final hypothesis was constructed, stating that Wilson Ridge was a lunette that formed during the Woodfordian when northwesterly winds prevailed. To test this theory, stratigraphic information was obtained from six backhoe trenches excavated along a south to north transect across the dune axis. In order to correlate stratigraphy among trenches, twelve cores were extracted with a Giddings coring machine (fig. 50).

According to Arbogast (1996a), 10 late Quaternary stratigraphic units were identified through backhoe trenching and coring, and their chronology was established with 12 radiocarbon ages (table 3). Several, well-developed buried soils were recognized, at various positions in cross section (fig. 51), that formed in sediments ranging from clay loam to sandy loam (fig. 52).

The basal unit (Unit I), which was recognized in trench 6 (fig. 53) and cores 2 to 4, 6 to 9, and 11 (figs. 51–52), underlies the lunette where it is about 3 m (9.8 ft) higher than in the playa. The unit consists of gleyed silt and clay containing fragmented gastropods. Humates at the top of Unit I in trench 6 dated to about 17,500 yr B.P. and provided a $\delta^{13}C$ value of -20.8‰.

Unit II was recognized in trenches 4 (figs. 51–53) and 5 (figs. 52, 54), as well as in cores 7 to 10. The unit contains numerous gastropod fragments and is extremely

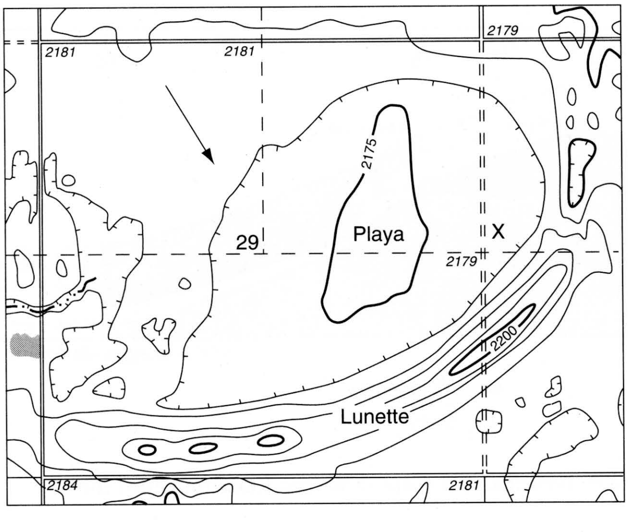


FIGURE 49—Topographic map in the vicinity of Wilson Ridge. Scale = 1:24,000 (Centerview 7.5-min. quadrangle, 1972). Pursuant to the landowner's request, the study transect was located where the unimproved road bisects the dune from north to south (x). The arrow reflects the direction of the prevailing winds that scoured the playa during arid intervals.

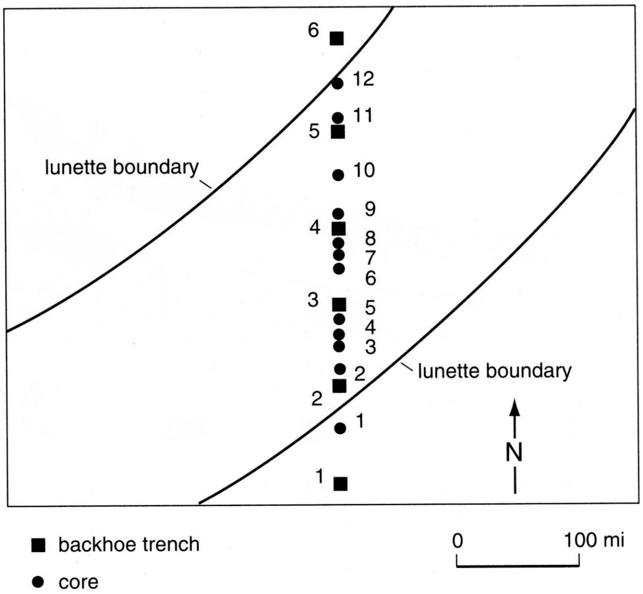


FIGURE 50—Position of backhoe trenches and cores extracted from Wilson Ridge (modified from Arbogast, 1996a).

calcareous; it is thickest (ca. 2 m; 6.6 ft) in the center of the dune, but pinches out or is truncated both up and downslope (fig. 51). Formed throughout the deposit is a well-developed buried soil, consisting of several Btb horizons. Humates in the upper part of Unit II dated to about 16,500 yr B.P. in trench 4 and approximately 17,200 yr B.P. in trench 5 (figs. 52, 55). In addition, $\delta^{13}C$ values of -11.9‰ were obtained from both samples.

Unit III consists of calcareous, sandy sediments, containing fragmented gastropods, and overlies Unit II in the center and north slope of the dune. The deposit, which has a maximum thickness of about 2.6 m (8.5 ft), was recognized in trenches 4 (figs. 51, 52, 55) and 5 (figs. 52, 54) as well as in cores 6 to 11. It appears to have been truncated on both the north and south slopes (fig. 51). Although no radiocarbon ages were obtained directly from Unit III, ages of approximately 17,200 and 9,800 yr B.P. obtained from trench 4 (figs. 52, 55) effectively bracket the deposit.

Capping Units I, II, and III is Unit IV, which is recognized in trenches 1 through 3 (figs. 56–58) and in cores 1 through 8. Unit IV consists of pale-brown

TABLE 3—RADIOCARBON AGES OBTAINED FROM WILSON RIDGE. All ages were obtained on the base soluble fraction of total soil humates.

Backhoe trench	Depth (m)	Laboratory number	Uncorrected ^{14}C age (yr B. P.)	$\delta^{13}C$ (‰)	Corrected ^{14}C age (yr B.P.)	Calibrated age[1] (yr B. P.)
Trench 1[2]	2.45–2.50	Tx-8005	10,220 ± 200	-18.3	10,330 ± 200	12,520(12,200)11,550
Trench 2[2]	1.18–1.23	Tx-7827	5,490 ± 140	-13.6	5,670 ± 140	6,640(6,450)6,310
Trench 3[2]	1.59–1.64	Tx-7996	8,680 ± 180	-14.2	8,860 ± 180	9,990(9,890)9,570
Trench 3[2]	2.65–2.70	Tx-7826	10,180 ± 200	-14.2	10,360 ± 200	12,550(12,240)11,670
Trench 3[2]	2.91–2.96	Tx-8004	12,000 ± 220	-18.1	12,110 ± 240	14,550(14,130)13,760
Trench 4[2]	2.60–2.65	Tx-7825	16,330 ± 380	-11.9	16,520 ± 400	20,070(19,450)18,960
Trench 5[2]	0.95–1.00	Tx-8014	9,460 ± 180	-13.0	9,840 ±180	11,660(11,000)10,870
Trench 5[2]	2.25–2.30	Tx-7824	16,950 ± 460	-11.9	17,180 ± 480	21,137(20,370)19,611
Trench 6[2]	0.73–0.78	Tx-8002	2,600 ± 100	-12.4	2,800 ± 100	3,020(2,870)2,770
Trench 6[2]	1.13–1.18	Tx-7695	3,220 ± 80	-13.2	3,410 ± 80	3,822(3,677)3,478
Trench 6[2]	1.94–1.99	Tx-7997	7,410 ± 160	-15.6	7,560 ± 160	8,440(8,340)8,130
Trench 6[2]	3.58–3.63	Tx-7995	17,480 ± 520	-20.8	17,540 ± 260	21,620(20,880)20,070

[1] Calibration from a conventional $\delta^{13}C$-corrected radiocarbon age to calibrated calendar years using a tree-ring curve. All calibrations reported here were based upon the 20-year atmospheric curve (see Linick et al., 1985, 1986; Kromer et al., 1986; Mook, 1986; and Stuiver et al., 1986). Program used for calibration is discussed in Stuiver and Reimer (1993).

[2] Reported in Arbogast (1996a).

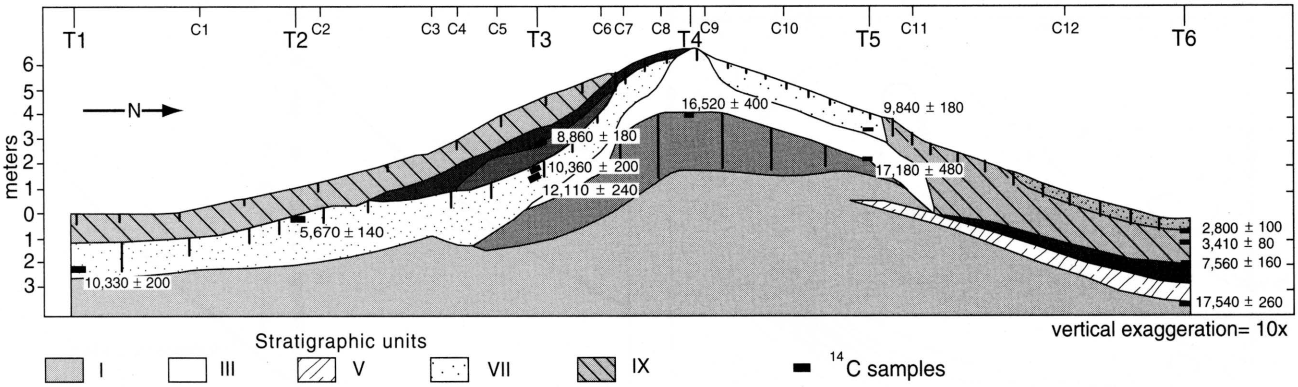


FIGURE 51—CROSS SECTION SHOWING THE POSITION OF STRATIGRAPHIC UNITS, TRENCHES, AND CORES AT WILSON RIDGE (modified from Arbogast, 1996a).

(10YR6/3; moist) silt and has a maximum thickness of about 1.4 m (4.6 ft). Inset against Unit III in the center of the dune, Unit IV overlies Unit I on the south slope. The stratum contains a well-developed soil recognized in most of the dune's south slope that merges with the surface soil at the dune crest (fig. 51). Humates in the upper part of the 3Ab in trench 3 dated to about 12,100 yr B.P. and provided a $\delta^{13}C$ value of -18.1‰.

Unit V includes laminated deposits of sand and silt (fig. 51) recognized in the playa in trench 6 (fig. 53) and in cores 11 and 12 on the dune's north slope. The maximum thickness of the unit is about 1.5 m (4.9 ft) (fig. 51). Although no radicarbon age was derived from the stratum, ages of about 17,500 and 7,600 yr B.P. in trench 6 (figs. 52, 53) bracket the deposit.

Unit VI is a lens of laminated silt and sand that was identified on the south flank of the dune in trench 3 (figs. 52, 58) and in cores 4 to 6. In trench 3 the deposit is about 1.4 m (4.6 ft) thick, but it thins both up and downslope (fig. 51). Sedimentary structures (e.g., small cross-beds,

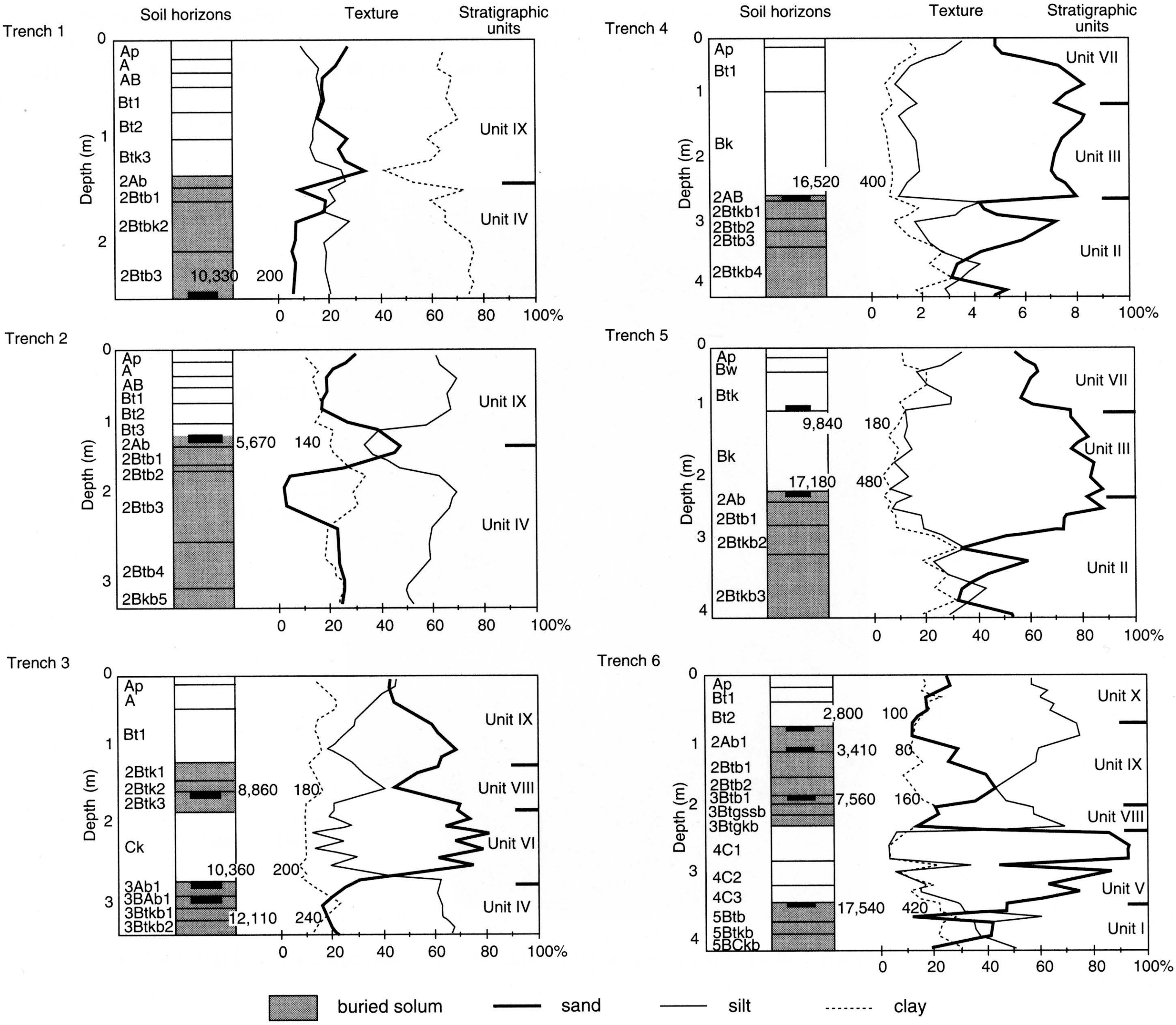


FIGURE 52—TRENCH AND SOIL STRATIGRAPHY, including radiocarbon ages, at Wilson Ridge (modified from Arbogast, 1996a).

convolutions) suggest that the sediment was saturated when it accumulated, possibly related to slumping (fig. 59). Many of the laminae are consistent in character (e.g., texture, color) with the underlying soil, suggesting it may have been a source for sediment. An age of about 10,400 yr B.P. ($\delta^{13}C$ value: -14.2‰) from the upper part of the 3Ab in trench 3 is a maximum-limiting age for Unit VI, and an age of about 8,900 yr B.P. ($\delta^{13}C$ value: -14.2‰) from the lower part of the overlying 2Btk3 gives the minimum-limiting age (figs. 51, 58). These ages, coupled with the character of the deposit, further suggest rapid accumulation. Given this scenario, Unit VI separates two buried soils in trench 3 that are otherwise welded upslope at the surface in cores 7 and 8 and downslope in trenches 1 and 2 and core 1.

Unit VII, a lens of sandy silt and sand, mantles Unit III on the upper north slope of the dune (fig. 51). Observed in trench 5 (figs. 52, 55) and cores 9 and 10 (fig. 51), it is similar to the underlying Unit III, but does not contain gastropods. As a result, it may indicate a different depositional interval. Humates from the lower part of the Btk horizon of Unit VII in trench 5 dated to about 9,800 yr B.P. and provided a $\delta^{13}C$ value of -13.0‰.

Unit VIII is a thin veneer of silty sediment on the north slope of the dune in trench 6 and core 12 (figs. 52, 53) and on the south slope of the dune in trench 3 and cores 3 to 7 (fig. 51, 58). In the playa, the deposit is about 42-cm (16.5-in)-thick where it overlies Unit V. Humates from the upper part of the 3Btb1 horizon in trench 6 provide a minumum-limiting age of around 7,600 yr B.P. and a $\delta^{13}C$ value of -15.6‰. The

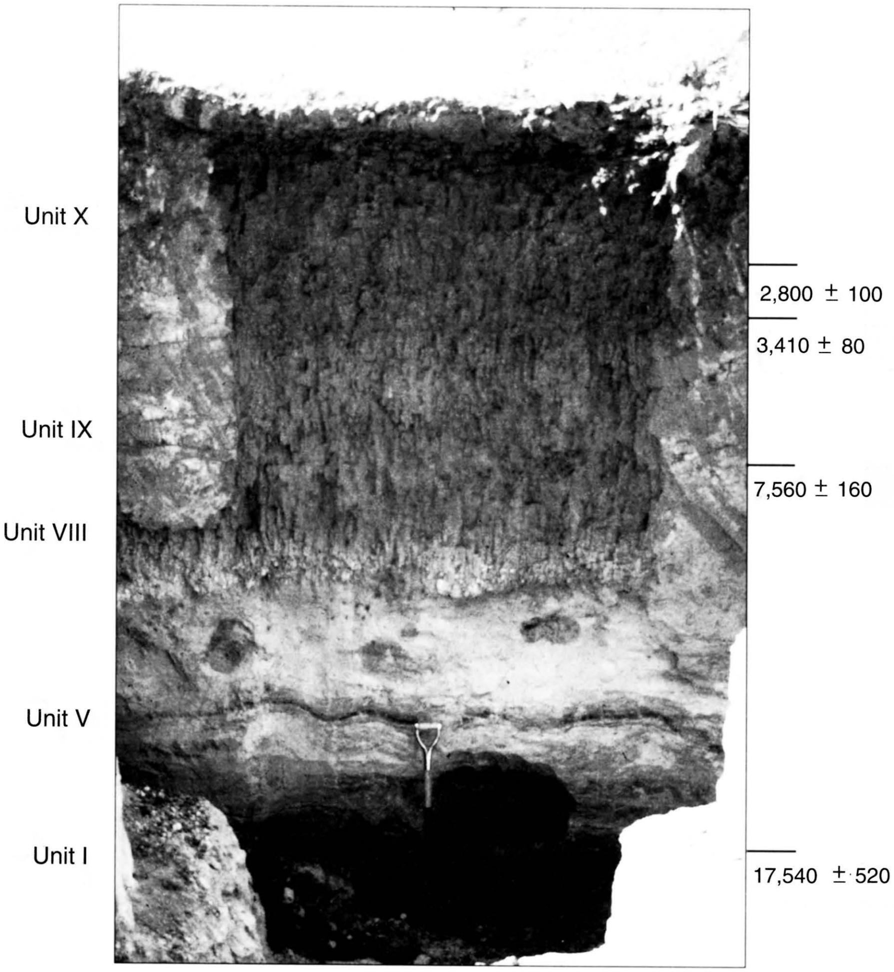


FIGURE 53—Trench 6, 4.20 m (13.78 ft) deep, at Wilson Ridge showing stratigraphic units and radiocarbon ages.

age of the unit can be better estimated on the south slope of the dune, where it is about 1.5 m (4.9 ft) thick, is capped by a truncated buried soil, and overlies the stratified deposits of Unit VI. Radiocarbon ages of about 8,900 yr B.P. ($\delta^{13}C$ value: -14.2‰) at the base of the deposit in trench 3 (figs. 51, 58), and approximately 5,700 yr B.P. ($\delta^{13}C$ value, -13.6‰) from the upper 2Ab in trench 2 (fig. 51, 57) provide maximum- and minimum-limiting ages, respectively, for Unit VIII on the south slope of Wilson Ridge.

Unit IX is a deposit of silty sediment on both the north and south flanks of the dune (figs. 51, 52). On the north side, the unit was recognized in trench 6 (figs. 52, 53) and cores 11 and 12 (figs. 51). The deposit is thickest (1.21 m; 3.9 ft) in trench 6, where it is bracketed by radiocarbon ages of about 7,600 and 3,400 yr B.P. On the south side of the dune, the stratum was recognized in trenches 1 to 3 and cores 1 to 6 where it has an average thickness of about 1.2 m (3.9 ft) (figs. 51, 52). A radiocarbon age of approximately 5,700 yr B.P. was derived from the 2Ab in trench 2 (figs. 52, 57) and provides a maximum-limiting age for Unit IX on the south side of the dune.

Unit X is a thin veneer of sediment, positively identified in the playa within trench 6 (figs. 52, 53). The unit is a 73-cm (28.7-in)-thick deposit of silt (fig. 52) that buried a well-developed soil soon after 2,800 yr B.P. Unit X may thinly mantle most of the dune, however, as suggested by the textural and pedogenic consistency (i.e., thermic, pachic Argiustoll) that is present, except for the north slope (Arbogast, 1996a).

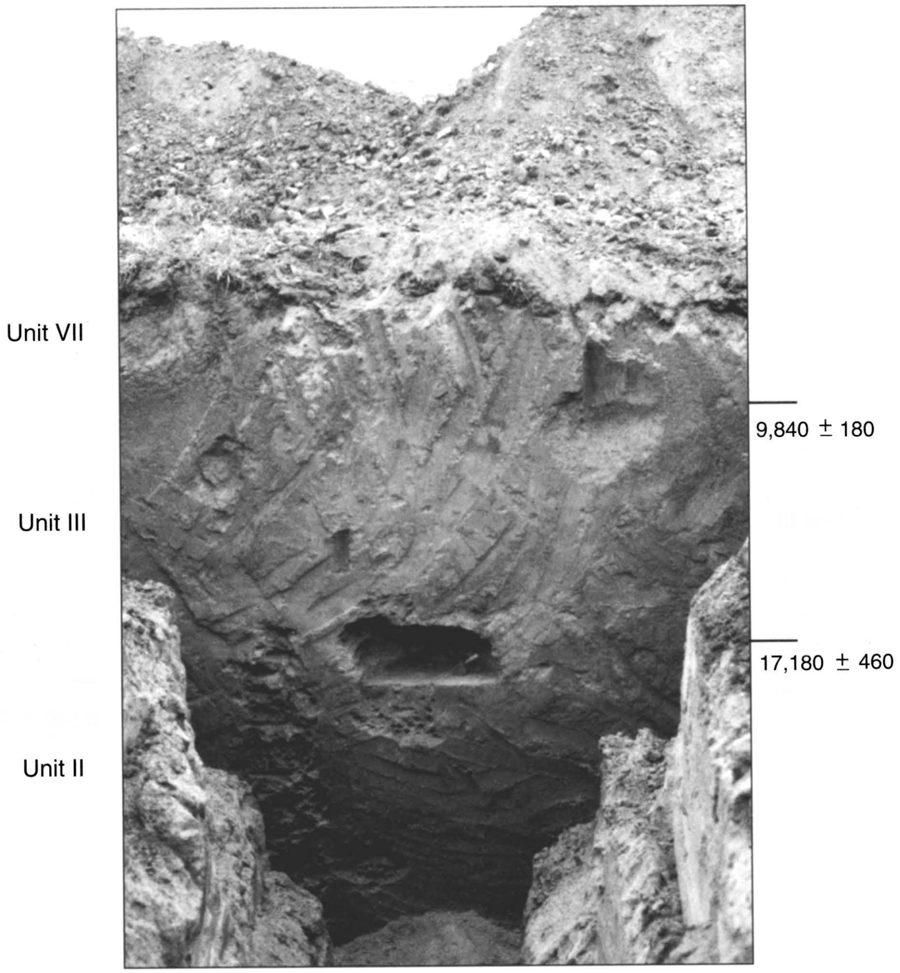


FIGURE 54—TRENCH 5, 4.18 M (13.71 FT) DEEP, at Wilson Ridge showing stratigraphic units and radiocarbon ages.

In summary, the stratigraphic and chronologic evidence indicate a complex late Quaternary history at Wilson Ridge (Arbogast, 1996a). Ten stratigraphic units, containing a total of six buried soils, are present and represent a variety of depositional events and facies. Based on the unique characteristics (e.g., sedimentology, radiocarbon ages) of each unit, the geomorphic history can be reconstructed (Arbogast, 1996a). In addition, radiocarbon-derived $\delta^{13}C$ values (Krishnamurthy et al., 1982) and faunal remains can be used to infer paleoclimatic variability through time.

From the data currently available, it is impossible to precisely determine when Wilson Ridge began to develop. As with other lunettes in the Great Plains (e.g., Reeves, 1965; Holliday, 1985, 1989), the dune is closely associated with an adjacent playa (fig. 49). The underlying deposit (Unit I; fig. 51) is consistent with other late Wisconsinan deposits on the Great Bend Sand Prairie (e.g., Edwards 1–4, Reno 4). A radiocarbon age of about 17,000 yr B.P. from a buried soil in Unit I provides an estimated minimum age for the deposit. According to Arbogast (1996a), this correlates reasonably well with studies conducted elsewhere in the Great Plains. Holliday (1985), for example, presented evidence that lunettes began to form in the southern High Plains around 30,000 yr B.P. In addition, Reeves (1965) obtained a radiocarbon age of approximately 19,000 yr B.P. directly beneath a lunette in Texas. Given that Wilson Ridge overlies a unit that is texturally and chronologically consistent to the one described by Reeves (1965), eolian sedimentation probably began at Wilson Ridge about 20,000 yr B.P. This age estimate also correlates with the onset of Peoria loess sedimentation in

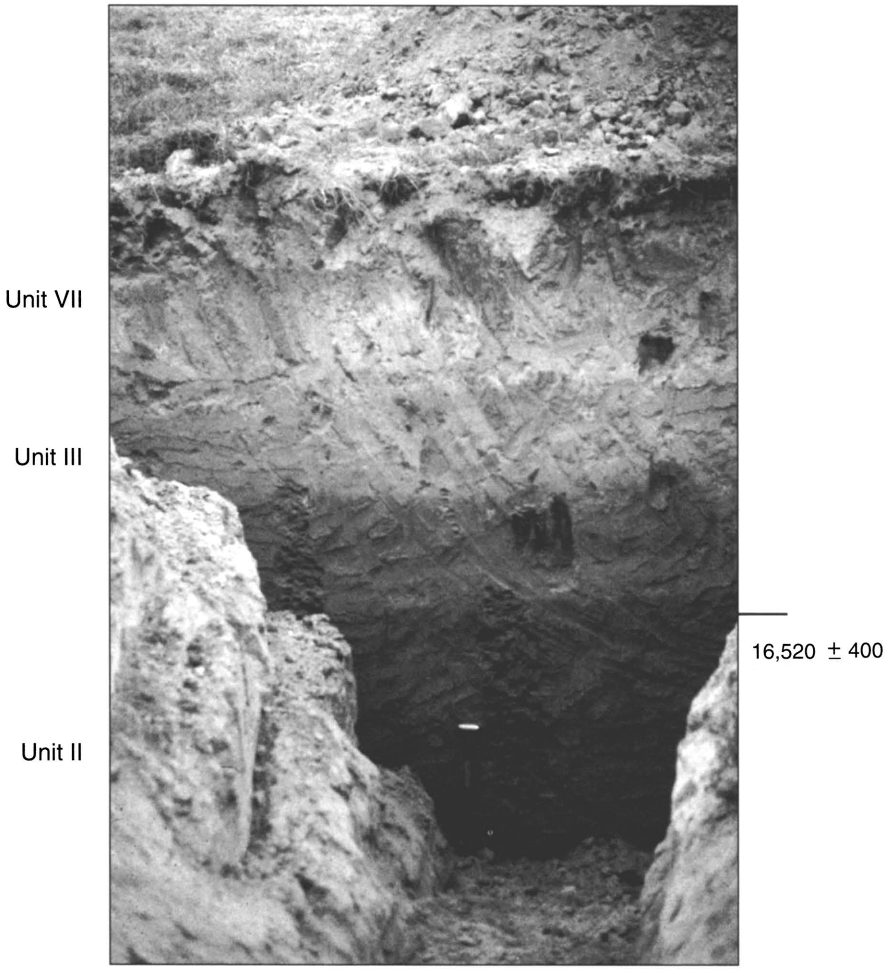


FIGURE 55—Trench 4, 4.22 m (13.85 ft) deep, at Wilson Ridge showing stratigraphic units and radiocarbon age.

the central Great Plains (Wells and Stewart, 1987; Johnson et al., 1993). Eolian deposition at Wilson Ridge apparently began because the playa was deflated and Unit II accumulated downwind of a small ridge or shoreline formed in Unit I.

From the available evidence it appears that deposition of Unit II continued, as a result of prevailing northwesterly winds, until approximately 17,000 yr B.P. The sediment is calcareous and includes numerous gastropod fragments that were probably transported from the playa. Following this interval of sedimentation, the playa stabilized, resulting in a well-developed soil in both the lake bed and on the dune. Values of $\delta^{13}C$ suggest vegetation edaphically varied during this interval of time: -20.9‰ from the playa indicates a higher proportion of C_3 plants, whereas -11.9‰ demonstrates that the well-drained dune was xeric.

Soon after the episode of stability of 17,000 yr B.P., another interval of eolian sedimentation transpired. According to Arbogast (1996a), this period of deposition lasted until around 12,000 yr B.P., largely occurred in two stages, and may have been driven by the climate changes associated with the major late Wisconsinan deglaciation (Ruddiman, 1987). The initial stage resulted in Unit III, a deposit of calcareous and sandy sediment on the north slope and crest of the dune that deflated from the playa. The latter interval is represented by Unit IV, which accumulated on the south slope of the dune. Unit IV is similar in texture (70% silt) and color (10YR6/3; moist) to Peoria loess, a unit of late-Wisconsinan loess (Wells and Stewart, 1987; Johnson, 1993, Johnson et al., 1993) that has been recognized immediately to the north and south of the Great Bend Sand Prairie (Feng, 1991; Feng et al.,

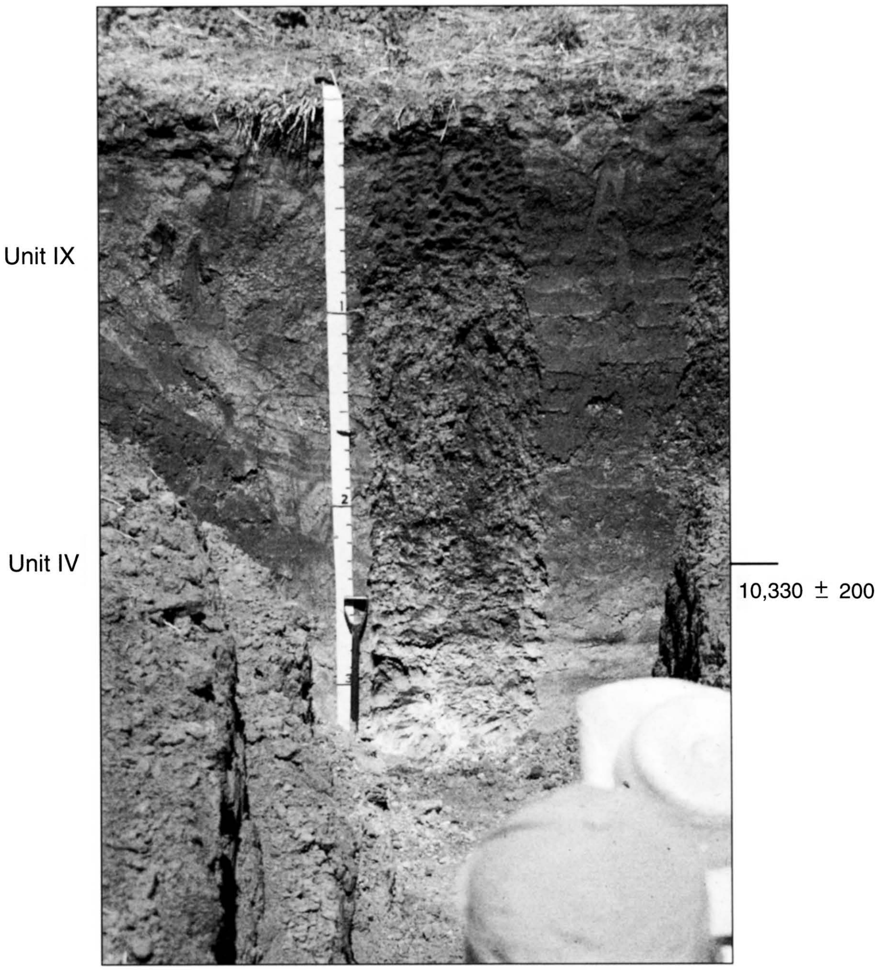


FIGURE 56—Trench 1, 3.31 m (10.86 ft) deep, at Wilson Ridge showing stratigraphic units and radiocarbon age.

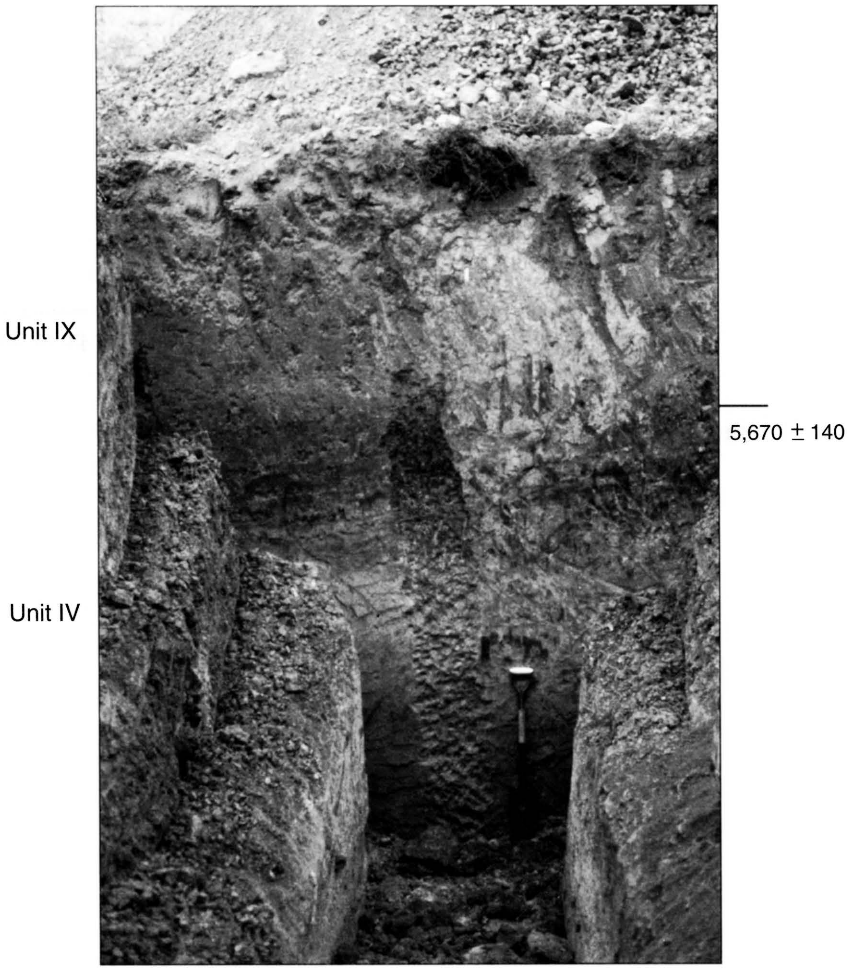


FIGURE 57—Trench 2, 3.22 m (10.56 ft) deep, at Wilson Ridge showing stratigraphic units and radiocarbon age.

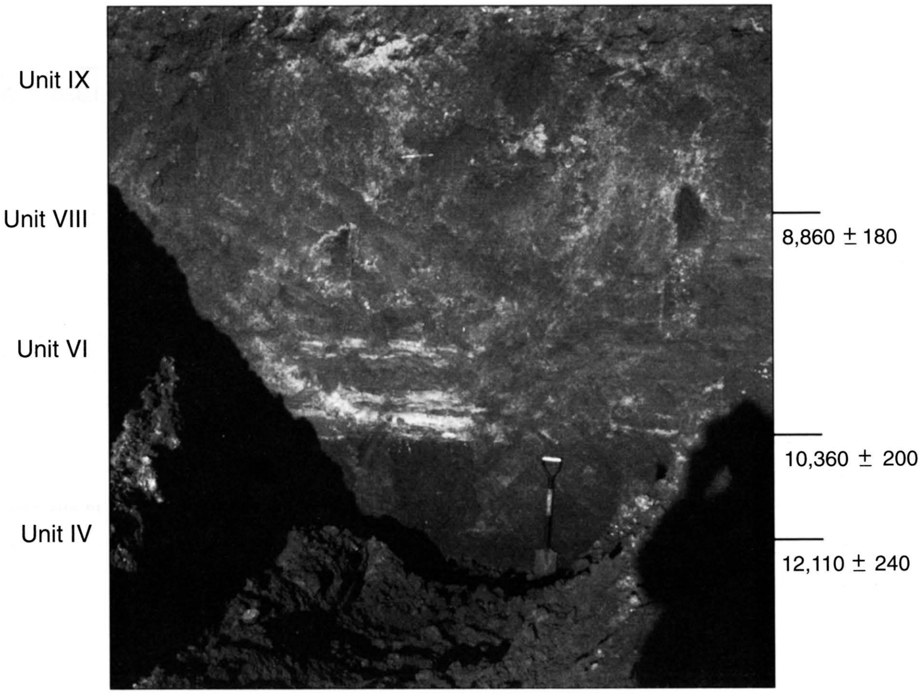


FIGURE 58—Trench 3, 3.24 m (10.63 ft) deep, at Wilson Ridge showing stratigraphic units and radiocarbon ages.

1994). As a result of this similarity to Peoria loess, it is conceivable that Unit IV may have originated from a distant source.

Subsequent to the deposition of Units III and IV, a well-developed soil formed in Unit IV on the upper part of the dune's south slope (fig. 51). Radiocarbon ages from the lower and upper boundaries of the soil suggest that landscape stability may have lasted for approximately two thousand years (i.e., from about 12,000 to 10,000 yr B.P.). This interval of pedogenesis implies a temporal correlation with the Brady soil, a geosol that represents a major period of landscape stability in the central Great Plains (Schultz and Stout, 1948; Frye and Leonard, 1951; Caspall, 1970, 1972; Feng et al., 1994) at the Pleistocene-Holocene climatic boundary (Johnson and May, 1992). Generally, the interval of time represented by the Brady geosol is thought to range from approximately 10,500 to 8,500 yr B.P. The vast majority of ages derived from the Brady, however, are from northern Kansas (e.g., Johnson, 1993) and Nebraska (e.g., Johnson and May, 1992; Johnson et al., 1993). According to Arbogast (1996a), the results from Wilson Ridge suggest that the Brady may be time-transgressive.

Given that the soil (Brady?) in Unit IV is so well developed, it is interesting that it occurs only on the south slope of the dune. This spatial variability may have occurred for one of two reasons: (1) a soil formed only on the lee slope of the dune because strong northwestly winds perpetually destabilized the north slope; or (2) a soil initially formed over the entire dune, but was truncated on the north slope during the early Holocene.

According to Arbogast (1996a), four variables suggest that the north slope of the dune was extensively eroded during the early Holocene: (1) the soil on the south slope is extremely well developed, suggesting a major period of landscape stability; (2) the preserved soil dips more steeply than the modern surface (moreover, it merges with the surface soil on the crest, suggesting the dune was higher at one time with the soil continuing to the north slope); (3) values of $\delta^{13}C$ shift from -18.1‰ in the lower boundary to -14.2‰ in the upper boundary of the soil on the south slope of the dune, which suggests a shift to a more arid and potentially unstable environment towards the end of the soil-forming interval; and (4) the soil was partially truncated on the lower part of the lee slope of the dune.

The nature of early Holocene deposits suggests that a significant period of landscape instability occurred soon after about 10,000 yr B.P., resulting in complete truncation of the Brady soil on the north slope and its partial truncation on the south slope. Specifically, the position and sedimentology of Unit VI (fig. 51) imply at least one major erosional event in the early Holocene. As mentioned previously, Unit V consists of well-laminated deposits of sand and silt, containing convolutions and load structures, that buried the soil in Unit IV. A radiocarbon age of approximately 8,900 yr B.P. was obtained from sediments directly above Unit VI. Coupled with the maximum-limiting age of about 10,000 yr B.P. (from the top of Unit IV), this provides a relatively narrow window (~1,000 yr) for the deposition of Unit VI to have occurred. Overall, the width of the depositional interval and the deposit's sedimentology indicate rapid accumulation. Stringers of organic-rich laminae in the unit, apparently derived from the underlying soil, suggest that mass wasting of upslope sediment truncated the soil near the crest of the dune. As a

FIGURE 59—Sedimentary structures, including laminations and load structures, in Unit VI in Trench 3 at Wilson Ridge.

result, the dune probably decreased in height and the position of the crest shifted slightly from north to south.

During the Holocene the geomorphic chronology is characterized by episodic deposition in a periodically more arid, destabilizing environment. Evidence suggests that the playa remained the sediment source during the early Holocene. A major unconformity exists between units V and VIII, suggesting that strong, northwestly winds deflated the lake bed. As a result, the north slope and crest of the dune would have frequently been destabilized, whereas the southern slope was protected. Accumulation of sandy sediment (Unit VII) began approximately 9,000 yr B.P. on the north slope of the dune. The onset of this sedimentary interval correlates with events recorded elsewhere in the region, including deposition of Bignell loess (Frye et al., 1968; Feng, 1991; Johnson, 1993) and localized eolian erosion in the southern High Plains (Holliday, 1989). A δ^{13}C value of -13.0‰ from Unit VII suggests increased aridity at the site during the early Holocene. In addition, the unit lacks the gastropods that are contained in underlying deposits, further suggesting increased aridity. Early Holocene deposition on the north slope was probably episodic because a well-developed soil formed in the playa around 7,500 yr B.P. This period of stability correlates reasonably well with soil formation in a playa located to the west in Haskell County (Mandel and Olson, 1995).

Despite the instability of the north slope, evidence suggests that the south slope was largely in equilibrium during the early Holocene. Other than deposition of Unit VI and some erosion of the Brady soil, extended pedogenesis prevailed. On the upper part of the south slope, a period of soil formation began soon after deposition of Unit VI. This period of stability lasted until about 5,800 yr B.P. Immediately south of the dune, an extended period of pedogenesis (~10,300–5,800 yr B.P) resulted in the development of a thick, cumulic soil.

It is suprising that the south slope of Wilson Ridge was stable throughout the early Holocene while the north slope was severely truncated, especially since regional and local evidence implies a more unstable environment. Documented early Holocene climate change (Webb, 1985; Crowley and North, 1991) reflects a gradual temperature increase, which culminated in the middle Holocene as an extended warm, dry period (Wright, 1970; Benedict and Olson, 1978; Barry, 1983; Holliday, 1989). This period is commonly referred to as the Altithermal (Antevs, 1955). Values of δ^{13}C, derived from the buried soil on the south side of Wilson Ridge (fig. 51), reflect the regional record, changing from -18.3‰ at about 10,300 yr B.P. to -13.6‰ around 5,800 yr B.P. Given the overwhelming evidence for a shift to more arid conditions, the south slope of Wilson Ridge should have destabilized in the early Holocene. According to Arbogast (1996a), the only logical explanation is that mobilizing winds remained northwesterly into the latter part of the early Holocene, resulting in a "windshadow" that protected the south slope.

According to Arbogast (1996a), Wilson Ridge has generally been a stable landform in the past 5,000 yr B.P. Local climate has generally remained warm and dry, as indicated by low δ^{13}C values (e.g., -12.4‰). Sedimentation in the playa (Unit IX) has been episodic with soil formation occurring between about 3,500 yr B.P. and 2,800 yr B.P. This pattern correlates to a period of relative stability, one that began around 4,000 yr B.P., in a playa in Haskell County (Mandel and Olson, 1995). On the south slope of the dune, an approximately 1-m-thick deposit of sandy silt (Unit IX) has accumulated in the past 5,000–6,000 years. This corresponds to the record derived from a lunette in the Sandhills of Texas, where eolian sedimentation occurred between about 6,000 yr B.P. and 4,500 yr B.P. (Holliday, 1989). In contrast to earlier deposits, which were mobilized by northwestly winds at Wilson Ridge, sedimentation on the south slope (Unit IX; fig. 51) apparently resulted from the same southerly winds that were documented by Arbogast (1996b) elsewhere on the Great Bend Sand Prairie in the late Holocene. At most localities across the dune and playa, the surface soil is well developed, indicating relatively long-term (ca. 1,000 yr) stability at the site. The lone exception is on the upper north slope where the surface soil is less well developed, suggesting that this area of the dune has recently been active and that northwestly winds continue to provide sediment (Arbogast, 1996a).

Discussion and Conclusions

Chronology of Late Quaternary Landform Evolution and Paleoenvironmental Change

After analyzing and integrating the data collected from the Great Bend Sand Prairie, a full reconstruction of late Quaternary landform evolution and paleoenvironmental change on uplands is possible. Given the lack of reliable dating techniques, earlier studies of dune fields in Kansas were largely qualitative in nature and based upon outdated glacial chronologies. Because of the high number of radiocarbon ages obtained in this study (tables 3, 4), previously accepted theories regarding landscape evolution in the region have been significantly revised. The chronology is organized in three parts: late Wisconsinan (ca. 21,000–10,000 yr B.P.), early to middle Holocene (ca. 10,000–4,000 yr B.P.), and late Holocene (4,000 yr B.P.–Present).

Late Wisconsinan (ca. 21,000–10,000 yr B.P.)

According to Arbogast and Johnson (1998), data obtained from late Wisconsinan deposits imply that the environment was more mesic than during the Holocene. The only macrofloral sample (white-spruce charcoal) of Woodfordian age indicates the presence of conifers during the glacial maximum (Johnson, 1991). The assemblage of faunal macrofossils (e.g., *Discus cronkhitei, Succinea avara,* an aquatic bivalve) further suggests that the environment was mesic because these species are presently extant only in the subalpine taiga of the Rocky Mountains (Leonard, 1952; Bequaert and Miller, 1973) and the more humid southeastern and northeastern parts of North America (Leonard, 1952) (fig. 60). Finally, values of $\delta^{13}C$ derived from Woodfordian deposits generally range from -25‰ to -17‰, suggesting that C_3 plants were more prevalent (fig. 61).

During the Woodfordian the cool, relatively moist environment apparently promoted accumulation of a very poorly sorted, generally noncalcareous deposit of sand, silt, and clay (silty sand) on the Great Bend Sand Prairie. A total of 13 radiocarbon ages were derived from the lower part of silty sand (i.e., sample locations buried by additional deposits of silty sand) exposed in backhoe trenches. Eleven of these ages range from about 24,000 yr B.P. to 10,000 yr B.P. (fig. 62). The extremely widespread silty sand underlies most of the study area but varies considerably in texture from a spatial perspective. At some localities (e.g. Edwards 4), the silty sand is consistently 60% sand, whereas at others (e.g., Reno 4, Belpre Trench) the deposits contain as much as 80% silt or 40% clay. The majority of study sites exhibit at least one, but typically two or three, fining-upward sequences. Overall, the spatial variability in the silty sand suggests lateral differences in facies. Fining-upward sequences within the silty sand at specific sites imply that subtle change in depositional regime also occurred on a temporal basis (Arbogast and Johnson, 1998).

It is difficult to determine with certainty the process which resulted in accumulation of the silty sand. At Cheyenne Bottoms, Fredlund (1995) reported a major unconformity between Farmdalian and Holocene deposits, which he argued occurred because climate was more arid and surface winds were stronger in the Woodfordian than in the Holocene. In general, results from this study contradict Fredlund's (1995) conclusions because the silty sand is so pervasive and demonstrably Woodfordian in age. Unfortunately, no diagnostic sedimentary structures were discerned in the silty sand in any of the sites studied.

There is indirect evidence, according to Arbogast and Johnson (1998), that the silty sand may be an eolian deposit. Specifically, the high percentages of silt in much of the silty sand suggest an eolian source, at least in part, because substantial deposits of Peoria loess accumulated in the central Great Plains during the Woodfordian (Frye and Leonard, 1952; Wells and Stewart, 1987; Johnson, 1993; Johnson et al., 1993), including sites immediately north (Barton County Landfill) and south (Pratt County Landfill) of the study area (Feng, 1991; Feng et al., 1994). Given the documented presence of Peoria loess surrounding the study area, deposition of eolian silt probably occurred on the Great Bend Sand Prairie as well (Arbogast and Johnson, 1998). In fact, Peoria loess may exist in scattered localities in the region, as suggested by deposits of buff-colored, relatively calcareous silt recognized at the Belpre Trench, Phillips Trench, and the south flank of Wilson Ridge. At the Belpre and Phillips Trenches, the silt overlies a buried soil considered to be equivalent to the Gilman Canyon Formation (Reed and Dreeszen, 1965), based upon a radiocarbon age of about 21,000 yr B.P. from the Belpre Trench. In addition, landsnails comparable to those described elsewhere in the Peoria loess (e.g., Leonard, 1952; Wells and Stewart, 1987) were recovered from the deposits at both localities.

Arbogast and Johnson (1998) also suggested that the silty sand may also, in part, be an eolian deposit similar to cover sands in Alaska (Lea and Waythomas, 1990) and Europe (e.g., Koster, 1988; Schwan, 1988), which accumulated downwind of glacial outwash streams. Given the position of the Arkansas River generally to the north of the Great Bend Sand Prairie and the prevailing northwesterly winds of the Woodfordian, the floodplain certainly could have been a source for eolian sediment.

Although eolian processes probably contributed to the silty sand, Arbogast and Johnson (1998) argued that the majority of evidence suggests that fluvial processes potentially dominated the Woodfordian depositional regime. The silty sand differs from cover sand in its lack of stratification and its very high clay content at some sites. Additionally, the surface of the unit is sharply and very unevenly truncated at two sites (e.g., Edwards 3, Stafford 10), suggesting that flowing water was responsible for erosion. These characteristics, coupled with the many fining-upward sequences and very poor sorting (Friedman, 1967), implied to Arbogast and Johnson (1998) that sedimentation largely occurred in a low-energy fluvial system. Given the textural variability in the silty sand

TABLE 4—RADIOCARBON AGES USED TO ESTIMATE THE AGE OF BURIED SOILS IN SILTY SAND, LOESS, AND DUNE SAND ON THE GREAT BEND SAND PRAIRIE. All ages from soils were obtained on the base soluble fraction of total soil humates, except for Tx-6479*, which was derived from charcoal. Abbreviations: ds = dune sand; uss = upper silty sand; lss = lower silty sand.

Location	Stratigraphic unit	Horizon (depth)	Lab no. (Tx-)	Uncorrected ^{14}C age	δ^{13}C-corrected ^{14}C age[1]	δ^{13}C (‰)	Calibrated age[2]
Buster Dune[3]	ds	2Ab (0.82–0.87)	7980	modern	modern	-17.9	modern
Harvey/Reno[3]	ds	2Ab (1.47–1.52)	7828	40 ± 100	210 ± 100	-14.2	320(160)60
Stafford 6[3]	ds	2Ab (0.53–0.58)	7983	110 ± 80	270 ± 80	-15.2	464(299)146
Rice Roadcut[3]	ds	2Ab (1.28–1.33)	7978	230 ± 80	380 ± 80	-16.0	504(467)315
Stafford 6[3]	ds	3Ab (1.13–1.18)	7982	320 ± 100	480 ± 100	-15.0	560(510)320
Rice Roadcut[3]	ds	2Ab (1.69–1.74)	7977	340 ± 80	490 ± 80	-15.9	559(517)454
Stafford 6[3]	ds	3Ab (1.47–1.52)	7981	360 ± 80	550 ± 80	-13.4	644(542)507
Reno 3[3]	ds	3Ab (2.57–2.62)	8119	520 ± 80	700 ± 80	-14.4	701(657)554
Reno 4[3]	ds	2Ab (1.10–1.15)	8012	510 ± 80	710 ± 80	-12.4	717(660)556
GMD 5 #2[4]	uss	2Ab (1.63–1.68)	9743	660 ± 60	810 ± 120	-15.5	910(700)650
Crocket Cutbank[3]	ds	3Ab (5.88–5.93)	8214	750 ± 80	880 ± 80	-16.6	913(774)695
Stafford 2[3]	ds	2Ab (1.14–1.19)	9313	840 ±80	1,030 ± 80	-13.3	1,056(936)792
14KW7[3]	ds	2Ab (5.12–5.17)	7777	920 ± 60	1,090 ± 120	-14.5	1,170(970)910
Stafford 11[3]	uss	3Btb1 (1.07–1.12)	8218	1,350 ± 80	1,500 ± 80	-15.7	1,502(1,354)1,301
Buster Quarry[3]	ds	3Ab (4.55–4.60)	7979	1,350 ± 100	1,500 ± 100	-15.8	1,510(1,350)1,300
GMD5 #7[4]	uss	2Ab (1.02–1.07)	6744	1,440 ± 160	1,620 ± 160	-14.7	1,700(1,520)1,340
GMD 5 #10[4]	uss	2Ab (1.43–1.48)	7700	2,130 ± 140	2,250 ± 140	-17.3	2,360(2,230)2,020
Cornwell Trench[3]	ds	2Ab (1.48–1.53)	7998	2,140 ± 100	2,310 ± 100	-14.3	2,380(2,340)2,150
Phillips Trench[3]	uss	2Btb1 (0.33–0.38)	8216	2,250 ± 100	2,400 ± 130	-15.8	2,710(2,360)2,330
Edwards 4[3]	uss	3Btb1 (1.60–1.65)	8314	2,610 ± 90	2,730 ± 180	-17.3	3,080(2,790)2,710
GMD 5 #10[4]	uss	2Ab (1.43–1.48)	6745	2,810 ±160	2,940 ± 160	-17.0	3,450(3,070)2,750
Edwards 1[3]	uss	2Btb1 (1.15–1.20)	8003	1,690 ±80	3,220 ± 80	-13.2	3,555(3,425)3,349
Crocket Cutbank[3]	ds	3C5 (8.06–8.11)	8215	3,160 ± 50	3,280 ± 100	-17.7	3,630(3,470)3,380
Cullison Quarry[3]	uss	3Btb1 (4.70–4.75)	8221	3,660 ± 100	3,820 ± 100	-14.8	3,822(3,677)3,478
GMD 5 #7[4]	uss	2Ab(2.02–2.07)	6877	4,530 ± 160	4,620 ± 160	-16.6	4,400(4,180)4,000
GMD 5 #9[4]	uss	2Ab (1.30–1.35)	6477	4,680 ± 200	4,840 ± 200	-15.2	6,290(6,180)5,990
Reno 4[3]	uss	3Btb1 (2.27–2.32)	8011	5,260 ± 120	5,370 ± 120	-17.7	7,020(6,720)6,390
GMD 5 #9[4]	uss	2Ab (1.40–1.45)	6478	5,740 ± 300	5,870 ± 300	-17.6	7,210(7,020)6,860
Stafford 3[3]	ds	2Ab (1.96–2.01)	8118	6,050 ± 160	6,160 ± 160	-17.9	7,900(7,690)7,580
Belpre Trench	uss	Bt3 (0.95–1.00)	8009	6,850 ± 140	6,930 ± 140	-20.2	8,440(8,340)8,130
Cullsion Quarry[3]	lss	4Btgb1 (5.65–5.70)	8220	7,770 ± 140	7,850 ± 140	-20.0	8,950(8,560)8,780
Belpre Trench	lss	Btk8 (2.20–2.25)	7975	8,490 ± 160	8,550 ± 160	-21.0	9,840(9,490)9,370
Edwards 1[3]	lss	3Btb1 (1.65–1.70)	7697	8,990 ± 560	9,050 ± 560	-22.2	10,883(10,000)9,453
Reno 5	lss	4C1 (3.10–3.15)	8013	8,910 ± 260	9,160 ± 260	-9.5	10,387(10,040)9,905
Stafford 2[3]	lss	4Btb (3.08–3.13)	8117	10,440 ± 720	10,460 ± 720	-23.6	13,096(12,370)10,964
Stafford 11[3]	lss	3Btb4 (1.82–1.87)	8219	10,970 ± 180	11,100 ± 180	-16.5	13,230(13,010)12,810
Stafford 4	lss	2Ab (1.66–1.71)	7999	12,640 ± 340	12,820 ± 340	-14.2	15,710(15,150)14,560
Edwards 3	lss	4Btkb (2.72–2.77)	8115	13,280 ± 780	13,290 ± 780	-24.7	16,840(15,870)14,690
GMD 5 #1[4]	uss	2Ab (2.49–2.54)	6742	13,620 ± 560	13,670 ±1,340	-20.9	17,939(16,380)14,389
Edwards 3	lss	3Btkb (1.79–1.84)	8116	15,260 ± 680	15,250 ± 680	-25.6	18,836(18,170)17,429
Reno 4[3]	lss	3Btb3 (3.50–3.55)	8010	16,590 ±360	16,670 ±360	-19.9	20,260(19,450)18,960
Stafford 1	uss	2Btb1 (0.98–1.03)	7679	16,760 ±440	16,810 ± 440	-21.9	20,593(19,830)19,213
GMD 5 #9[4]	lss	2Bk2 (2.02)	6479*	17,950 ±660	17,970 ±660	-23.7	22,300(21,450)20,510
Edwards 2	uss	2btb1 (1.23–1.28)	7698	19,370 ± 1,000	19,460 ±1,000	-19.8	na
Stafford 10	lss	3Btb (1.45–1.50)	8114	20,150 ± 740	20,230 ± 740	-20.1	na
Belpre Trench	lss	Btkg13 (3.44–3.49)	7976	20,620 ± 500	20,670 ± 500	-22.3	na

[1] For a discussion of the δ^{13}C-correction procedure, see Stuiver and Polach (1977) and Taylor (1987).

[2] Calibrated from conventional δ^{13}C-corrected radiocarbon age to calendar years using a tree-ring curve. All calibrations reported here were based upon the 20-year atmospheric curve (e.g., Linick et al., 1985; Stuiver et al., 1986). Program used is discussed in Stuiver and Reimer (1993).

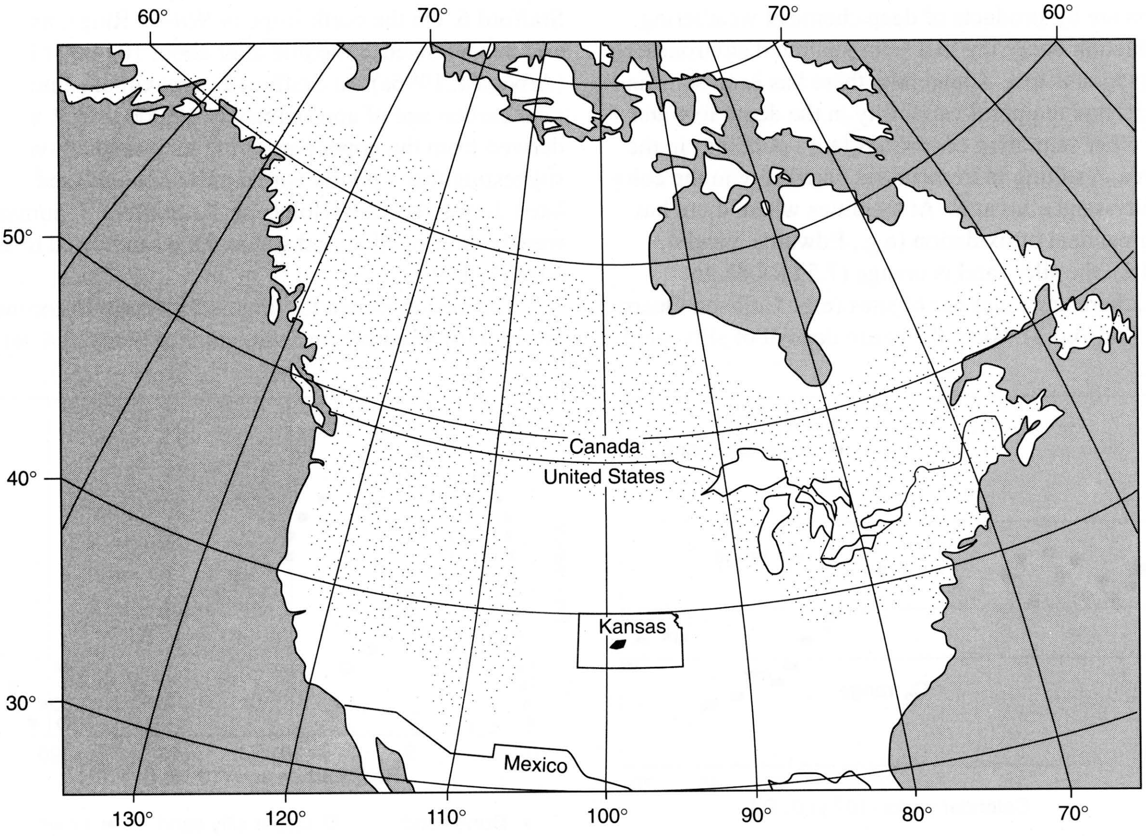
[3] Reported in Arbogast (1996b).

[4] Reported in Johnson (1991).

throughout the study area, it appears that the facies included main channel, secondary channel, and backswamp. Radiocarbon ages from trench 6 (figs. 51–53) at Wilson Ridge indicate the presence of at least one playa in the region during the late Wisconsinan. Although it is impossible to determine with the data now available, perhaps the study area was covered by a series of interconnected lakes with a very low gradient. Such a scenario is logical and may account for the near ubiquity of the silty sand, while allowing for the tremendous spatial and random diversity in the deposit.

The discrepancy between Fredlund's (1995) conclusions and the results of this study are interesting and, according to Arbogast and Johnson (1998), may shed further light on depositional processes associated with the silty sand on the Great Bend Sand Prairie. Given that Fredlund's (1995) data were derived from only one core, they may be biased. If, however, a major Woodfordian unconformity does exist at Cheyenne Bottoms, Arbogast and Johnson (1998) argued that it should also exist on the Great Bend Sand Prairie because of their proximity (<20 km; 12.4 mi) and topographic similarity (i.e., poorly drained lowlands). The primary difference between the two regions is drainage area. Cheyenne Bottoms is located within a structurally controlled, closed basin fed by Blood and Deception Creeks (Latta, 1950; Bayne, 1977). The Great Bend Sand Prairie, in contrast, is part of the much larger Arkansas River basin, which originates in the Rocky Mountains. At times during the Woodfordian, the Arkansas River was probably a meltwater stream (Schumm and Brackenridge, 1987), with widely variable discharge as mountain glaciers fluctuated. Johnson and Dort (1988) described paleomeanders on Pleistocene terraces in western Kansas that suggest that at other times the Arkansas was a much narrower, deeper stream, one that carried fine-textured sediment at much higher discharge. Given that the gradient of the Arkansas declines sharply in the vicinity of the Great Bend Sand Prairie (Fent, 1950), Arbogast and Johnson (1998) argued that the Arkansas might have regularly flooded the study area during the late Wisconsinan, which may have contributed to deposition of the silty sand in a marshy environment.

Although the silty-sand alluvium is likely a product of regional processes, evidence from Edwards 2, Stafford 1, Stafford 10, and Wilson Ridge indicates that local deposition of eolian sediments occurred. At Edwards 2, Stafford 1, and Stafford 10, radiocarbon ages of Woodfordian age were derived from the upper part of silty sand (fig. 62), suggesting that overlying eolian sands may be late Wisconsinan deposits. At Wilson Ridge, which is situated on the southern margin of a playa, two buried soils were recognized that date to the Woodfordian (fig. 51). Eolian deposition apparently began approximately 20,000 yr B.P., a period of lunette formation recognized by Reeves (1965)



▰ Great Bend Sand Prairie ⬚ Distribution of living *Discus cronkhitei*

FIGURE 60—Present distribution of *Discus cronkhitei* in North America (modified from Leonard, 1952).

in the southern High Plains, when northwesterly winds deflated sediment from the nearby lake bed. Sedimentation continued until around the glacial maximum when a well-developed soil formed in the dune. Eolian sedimentation was subsequently renewed until approximately 12,000 yr B.P., when another soil formed (Arbogast, 1996a), one generally equivalent in age to the Brady soil (Schultz and Stout, 1948; Frye and Leonard, 1951; Caspall, 1970; Johnson and May, 1992).

All of the described Woodfordian deposits on the Great Bend Sand Prairie have been tremendously altered by post-depositional processes (Arbogast and Johnson, 1998). Specifically, soil formation in the silty-sand alluvium has been very strong, potentially obliterating any diagnostic sedimentary structures. In each instance where the silty-sand alluvium was described, buried soils consisted of stacked Btb horizons that spanned the entire thickness of the exposed section. Well-defined clay films, traceable in sola over several meters, were commonly observed in exposures and thin sections (e.g., figs. 15A,B), indicating that soil formation was intense and of long duration. Slickensides were also noted on occasion (e.g., Cullison Quarry), suggesting that sola in the silty-sand alluvium have expanded and contracted over time due to wetting and drying.

Late Wisconsinan deposits have also been altered by oxidation and reduction, probably in conjunction with soil formation. According to Birkeland (1984), oxidation and reduction are by-products of deep chemical weathering, with reduction occurring in a semi-saturated environment where oxygen is low. Apparently, there has been substantial spatial and temporal variability in the depth to water table or other saturating conditions (e.g., ponding) in the study area, resulting in tremendous variability in the color of the silty-sand alluvium. At localities where there has been a great deal of oxidation (e.g., Edwards 3 and 4, Stafford 2), the silty sand is orange (7.5YR4/6). In contrast, there are a number of sites (e.g., Cullison Quarry, Reno 4, Stafford 3) where the entire deposit of silty-sand alluvium is heavily gleyed and mottled, suggesting consistently high water tables for a long period of time. In another comparison, the degree of oxidation and reduction varied sharply with depth at Edwards 1 and 2, implying that the degree of saturation differed over time. At two sites (Belpre Trench, Phillips Trench), there seems to have been very little chemical weathering, because late Pleistocene deposits range in color from brown (10YR3/2; moist) to black (10YR2/1; moist).

Early to Middle Holocene (ca. 10,000–4,000 yr B.P.)

In contrast to the evidence of a mesic landscape in the late Wisconsinan, the data suggests that early and middle Holocene environments became increasingly arid and unstable (Arbogast and Johnson, 1998). The primary evidence for greater aridity is termination of silty-sand deposition, a widespread and probable fluvial facies in the late Wisconsinan, and extinction of landsnails. Secondarily, $\delta^{13}C$ values suggest the growing dominance of C_4 grasses (fig. 61).

As the climate became more xeric during the early Holocene, eolian processes, rather than fluvial, became the most important means of sediment transport on uplands of the Great Bend Sand Prairie. Direct evidence for sediment transport by wind during the early and middle Holocene was recognized at Wilson Ridge, Stafford 2, Stafford 4, and Stafford 6. On the north slope of Wilson Ridge, an eolian unit accumulated sometime after about 10,000 yr B.P. (Arbogast, 1996a). At Stafford 2, a maximum-limiting radiocarbon age of approximately 10,000 yr B.P. was derived from the upper part of the silty-sand alluvium, suggesting that a poorly sorted unit of eolian sand accumulated during the early Holocene. At Stafford 3, humates in a weakly developed buried soil in dune sand dated to around 6,000 yr B.P.

The best evidence for large-scale, early Holocene transport of wind-blown sediment is at Stafford 4, where

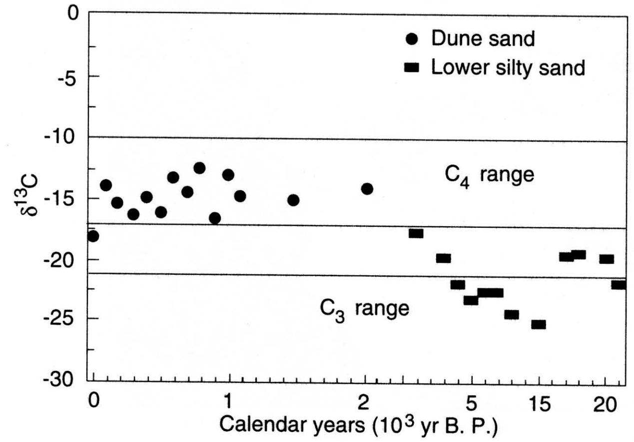


FIGURE 61—Distribution of $\delta^{13}C$ values and radiocarbon age from the lower silty sand and dune sand on the Great Bend Sand Prairie (modified from Arbogast and Johnson, 1998).

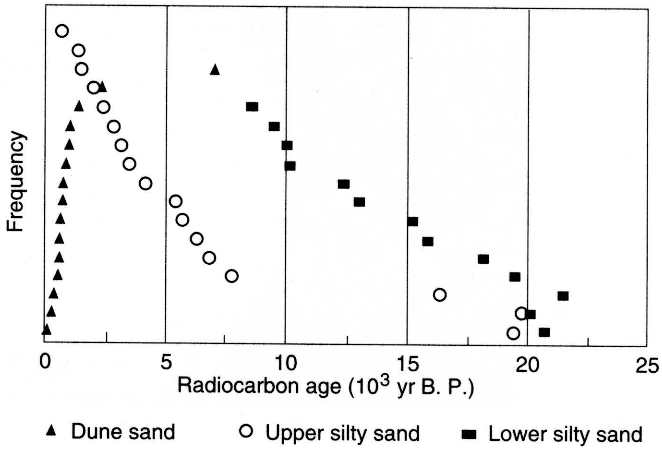


FIGURE 62—Radiocarbon ages and stratigraphic position on the Great Bend Sand Prairie (modified from Arbogast and Johnson, 1998).

grayish, unweathered loess overlies a buried soil dated to about 12,000 yr B.P. (Arbogast and Johnson, 1998). In fact, loess was recognized in reconnaissance over a relatively large (ca. 97 km^2; 60.2 mi^2) area north of Stafford. Given the underlying radiocarbon age at Stafford, the loess could be Pleistocene loess or Bignell loess (Frye et al., 1968), an early Holocene loess which has been recognized to the north and south of the study area at the Barton and Pratt County landfills, respectively (Feng, 1991). In general, Bignell loess is thought to be reworked sediment from the Brady soil, which accounts for its grayish color (Frye et al., 1968; Johnson and May, 1992; Johnson, 1993). Bignell loess is typically thin and discontinuous, filling slight depressions on upland topography (Frye et al., 1968). If the loess north of Stafford is Bignell loess, it represents one of the largest uninterrupted expanses of the unit in the central Great Plains (Arbogast and Johnson, 1998).

In addition to the direct evidence for increasingly arid conditions during the early and middle Holocene, the range of radiocarbon ages from the upper part of silty-sand alluvium suggests eolian processes were periodically mobilizing dunes over the deposit during the middle Holocene. Of the 17 ages derived from upper silty sand, 14 date to the Holocene, with five ranging between about 7,000 yr B.P. and 4,500 yr B.P. (fig. 62). Although the middle Holocene ages are probably not reliable indicators of when exposure of silty-sand alluvium last occurred at a given site, the older ages certainly suggest that the deposit has been buried for a longer period of time than those sites that have recently been deflated.

A better indicator for early and middle Holocene exposure of the silty-sand alluvium is the truncation of the upper part of the silty sand, which suggests that extensive eolian erosion occurred sometime after soil formation. The silty sand, of demonstrated late Wisconsinan age, contains an extremely well developed buried soil at every site described in this study. Although Bt horizons are well preserved in the unit, overlying A horizons are rarely observed, suggesting that they were stripped during erosional episodes in the Altithermal. Regionally, widespread eolian sand mobilization has been recognized during the middle Holocene in the southern High Plains (Holliday, 1989, 1995). Given the demonstrably drier climate of the early and middle Holocene, Arbogast and Johnson (1998) argued that the surface of the silty sand was probably exposed to deflating winds.

Late Holocene (ca. 4,000 yr B.P.–present)

Although the evidence for early and middle Holocene environmental change is largely implied and generalized, a more detailed reconstruction for late Holocene landscape evolution is possible. From a climatic perspective, values of $\delta^{13}C$ (derived from buried soils in dune sand) suggest that C_4 grasses have inhabited the area throughout the late Holocene (fig. 61). According to Arbogast (1996b), this indicates that a warm and semi-arid environment has persisted throughout the past few thousand years.

Within the semi-arid environment, eolian processes have dominated and sand dunes have periodically mobilized. Contemporary dunes are generally late Holocene deposits and are found in several forms according to their degree of development: high-relief sand sheet, compound sub-parabolic, compound parabolic, and parabolic. Potentially, three source areas exist for late Holocene dunes: (1) Holocene floodplains, (2) exposures of silty sand, and (3) older dune sand. Although each has probably contributed to some extent, the lack of Altithermal dune sand implied to Arbogast (1996b) that late Holocene dunes are reworked middle Holocene eolian sand deposits.

According to Arbogast and Johnson (1998), sedimentological data verify that dune sand resulted from different depositional processes (reflecting the increased aridity of the late Holocene) from those that caused the accumulation of silty-sand alluvium. Specifically, the sedimentological variables that best distinguish silty-sand alluvium from dune sand are mean particle size and sorting; when these variables are plotted with regard to one another, distinguishable populations result. Mean particle size in silty-sand alluvium is coarse to fine silt, indicating the high percentages of silt in the deposit. In contrast, average texture in dunes ranges from coarse silt to very fine sand. Dune sand is moderately to poorly sorted, with a near linear relationship to mean texture. Conversely, silty-sand alluvium is poorly to very poorly sorted with a much poorer relationship to texture than dune sand. According to Friedman (1967), eolian sands are better sorted because they are transported largely by saltation. Given the clear difference between the two deposits (fig. 63), Arbogast and Johnson (1998) concluded that each deposit resulted from diachronous processes, with silty sand resulting largely from alluviation and dune sand accumulating in an eolian environment.

The orientation of dune limbs is a good indicator of prevailing paleowinds. On the Great Bend Sand Prairie, most parabolic dune fields have limbs that are largely oriented southwesterly (fig. 64), though some dunes are oriented

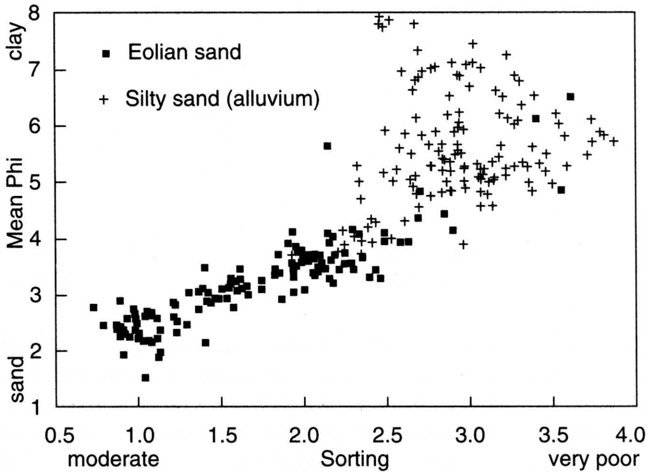


FIGURE 63—Scatterplot of mean phi and sorting in the silty sand and dune sand (modified from Arbogast and Johnson, 1998).

northwesterly. Wind data from Hutchinson, Kansas, immediately to the east of the Great Bend Sand Prairie, illustrate the probable cause for dune orientation. During the winter months prevailing winds are northwesterly (fig. 65A), whereas in the summer they are southerly to southwesterly (fig. 65B). When annual data are summarized, the region can be characterized as a high-energy environment with multidirectional winds; drift potential (DP), the relative sand-moving capability of wind, is 1,090. Because summer winds are strongest, the resultant drift direction (RDD) is northeasterly overall (fig. 65C; D. R. Muhs, personal communication, 1995; Arbogast, 1996b). Given the orientation of most parabolic dunes, summer winds are considered to be largely responsible for dune migration. Northwesterly oriented dunes indicate that mobilization of sand has also occurred during the winter.

Evidence for regional landscape instability about 1,000 yr B.P., potentially resulting from increased dryness (Hall, 1982), is well documented throughout the central Great Plains. For example, following a period of stability at approximately 1,200 yr B.P., streams throughout the region entrenched dramatically (e.g., Johnson and Martin, 1987; Johnson and Logan, 1990; May, 1992; Arbogast and Johnson, 1994). Significant mobilization and deposition of

FIGURE 64—Aerial photograph of a parabolic dune field in secs. 8, 9, T. 23 S., R. 17 W. The Arkansas River is in the northwest corner. Note the well defined, crescentic-shaped dunes in the left portion of the photograph. Bright spots are blowouts where sand is locally active. Distance across the photograph is approximately 3.2 km (2 mi).

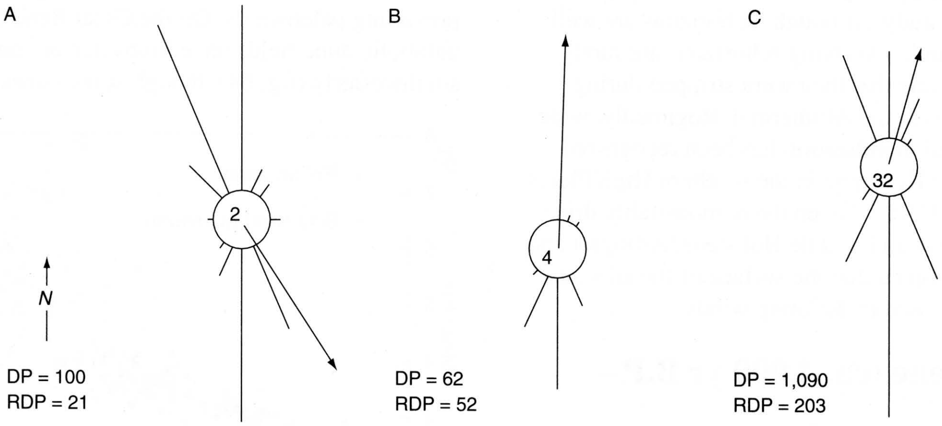


FIGURE 65—(A) February sand-rose diagram (complex) from Hutchinson, Kansas. Winds are multidirectional, with the prevailing wind from the northwest and a southeasterly resultant drift direction (RDD; from D. R. Muhs, personal communication, 1995; Arbogast, 1996a). Drift potential (DP) measures (in vector units) the relative sand moving capability of wind, whereas the RDP quantifies the resultant drift potential. The reduction factor is the number by which the vector-unit total of each sand-rose arm was divided so the longest arm would plot at <50 mm. Here, it is 2 (Fryberger and Dean, 1979). (B) July sand-rose diagram (wide unimodal) for Hutchinson. Winds are dominantly southerly with a north to northeasterly RDD. Here, the reduction factor is 4. (C) Annual summary of sand-rose diagrams (complex for Hutchinson). In this diagram, the reduction factor is 32.

eolian sand since 1,000 yr B.P. is documented at five widely scattered sites on the Great Bend Sand Prairie. At the Crocket Cutbank, a maximum-limiting age of around 900 yr B.P. was derived from a buried soil underlying approximately 6 m (20 ft) of dune sand. Similarly, buried soils in dune sand at Reno 3 and Reno 4 dated to about 700 yr B.P. At the Rice Roadcut, ages of around 500 yr B.P. and 400 yr B.P. were obtained from the bottom and top, respectively, of a buried soil in dune sand. At Stafford 6, two periods of soil formation apparently occurred in the past 600 years. Two ages were obtained from the lowermost soil buried at the site: approximately 600 yr B.P. and 500 yr B.P. from the bottom and top, respectively. Following deposition of about 60 cm (24 in) of eolian sand, a brief period of soil formation occurred around 300 yr B.P.

According to Arbogast (1996b), radiocarbon dating of buried soils in late Holocene dune sand conservatively suggests that four distinct periods of soil formation are preserved: around 2,300, 1,400, 1,000, and from 700 to 200 yr B.P. When the two buried soils from Stafford 6 are included, the record is even better resolved because they provided distinct ages of about 540 and 300 yr B.P,. As a result, Arbogast (1996b) concluded that at least five periods of late-Holocene soil formation, generally bracketing intervals of eolian sand mobilization, are preserved in dunes of the Great Bend Sand Prairie at approximately 2,300, 1,400, 1,100 to 900, 700 to 500, and 300 calibrated yr B.P.

Results from this study compare favorably with other findings derived from the central Great Plains in the past ten years. Late Holocene dune formation has been reported throughout the region (e.g., Ahlbrandt et al., 1983; Forman and Maat, 1990; Forman et al., 1995; Madole, 1986, 1994, 1995; Muhs, 1985; Ponte et al., 1994; Muhs et al., 1996), with significant reactivation of dunes in both northeastern Colorado and the Great Bend Sand Prairie in past 1,500 years. Mobilization of eolian sand in the Colorado piedmont and Kansas was apparently episodic. According to Arbogast (1996b), periods of stability and soil formation may have occurred regionally in dunes at approximately 1,400 yr B.P. and 900 yr B.P. Arbogast (1996b) demonstrated, however, that stability has transpired more frequently in the past few hundred years in south-central Kansas because soils dated at about 700–500 yr B.P. and 300 yr B.P. on the Great Bend Sand Prairie have not yet been recognized in northeastern Colorado. This is consistent with the current, more arid conditions in northeastern Colorado, which receives less than 40 cm (16 in) of precipitation annually (NOAA, 1982), as opposed to the Great Bend Sand Prairie, which gets about 75 cm (30 in) of precipitation a year.

Although dunes have mobilized in the late Holocene on the Great Bend Sand Prairie, instability has differed spatially. Surface soil development varies in many dune fields, including sola with A/Bt/C or A/AC/C horizonation, providing an indirect measure of the spatial deviation in dune stability. Calibrated radiocarbon ages from the top and bottom of buried soils at the Rice Roadcut and Stafford 6 suggest that soils with A/AC/C profiles formed within a 300-year period. As a result, many dunes (e.g., Crocket Cutbank, Cullison Quarry) have been stable for a relatively brief period of time. In contrast, dunes with A/Bt/C horizonation in their surface soils must have been stable for a longer period of time (e.g., Birkeland, 1984). These findings are consistent with accounts of early 19th-century explorers, reported by Muhs and Holliday (1995), who noted both active and inactive dunes along the Arkansas River valley.

There is also direct evidence that instability in dunes varies spatially in the study area. At the Buster Quarry, for example, a "modern" age was derived from a soil buried by 80 cm (31 in) of stabilized dune sand. Barbed wire was found overlying a buried soil in a dune at the Cullison Quarry. Blowouts are commonly occurring features in many dune fields, contributing to the "chaotic" appearance defined by Smith (1940) and indicating that many dune fields are presently near the threshold for instability (e.g., Muhs and Maat, 1993).

Contributions of This Study

This study has added to the chronology of late Quaternary landform evolution and climate change in the central Great Plains in several respects. Although preliminary research (Johnson, 1991; Feng, 1991) characterized the silty sand as Peoria loess, this study illustrates that the unit is dominantly fluvial in nature, with several facies that may have accumulated in a marshy environment of interconnected lakes. Another result of this study is the recognition that the Brady soil, previously identified only in loess and alluvium, also occurs in dunes (e.g., Wilson Ridge). Prior to this study, lunettes (e.g., Wilson Ridge) were not recognized north of the southern High Plains. In addition, the orientation of Wilson Ridge indicates that northwesterly winds prevailed as far south as Kansas during the late Wisconsinan. This study also demonstrates that the Bignell loess, previously thought to exist only in small, discontinuous areas, may be present in a relatively broad expanse in the loess plain north of Stafford on the Great Bend Sand Prairie. Another result of the study is the recognition that most dunes in the Great Bend region are younger than was previously thought. Early researchers (e.g., Smith, 1940; Simonett, 1960) of eolian sand mobilization along the Arkansas River assumed, without absolute age control, that dunes were late Wisconsinan landforms because they were on the southern side of the river (i.e., northwesterly winds produced the dunes). Although late Wisconsinan eolian sand deposits do occur (e.g., Wilson Ridge), results from this study demonstrate that most dunes in the Great Bend region are very young landforms with significant activation in the last 1,000 years due to prevailing southwesterly winds. Finally, the study demonstrates that surface soils in dunes, consisting of A/AC/C profiles, form in 200 to 300 years.

The record of late Holocene eolian sand stability and mobilization on the Great Bend Sand Prairie is especially significant in the context of greenhouse warming scenarios. Intuitively, it would appear that a dramatic shift toward a more arid climate must occur for dunes in the region to become mobilized in a widespread fashion. Although droughts have occurred periodically (e.g., 1890's, 1930's, 1950's) since climate data have been compiled for the area, they have not been severe enough to promote extensive eolian sand mobilization. On the contrary, these and other results (e.g., Madole, 1994) recently derived from the central Great Plains document intensive dune formation in the past 1,000 years, even though atmospheric circulation models indicate little variation in average surface temperature and annual precipitation during the last millenium (Kutzbach, 1987). Muhs and Maat (1993) predicted that dunes in the Great Plains could increase one activity class (e.g., inactive to active crests), using Lancaster's (1988) dune mobility index, assuming a 4°C increase in temperature as per many GCM's (Hansen et al., 1988; Wetherald and Manabe, 1988; Wendland, 1993). This study verifies the climatic sensitivity of dunes and sand sheets of the central Great Plains, indicating that the threshold for soil stability can be crossed much more easily than once thought (e.g., Smith, 1940; Simonett, 1960). These results are especially significant for land-use planning in the region if the levels of projected greenhouse warming and drying actually occur.

Future Research

Much remains to be done in the Great Bend Sand Prairie before a comprehensive model of late Quaternary landscape evolution can emerge. Specifically, future study in the region should focus on the origin and post-depositional alteration of the silty sand, and the construction of a detailed chronology of late Holocene eolian sand mobilization.

Although intensive sampling and textural analyses of the silty sand were conducted in this study, the precise mode of sedimentation remains unclear. Sedimentological evidence (e.g., sorting, fining-upward sequences, spatial variability in texture) suggests deposition in a very low energy fluvial environment of some kind during the late Wisconsinan. Paleomeanders along the Arkansas River west of Great Bend indicate that discharge was much higher during the late Wisconsinan (Johnson, 1988) and may have maintained a network of playas (e.g., Wilson Ridge) or shallow lakes on the Great Bend Sand Prairie that were interconnected by distributary streams. Given widespread deposition of Peoria loess in the area surrounding Great Bend at the same time, however, eolian silt sedimentation must have occurred in the silty sand as well. Future research should attempt to quantify the relative input of fluvial and eolian processes in this widespread, late Wisconsinan unit. In addition, an attempt should be made to clarify the history of stability, soil formation, and weathering in the unit. In particular, it would be useful to determine whether the well-developed soils in the silty sand formed in one extended event or through the episodic burial and exhumation that has apparently occurred because of surficial, mobilizing sand.

Future research should also be conducted to refine the chronology of late Holocene eolian sand mobilization and stability of dunes in the study area. In this study, all radiocarbon ages on buried soils generally estimate when periods of stability occurred, as opposed to dating the interval when deposition of dune sand transpired. In the future, optical simulated luminescence (OSL) dating should be used to establish when eolian sand mobilized. In Wyoming, Stokes and Gaylord (1993) successfully used OSL to refine the chronology of dune instability in the Ferris dune field. At Great Bend, there is an excellent opportunity to enhance the history of eolian sand mobilization through OSL dating at sites such as the Crocket Cutbank, Rice Roadcut, and Stafford 6, where good radiocarbon control on buried solums has been established in this study.

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References

Ahlbrandt, T. S., and Fryberger, S. G., 1980, Geologic and paleoecologic studies of the Nebraska Sand Hills: U.S. Geological Survey Paper 1120-A, 28 p.

Ahlbrandt, T. S., Swinehart, J. B., and Maroney, D. G., 1983, The dynamic Holocene dune fields of the Great Plains and Rocky Mountain basins, USA; *in,* Eolian Sediments and Processes, M. E. Brookfield and T. S. Ahlbrandt, eds.: New York, Elsevier, p. 379–406.

Allison, E., 1965, Organic carbon; *in,* Methods of Soil Analysis (part 2), C. A. Black, ed.: Madison, Wisconsin, American Association of Agronomy, 1,572 p.

Andreas, A. T., 1893, History of the State of Kansas—Containing the full account of its growth from an uninhabited territory to a wealthy, important State: Atchison County Historical Society, 1,616 p.

Andrews, J. T., 1987, The Late Wisconsin glacial and deglaciation of the Laurentide Ice Sheet; *in,* North America and Adjacent Oceans during the Last Glaciation, W. F. Wright and H. E. Wright, Jr., The Geology of North America, v. K-3: Boulder, Colorado, Geological Society of America, p. 13–37.

Antevs, E., 1955, Geologic-climatic dating in the West: American Antiquity, v. 20, p. 317–335.

Arbogast, A. F., 1995, Paleoenvironments and desertification on the Great Bend Sand Prairie in Kansas: Ph.D. dissertation, University of Kansas, Lawrence, 385 p.

_____, 1996a, Late Quaternary evolution of a lunette in the central Great Plains—Wilson Ridge, Kansas: Physical Geography, v. 17, no. 4, p. 354–370.

_____, 1996b, Stratigraphic evidence for late-Holocene aeolian sand mobilization and soil formation in south-central Kansas, USA: Journal of Arid Environments, v. 34, p. 403–414.

Arbogast, A. F., and Johnson, W. C., 1994, Climatic implications of the late Quaternary alluvial record of a small drainage basin in the central Great Plains: Quaternary Research, v. 41, p. 298–305.

_____, 1998, Late-Quaternary landscape response to environmental change in south-central Kansas: Association of American Geographers Annals, v. 88, p. 125–146.

Barry, R. G., 1983, Late-Pleistocene climatology; *in,* Late-Quaternary Environments of the United States, the Late Pleistocene, v. 1, H. E. Wright, and S. C. Porter, eds.: Minneapolis, University of Minnesota Press, p. 390–407.

Bartlein, P. J., Webb III, T., and Fleri, E., 1984, Holocene climatic change in the northern Midwest—Pollen-derived estimates: Quaternary Research, v. 22, p. 361–374.

Bayne, C. K., 1977, Geology and structure of Cheyenne Bottoms, Barton County, Kansas: Kansas Geological Survey, Bulletin 211, pt. 2, 12 p.

Benedict, J. B., 1973, Chronology of cirque glaciation, Colorado Front Range: Quaternary Research, v. 3, p. 584–599.

Benedict, J. B., and Olson, B. L., 1978, The Mount Albion complex—A study of prehistoric man and the Altithermal: Research Report of the Center for Mountain Archaeology, v. 1, 213 p.

Bequaert J. C., and Miller, W. B., 1973, The mollusks of the arid southwest: Tucson, University of Arizon Press, 271 p.

Birkeland, P. W., 1984. Soils and Geomorphology: New York, Oxford University Press, 372 p.

Bradbury, J. P., 1980, Late Quaternary vegetation history of the central Great Plains and its relationship to eolian processes in the Nebraska Sand Hills: U.S. Geological Survey, Professional Paper 1120C, p. 25–36.

Bradley, R. S., 1985, Quaternary Paleoclimatology: Boston, Allen and Unwin, 472 p.

Brady, R. G., 1989, Geology of the Quaternary dune sands in eastern Major and southern Alfalfa Counties, Oklahoma: Ph.D. dissertation, Oklahoma State University, 166 p.

Caspall, F. C., 1970, The spatial and temporal variations of loess deposition in northeastern Kansas: Ph.D. dissertation, University of Kansas, Lawrence, 294 p.

_____, 1972, A note on the origin of the Brady paleosol in northeastern Kansas: Proceedings of the Association of American Geographers, v. 4, p. 19–24.

Cerling, T. E., and Quade, J., 1993, Stable carbon and oxygen isotopes in soil carbonates; *in,* Climate Change in Continental Isotopic Records, P. K. Swart, K. C. Lohmann, J. McKenzie, and S. Savin, eds.: Geophysical Monograph 78, 278 p.

CLIMAP Members, 1981, Seasonal reconstruction of the earth's surface at the last glacial maximum: Geological Society of America, Map Chart MC-36.

COHMAP Members, 1988, Climatic changes of the last 18,000 years—Observations and model simulations: Science, v. 24, p. 1,043–1,052.

Courtier, W. H., 1934, Physiography and geology of south-central Kansas: Ph.D. dissertation, University of Kansas, Lawrence, 94 p.

Crowley, T. J., and North, G. T., 1991, Paleoclimatology: New York, Oxford University Press, 339 p.

Day, P. R., 1965, Particle fractionation and particle-size analyses; *in,* Methods of Soil Analysis, Part I—Physical and Mineralogical Properties, Including Statistics of Measurement and Sampling, C. A. Black, ed.: American Society of Agronomy, p. 545–567.

Dean, W. E., Jr., 1974, Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition—Comparison with other methods: Journal of Sedimentary Petrology, v. 44, p. 242–248.

Deines, P., 1980, The isotopic composition of reduced organic carbon; *in,* Handbook of Environmental Isotope Geochemistry—The Terrestrial Environment, P. Fritz and J. Fontes, eds.: New York, Elsevier, 135 p.

Delaune, R. D., 1986, The use of $\delta^{13}C$ signature of C_3 and C_4 plants in determining past depositional environments in rapidly accreting marshes of the Mississippi River deltaic plain, Louisiana, USA: Chemical Geology (Isotope Geoscience Section), v. 59, 315–320.

Delcourt, H. R., 1979, Late-Quaternary vegetation history of the eastern Highland Rim and adjacent Cumberland Plateau of Tennessee: Ecological Monograph, v. 49, p. 255–280.

Delcourt, P. A., and Delcourt, H. R., 1983, Late-Quaternary vegetational dynamics and community stability reconsidered: Quaternary Research, v. 19, p. 265–271.

Dodge, D. A., and Roth, W. E., 1978, Soil Survey of Pawnee County, Kansas: U.S. Department of Agriculture, Soil Conservation Service, 99 p.

Dodge, D. A., Hoffman, B. R., and Horshch, M. L., 1978, Soil Survey of Stafford County, Kansas: U.S. Department of Agriculture, Soil Conservation Service, 59 p.

Dreimanis, A., 1962, Quantitative gasometric determination of calcite and dolomite by using Chittick apparatus: Journal of Sedimentary Petrology, v. 32, no. 3, p. 520–529.

Fader, S. W. and Stullken, L. E., 1978, Geohyrdrology of the Great Bend Prairie, south-central Kansas: Kansas Geological Survey, Irrigation Series 4, 19 p.

Feng, Zhao-Dong, 1991, Temporal spatial variations in the loess depositional environment of central Kansas during the past 400,000 years: Ph.D. dissertation, University of Kansas, Lawrence.

Feng, Zhao-dong, Johnson, W. C., Sprowl, D. R., and Lu, Yan-chou, 1994, Loess accumulation and soil formation in central Kansas, United States, during the past 400,000 years: Earth Surface Processes and Landforms, v. 19, p. 55–67.

Fent, O. S., 1950, Pleistocene drainage history of central Kansas: Kansas Academy of Science, v. 53, no. 1, p. 81–90.

Folk, R. L., and Ward, W. C., 1957, Brazos River bar—A study in the significance of grain size parameters: Journal of Sedimentary Petrology, v. 27, p. 3–26.

Forman, S. L., and Maat, P., 1990, Stratigraphic evidence for late Quaternary dune activity near Hudson on the Piedmont of northern Colorado: Geology, v. 75, p. 745–748.

Forman, S. L., Goetz, A. F. H., and Yuhas, R. H., 1992, Large-scale stabilized dunes on the High Plains of Colorado—Understanding the landscape response to Holocene climates with the aid of images from space: Geology, v. 20, p. 145–148.

Forman, S. L., Oglesby, R., Markgraf, V., and Stafford, T., 1995, Paleoclimatic significance of Late Quaternary eolian deposition on the Piedmont and High Plains, central United States: Global and Planetary Change, v. 11, p. 35–55.

Fredlund, G. G., 1989, Paleovegetational reconstruction at the North Cove site; *in,* Archaeological Investigations at the North Cove Site, Harlan County Lake, Harlan County, Nebraska, J. J. Adair, ed.: Report submitted to the U.S. Army Corps of Engineers, Kansas City, Missouri.

_____, 1995, Late Quaternary pollen record from Cheyenne Bottoms, Kansas: Quaternary Research, v. 43, no 2. p. 67–79.

Fredlund, G. G., and Jaumann, P. J., 1987, Late Quaternary palynological and paleobotanical records from the central Great Plains; *in,* Quaternary Environments of Kansas, W. C. Johnson, ed.: Kansas Geological Survey, Guidebook Series 5, p. 167–178.

Fredlund, G. G., Johnson, W. C., and Dort, W., Jr., 1985, A preliminary analysis of opal phytoliths from the Eustis ash pit, Frontier County, Nebraska: Nebraska Academy of Sciences, Institute for Tertiary-Quaternary Studies, TER-QUA Symposium Series, v. 1, p. 147–162.

Friedman, G. M., 1967, Dynamic processes and statistical parameters compared for size frequency distribution of beach and river sands: Journal of Sedimentary Petrology, v. 37, no. 2, p. 327–354.

Frye, J. C., and Leonard, A. B., 1951, Stratigraphy of late Pleistocene loess of Kansas: Journal of Geology, v. 59, no. 4, p. 287–305.

_____, 1952. Pleistocene geology of Kansas: Kansas Geological Survey, Bulletin 109, pt. 3, p. 29–48.

Frye, J. C., Willman, H. B., and Glass, H. D., 1968, Correlation of Midwestern loesses with the glacial succession; *in,* Loess and Related Deposits of the World, C. B. Schultz and J. C. Frye, eds.: Proceedings, 7th Congress, International Association of Quaternary Research, University of Nebraska Press, Lincoln, p. 3–21.

Fryberger, S. G., and Dean, G., 1979, Dune forms and wind regime; *in,* A Study of Global Sand Seas, E. D. McKee, ed.: Chapter F, U.S. Geological Survey Professsional Paper 1052, p. 137–169.

Gile, L. H., 1979, Holocene soils in eolian sediments of Bailey County, Texas: Soil Science Society of America Journal, v. 43, p. 994–1,003.

Gruger, J., 1973, Studies on the late Quaternary vegetation history of northeastern Kansas: Geological Society of America, Bulletin 84, p. 237–250.

Hack, J. T., 1941, Dunes of the western Navajo country: Geographical Review, v. 31, p. 240–263.

Hall, S. A., 1982, Geology of Delaware Canyon; *in,* The Late Holocene Prehistory of Delaware Canyon, Oklahoma, C. R. Ferring, ed.: Contributions in Archeology 1, Institute of Applied Sciences, North Texas State University, Denton, p. 47–63.

_____, 1990, Channel trenching and climatic change in the southern Great Plains: Geology, v. 18, no. 3, p. 342–345.

Hansen, J., Fung, I., Lacis, A., Rind, D. S., Ruedy, R., and Russell, G., 1988, Global climate changes as forecast by the Goddard Institute for Space Studies three dimensional model: Journal of Geophysical Research, v. 93, p. 9,341–9,364.

Hay, R., 1893, The geology of the Great Plains: Kansas Academy of Science Transactions, v. 13, p. 3–6.

Haworth, E., 1897, Physiography of western Kansas: Kansas Geological Survey, Bulletin 2, p. 247–284.

Hoefs, J., 1980, Stable Isotope Geochemistry: New York, Springer-Verlag, 208 p.

Hoffman, B. R., Glover, R. K., and Roth, W. E., 1986, Soil survey of Kiowa County, Kansas: U.S. Department of Agriculture, Soil Conservation Service, 124 p.

Holliday, V. T., 1985, Holocene soil-geomorphological relations in a semi-arid environment; *in,* Soils and Quaternary Landscape Evolution, J. Boardman ed.: Chichester, United Kingdom, John Wiley and Sons, p. 325–357.

_____, 1987, Eolian processes and sediments of the Great Plains; *in,* Geomorphic Systems of North America, Centennial Special, v. 2, W. L. Graf, ed.: Boulder, Colorado, Geological Society of America, p. 195–202.

_____, 1989, Middle Holocene drought on the Southern High Plains: Quaternary Research, v. 31, p. 74–82.

_____, 1995, Stratigraphy and geochronology of the dune fields on the Southern High Plains: Geological Society of America Annual Meeting, North-central Section, Abstracts and Programs, p. 59.

Hopkins, D. M., 1975, Time-stratigraphic nomenclature for the Holocene Epoch: Geology, v. 3, p. 10.

Jaumann, P. J., 1991, Evidence for late Quaternary boreal environments in the Arkansas River valley, south-central Kansas—Theoretical aspects of paleoecology and climatic inferences: M.A. thesis, University of Kansas, Lawrence, 283 p.

Johnson, E. 1986, Late Pleistocene and early Holocene paleoenvironments on the Southern High Plains (USA): Géographie Physique et Quaternaire, v. 40, 249–261.

_____, 1987, Lubbock Lake: Late Quaternary Studies on the Southern High Plains, College Station, Texas A&M University Press, 179 p.

Johnson, W. C., 1988, Paleochannels of the Arkansas River, western Kansas, and hydrologic implications: Association of American Geographers, Program and Abstracts, p. 90.

_____, 1990, Age determinations on the Gilman Canyon Formation and Brady paleosol in Kansas: American Quaternary Association, Program and Abstracts, p. 21.

_____, 1991, Buried soil surfaces beneath the Great Bend Prairie of central Kansas and archeological implications: Current Research in the Pleistocene, v. 8, p. 108–110.

_____, 1993, Surficial geology and stratigraphy of Phillips County, Kansas, with emphasis on the Quaternary Period: Kansas Geological Survey, Technical Series 1, 66 p.

Johnson, W. C., and Dort, W., 1988, Paleochannels of the Arkansas River, Western Kansas, and hydrologic implications: Association of American Geographers, Program and Abstracts, p. 90.

Johnson, W. C., and Logan, B., 1990, Geoarcheology of the Kansas River basin, central Great Plains; *in,* Archeological Geology of North America, N. P. Lasca and J. Donahue, eds.: Geological Society of America, Decade of North American Geology, Centennial Special Volume 4, p. 267–299.

Johnson, W. C., and Martin, C. W., 1987, Holocene alluvial-stratigraphic studies from Kansas and adjoining states of the east-central Plains; *in,* Quaternary Environments of Kansas, Johnson, W. C., ed.: Kansas Geological Survey, Guidebook Series 5, p. 109–121.

Johnson, W. C., and May, D. W., 1992, The Brady Geosol as an indicator of the Pleistocene/Holocene boundary in the central Great Plains: American Quaternary Association, Programs and Abstracts, p. 69.

Johnson, W. C., May, D. W., and Valastro, S., 1993, A 36,000-year chrono-, bio-, and magneto-stratigraphic record from loess of south-central Nebraska: Association of American Geographers, 89th Annual Meeting, Abstracts with Programs, p. 115.

Kansas Statistical Abstracts, 1992, Institute for Public Policy and Business Research, University of Kansas, Lawrence, p. 289.

Knox, J. C., 1983, Responses of river systems to Holocene climates; *in,* Late Quaternary Environments of the United States—The Holocene, v. 2, H. E. Wright, Jr., ed.: Minneapolis, University of Minnesota Press, p. 26–41.

Koster, E. A., 1988, Ancient and modern cold-climate aeolian sand deposition—A review: Journal of Quaternary Science, v. 3, p. 69–83.

Krishnamurthy, R. V., Deniro, M. J., and Pand, R. K., 1982, Isotope evidence for Pleistocene climatic changes in Kashmir, India: Nature, v. 298, p. 640–641.

Kromer, B., Rhein, M., Bruns, M., Schock-Fischer, H., Munnich, K. O., Stuiver, M., and Becker, B., 1986, Radiocarbon calibration data for the 6th and 8th millenia B.C.: Radiocarbon, v. 28, p. 954–960.

Krumbein, W. C., 1934, Size frequency distributions of sediments: Journal of Sedimentary Petrology, v. 4, p. 65–77.

Kuchler, A. W., 1974, A new vegetation map of Kansas: Ecology, v. 55, p. 586–604.

Kutzbach, J. E., 1981, Monsoon climate of the early Holocene—Climate experiment with the Earth's orbital parameters for 9,000 years ago: Science, v. 214, p. 61.

_____, 1985, Modeling of paleoclimates: Advances in Geophysics, v. 28A, p. 159–196.

_____, 1987, Model simulations of the climatic patterns during the deglaciation of North America; *in,* North America and Adjacent Oceans During the Last Deglaciation, W. F. Ruddiman and H. E. Wright, Jr., eds., The Geology of North America, v. K-3: Boulder, Colorado, Geological Society of America, p. 425–446.

Kutzbach, J. E., and Wright, H. E., 1985, Simulation of climate of 18,000 yr B.P.—Results for the North American/North Atlantic/European sector and comparison with the geologic record: Quaternary Science Reviews, v. 4, p. 147–187.

Kutzbach, J. E., Guetter, P. J., Behling, P. J., and Selin, R., 1993, Simulated climatic changes—Results of the COHMAP climate-model experiments; *in,* Global Climates Since the Last Glacial Maximum, J. E. Kutzbach, T. Webb, III, W. F. Ruddiman, F. A. Street-Perrott, and P. J. Bartlein, eds.: Minneapolis, University of Minneapolis Press, p. 24–93.

Lancaster, N., 1988, Development of linear dunes in the southwestern Kalahari, southern Africa: Journal of Arid Environments, v. 14, p. 233–244.

Latta, B. F., 1950, Geology and groundwater resources of Barton and Stafford Counties: Kansas Geological Survey, Bulletin 88, 228 p.

Lea, P. D., and Waythomas, C. F., 1990, Late-Pleistocene Eolian Sand Sheets in Alaska: Quaternary Research, v. 34, 269–281.

Leonard, A. B., 1951, Stratigraphic zonation of the Peoria loess in Kansas: Journal of Geology, v. 59, no. 4, p. 323–332.

_____, 1952, Illinoian and Wisconsinan molluscan faunas in Kansas: University of Kansas Paleontological Contributions, Mollusca, pt. 4, 38 p.

Linick, T. W., Suess, H. E., and Becker, B., 1985, La Jolla measurements of radiocarbon in south German oak tree-ring chronologies: Radiocarbon, v. 27, p. 20–32.

Linick, T. W., Long, A., Damon, P. E., and Ferguson, C. W., 1986, High-precision radiocarbon dating of bristlecone pine from 6,554 to 5,350 B.C.: Radiocarbon, v. 28, 943–953.

Logan, B., Arbogast, A. F., and Johnson, W. C., 1993, Geoarcheology of the Kansas Sand Prairies: University of Kansas, Museum of Anthropology, Project Report Series No. 83, 95 p.

Lugn, A. L., 1935, The Pleistocene geology of Nebraska: Nebraska Geological Survey, Bulletin 10, 223 p.

Madole, R. F., 1986, Pleistocene and Holocene dune stratigraphy, Wray Dune Field, Colorado–Nebraska: American Quaternary Association Abstracts, p. 148.

_____, 1994, Stratigraphic evidence of desertification in the west-central Great Plains within the past 1,000 yr: Geology, v. 22, p. 483–486.

_____, 1995, Spatial and temporal pattterns of late Quaternary eolian deposition, eastern Colorado, USA.: Quaternary Science Reviews, v. 14, p. 155–177.

Mandel, R. D., and Olson, C. G., 1995, Soils and Quaternary landscape evolution in the Cimarron River valley, Haskell County, Kansas; *in,* Landscape Evolution of the Cimarron River Valley, Southwestern Kansas, D. Porter, ed.: Friends of the Pleistocene, South-central Cell, Guidebook, p. 24–35.

Martin, C. W., and Johnson, W. C., 1995, Variation in radiocarbon ages of soil organic matter fractions from late Quaternary buried soils: Quaternary Research, v. 43, p. 232–237.

Martin, L. D., 1984, The effect of Pleistocene and recent environments on man in North America: Current Research in the Pleistocene, v. 1, p. 73–75.

Martin, L. D., and Hoffman, R. S., 1987, Pleistocene faunal provinces and Holocene biomes of the central Great Plains; *in,* Quaternary Environments of Kansas, W. C. Johnson, ed.: Kansas Geological Survey, Guidebook Series 5, p. 159–165.

Martin, L. D., and Martin, J. B., 1987, Equability in the late Pleistocene; *in,* Quaternary Environments of Kansas, W. C. Johnson, ed.: Kansas Geological Survey, Guidebook Series 5, p. 123–127.

May, D. W., 1992, Late Holocene valley-bottom aggradation and erosion in the South Loup River valley, Nebraska: Physical Geography, v. 13, p. 115–132.

McKee, E. D., 1979, Introduction to a study of global sand seas; *in,* A Study of Global Sand Seas, Chapter A, E. D. McKee, ed.: U.S. Geological Survey Professional Paper 1052, p. 305–399.

Melton, F. A., 1940, A tentative classification of sand dunes—Its application to dune history in the Southern High Plains: Journal of Geology, v. 57, p. 13–145.

Mettler, D. E., 1955, Dune sands of the Syracuse area in Kansas: M.S. thesis, University of Kansas, Lawrence, 95 p.

Merriam, D. F., 1963, Geologic history of Kansas: Kansas Geological Survey, Bulletin 162, 317 p.

Mook, W. G., 1986, Business meeting—Recommendations/resolutions adopted by the twelfth international radiocarbon conference: Radiocarbon, v. 28, p. 799.

Moore, R. C., 1920, Geology of Kansas: Kansas Geological Survey, Bulletin 6, 98 p.

Muhs, D. R., 1985, Age and paleoclimatic significance of Holocene sand dunes in northeastern Colorado: Association of American Geographers Annals, v. 75, p. 566–582.

_____, 1991, The potential response of Great Plains eolian sands to greenhouse warming and precipitation reduction: Geological Society of America, Abstracts with Programs, v. 23, no. 5, p. A285.

Muhs, D. R., and Holliday, V. T., 1995, Active dune sand on the Great Plains in the 19th century—Evidence from accounts of early explorers: Quaternary Research, v. 43, no. 1, 198–208.

Muhs, D. R., and Maat, P. B., 1993, The potential response of eolian sands to greenhouse warming and precipitation reduction on the Great Plains of the USA.: Journal of Arid Environments, v. 25, p. 351–361.

Muhs, D. R., Stafford, T. W., Jr., Bush, C. A., Hedge, C. E., Swinehart, J. B., Cowherd, S. D., and Mahan, S. A., 1995, Geochemistry of the Nebraska Sand Hills—Evidence for mineralogical maturity in a recently active sand sea: Geological Society of America, Abstracts with Programs, v. 27, no. 3, p. 75.

Muhs, D. R., Stafford, T. W., Jr., Cowherd, S. D., Mahan, S. A., Kihl, R., Maat, P. B., Bush, C. A., and Nehring, J., 1997, Origin of late Quaternary dune fields of northeastern Colorado: Geomorphology, v. 17, p. 129–149.

NOAA, 1982, Climatography of the United States, no. 81—Monthly Normals of Temperature, Precipitation, and Heating and Cooling Degree Days 1951–1980, Colorado: Ashville, North Carolina, U.S. Government Printing Office, p. 19.

The North American Commission of Stratigraphic Nomenclature, 1983, North American Stratigraphic Code: The American Association of Petroleum Geologists, Bulletin 67, p. 841–875.

Nordt, L. C., Bouton, T. W., Hallmark, C. T., and Waters, M. R., 1994, Late Quaternary vegetation and climatic changes in central Texas based on the isotopic composition of organic carbon: Quaternary Research, v. 41, p. 109–120.

Olson, C. G., Porter, D. A., Ransom, M. D, Nettleton, W. D., 1995, Source and distribution of eolian surficial material, central and southern High Plains: Geological Society of America Annual Meeting, North-central Section, Abstracts with Programs, v. 27, no. 3, p. 78.

Pierce, H.G., 1987, The gastropods, with notes on other invertebrates; *in,* Geology and soils of Lubbock Lake, Late Quaternary Studies on the Southern High Plains, E. Johnson, ed.: College Station, Texas A&M University Press, p. 41–48.

Ponte, M. R., Loope, D. R., and Swinehart, J. B., 1994, Significance of interbedded eolian sand and peat beneath interdunes of the central Nebraska Sand Hills: Geological Society of America, Abstracts with Programs, v. 26, no 7, p. A62.

Prante, M. C., 1989, Grain size—A progam to aid pedologic particle-size analysis, University of Kansas, Department of Geography (unpublished computer program).

Reed, E. C., and Dreeszen, V. H., 1965, Revision of the classification of the Pleistocene deposits of Nebraska: Nebraska Geological Survey, Bulletin 23, p. 65.

Reeves, C. C., Jr., 1965, Chronology of west Texas pluvial lake dunes: Journal of Geology, v. 73, p. 504–508.

_____, 1966, Pluvial lake basins of west Texas: Journal of Geology, v. 74, p. 269–291.

Rosner, M. L., 1988, The stratigraphy of the Quaternary alluvium in the Great Bend Prairie: M.S. thesis, University of Kansas, Lawrence, 135 p.

Ruddiman, W. F., 1987, Synthesis—The ocean/ice sheet record; *in,* North America and Adjacent Oceans during the last Deglaciation, W. F. Ruddiman and H. E. Wright, Jr., eds., The Geology of North America, v. K-3: Boulder, Colorado, Geological Society of America, p. 463–478.

Schlesinger, M. E., 1989, Model projections of the climatic changes induced by increased atmospheric CO_2; *in,* Climate and Geo-Sciences, A. Berger, S. H. Schneider, and J. C. Duplessy, eds.: Dordrecht, Netherlands, D. Reidel, p. 375–415.

Schoewe, W. H., 1949, The geography of Kansas, part 2—Physical geography: Kansas Academy of Science Transactions, v. 52, no. 3, p. 263–329.

Schultz, C. B., and Stout, T. M., 1948, Pleistocene mammals and terraces in the Great Plains: American Journal of Science, v. 243, p. 231–244.

Schumm, S. A., and Brackenridge, G. R., 1987. River Responses; *in,* North America and Adjacent Oceans during the last Deglaciation, W. F. Ruddiman and H. E. Wright, Jr., eds., The Geology of North America, v. K-3: Boulder, Colorado, Geological Society of America, p. 463–478.

Schwan, J., 1988, The structure and genesis of Weichselian to Early Holocene aeolian sand sheets in western Europe: Sedimentary Geology, v. 55, p. 197–232.

Simonett, D. S., 1960, Development and grading of dunes in western Kansas: Association of American Geographers Annals, v. 50, p. 216–241.

Smith, H. T. U., 1938, Quaternary dune building in Kansas: Geological Society of America Proceedings, p. 115.

_____, 1939, Sand dune cycle in western Kansas: Geological Society of America, Bulletin 50, p. 1,934–1,935.

_____, 1940, Geologic studies in southwestern Kansas: Kansas Geological Survey, Bulletin 34, 212 p.

_____, 1965, Dune morphology and chronology in central and western Nebraska: Journal of Geology, v. 73, p. 557–578.

Soil Survey Staff, 1987, Soil survey laboratory methods and procedures for collecting soil samples, USDA-SCS, Soil Survey Investigation Report, No. 1.: Washington, D.C., U.S. Government Printing Office.

Stokes, S., and Gaylord, D. R., 1993, Optical dating of Holocene dune sands in the Ferris dune field, Wyoming: Quaternary Research, v. 39, p. 274–281.

Stuiver, M., 1969, Yale natural radiocarbon measurements IX: Radiocarbon, v. 11, p. 545–648.

Stuiver, M., and Polach, H. A., 1977, Reporting of ^{14}C data: Radiocarbon, v. 19, p. 355–363.

Stuiver, M., Pearson, G. W., and Braziunas, T., 1986, Radiocarbon age calibration of marine samples back to 9,000 cal. yr B.P.: Radiocarbon, v. 28, p. 980–1,021.

Stuiver, M., and Reimer, P. J., 1993, Radiocarbon Calibration Program Revision 3.0: Radiocarbon, v. 35, p. 215–230.

Swinehart, J. B. 1990, Wind-blown deposits; *in,* An Atlas of the Sandhills, A. Bleed and C. Flowerday, eds., Conservation and Survey Division, Institute of Agriculture and Natural Resources, Resource Atlas No. 5a: Lincoln, University of Nebraska, p. 43–56.

Taylor, T. L., 1987, Radiocarbon Dating—An Archeological Perspective: Orlando, Florida, Academic Press, 212 p.

Washington, W. M., and Meehl, G. A. 1984, Seasonal cycle experiment on the climatic sensitivity to doubling of CO_2 with an atmospheric general circulation model coupled to a simple mixed-layer ocean model: Journal of Geophysical Research, v. 89, p. 9,475–9,503.

Watts, W. A., and Wright, H. E., Jr. 1966, Late-Wisconsin pollen and seed analysis from the Nebraska Sand Hills: Ecology, v. 47, p. 202–210.

Webb, T., 1985, Holocene palynology and climate*; in,* Paleoclimate Analysis and Modeling, A. D. Hecht, ed.: New York, Wiley-Interscience, p. 163–196.

Wells, G. L., 1983, Late-glacial circulation over central North America revealed by aeolian features; *in,* Variations in the Global Water Budget, A. Street-Perrott, M. Beran, and R. Ratcliffe, eds.: Dorchrecht, Netherlands, D. Reidel, p. 317–330.

Wells, P. V., and Stewart, J. D., 1987, Cordilleran-boreal taiga and fauna on the central Great Plains of North America, 14,000–18,000 years ago: American Midland Naturalist, v. 118, p. 94–106.

Wendland, W. M., 1993, Kansas climate with global warming—Agricultural and other economic impacts: Transactions of the Kansas Academy of Science, v. 96, p. 161–166.

Wendorf, F., and Hester, J. J., 1975, Late Pleistocene Environments of the Southern High Plains: Ft. Burgwin Research Center Publication 9, 290 p.

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

Wetherald, R. T., and Manabe, S., 1988, Cloud feedback processes in a general circulation model: Journal of the Atmospheric Sciences, v. 45, p. 1,397–1,415.

Wright, H. E., Jr., 1970, Vegetational history of the Central Plains; *in,* Pleistocene and Recent Environments of the Central Great Plains, W. Dort, Jr. and J. K. Jones, eds.: Lawrence, Kansas, University Press of Kansas, p. 157–172.

Wright, H. E., Almendinger, J. C., and Gruger, J., 1985, Pollen diagram from the Nebraska Sandhills and the age of the dunes: Quaternary Research, v. 24, p. 115–120.

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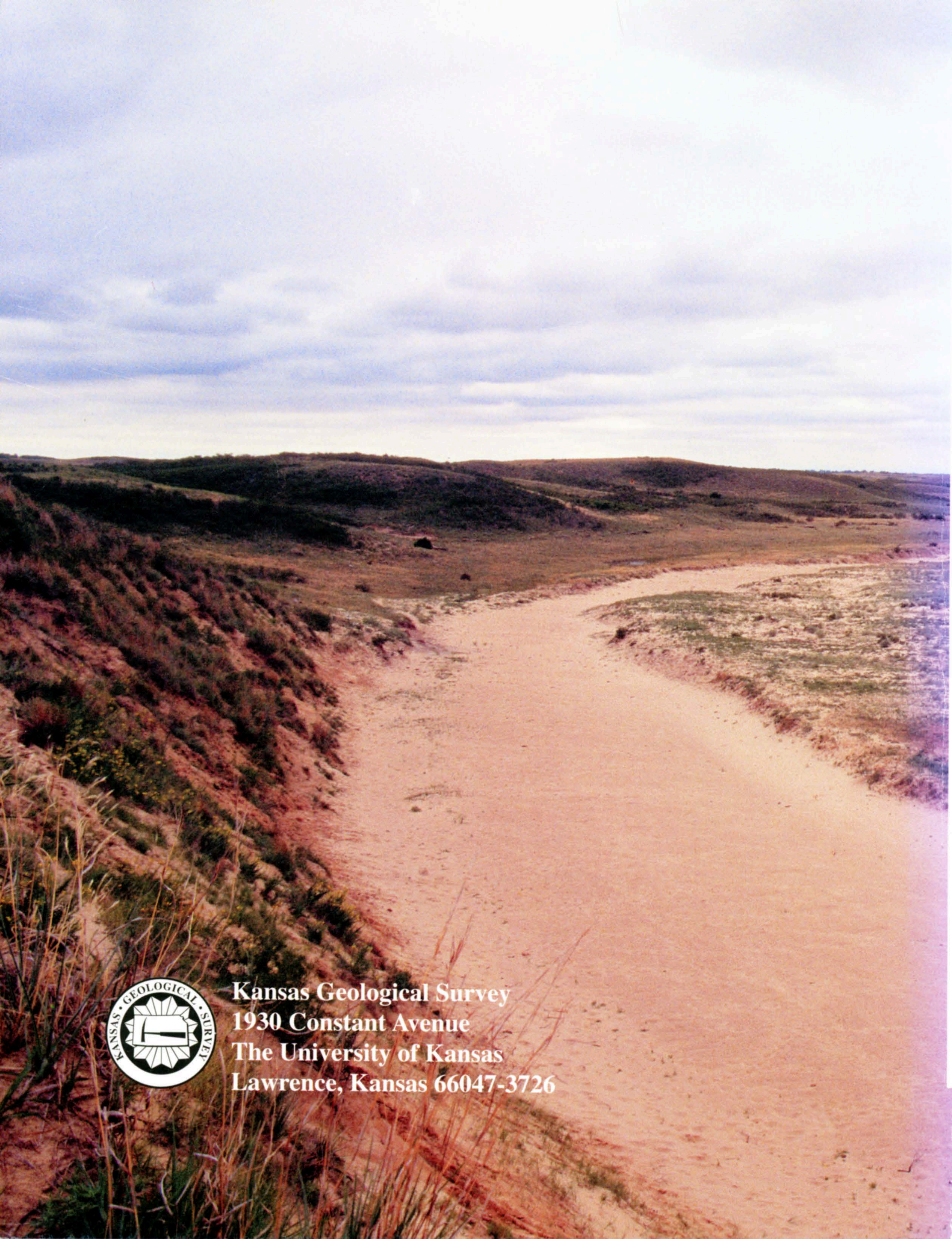

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