LATE QUATERNARY LANDFORM EVOLUTION IN THE REPUBLICAN RIVER BASIN, NEBRASKA

by

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ABSTRACT

Integration of sedimentological and radiocarbon data collected from upland, valley side, and valley bottom deposits has revealed a chronology of landform evolution during the last 30,000 years along an 80-km reach of the Republican River basin, south central Nebraska. Beginning shortly after 26,200 yr B.P., the Peoria loess, which is preserved on uplands, was deposited, burying a paleosol that appears to be the temporal equivalent of the Gilman Canyon Formation. Weakly developed paleosols within the Peoria loess indicate that deposition was episodic. Peoria loess deposition continued until just before 13,000 yr B.P., at which time an episode of entrenchment occurred. Between 13,000 and 11,400 yr B.P., the recently-cut valleys were filled with late Peoria loess and reworked Peoria loess. Floral and faunal data from that period indicate a higher water table and cooler and wetter climatic conditions than those of the present.

Surface stability and regional soil development dominated between 11,400 and 10,200 yr B.P. This soil, which is the temporal equivalent of the Brady paleosol, was buried after 10,200 yr B.P. by the Bignell loess on

uplands and by fluvially-deposited sediments in the valley. Bignell loess deposition appears to have ended around 8000 yr B.P., but alluviation in the valley continued until approximately 4500 yr B.P., probably under progressively warmer and drier climatic conditions. Soil formation occurred around 4500 yr B.P., perhaps as the climate became more mesic, and was then followed by renewed eolian depostion of reworked Peoria loess on valley sides. At some time between 4500 and 3700 yr B.P., the Republican River entrenched about 10 m.

The late Holocene record (post-4000 yr B.P.) has been characterized by stability on uplands and valley sides and alternating floodplain stability and rapid floodplain aggradation on the valley floor. Floodplain accretion took place between post-3000 and pre-2700, post-2700 and pre-2000, and post-2000 yr B.P. Floodplain stability, as denoted by paleosols, occurred around 3700 to 3000, 2700, and 2000 yr B.P. This pattern of alternating floodplain aggradation and stability is thought to have resulted from changes in flood magnitude and frequency, with stability taking place during periods of infrequent flooding or low magnitude flooding, and floodplain aggradation occurring during times of frequent flooding or high magnitude flooding. At some time after 2000 B.P., there was entrenchment by the Republican River of about 7 m, followed by gradual floodplain accretion. The synchrony between the late Holocene chronology reported here and that documented elsewhere in the central Great Plains implies a climate-control of fluvial landform change. Additionally, the late Holocene record demonstrates that river systems are highly sensitive to fluctuations in environmental conditions.

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CHAPTER 1: INTRODUCTION

This Study

As revealed in recent summaries (e.g., Johnson and Martin 1987; Osterkamp et al. 1987; Johnson and Logan, In press), the chronology of late Quaternary sedimentation and erosional activity in the central Great Plains is just beginning to emerge. With growing concern about human impact on environmental systems, including fluvial and climatic systems, studies that assess past landform changes, their causes, and the rates at which they progressed will be requisite to predicting future change (see Revelle 1982; Porter 1983; Oliver 1986; COHMAP 1988). By providing a chronology of late Quaternary (last 30,000 years) landform evolution in an area that has, to this point, received limited attention, this study adds to the small volume of research for the central Great Plains. The radiocarbon control used here also permits comparison of the resultant chronology of landform evolution to chronologies that have been constructed for areas within and peripheral to the central Great Plains. Additionally, the chronology can be compared to the forthcoming paleoenvironmental record of the region, thereby revealing the type of geomorphic activity that results from a given environmental regime.

Specifically, the objectives of this study are as follows:

1) **Distinguish among fluvial, colluvial, and eolian sediments along an 80-km reach of the Republican River in south central Nebraska, and determine the temporal and spatial relationship among them.**

2) **Reconstruct the upland and lowland chronology in the study area, and, where possible, document the rate at which change occurred.**

3) **Make revisions in the "classical" late Quaternary stratigraphy of the area, most of which was constructed prior to the advent of radiocarbon dating.**

4) **Speculate on the paleoclimatic conditions that caused landform change.**

This research focuses on Harlan Lake and an associated upstream reach of the Republican River in south-central Nebraska (Figure 1:1). Seasonal fluctuations in the level of the lake have provided exposures of upland and valley side deposits (Plate 1:1), and lateral migration of the Republican River has produced exposures of fluvial deposits in the valley (Plate 1:2). In addition, the study area is positioned midway between the Loup River system, which has been the focus of extensive work by May (1986, 1989), and the

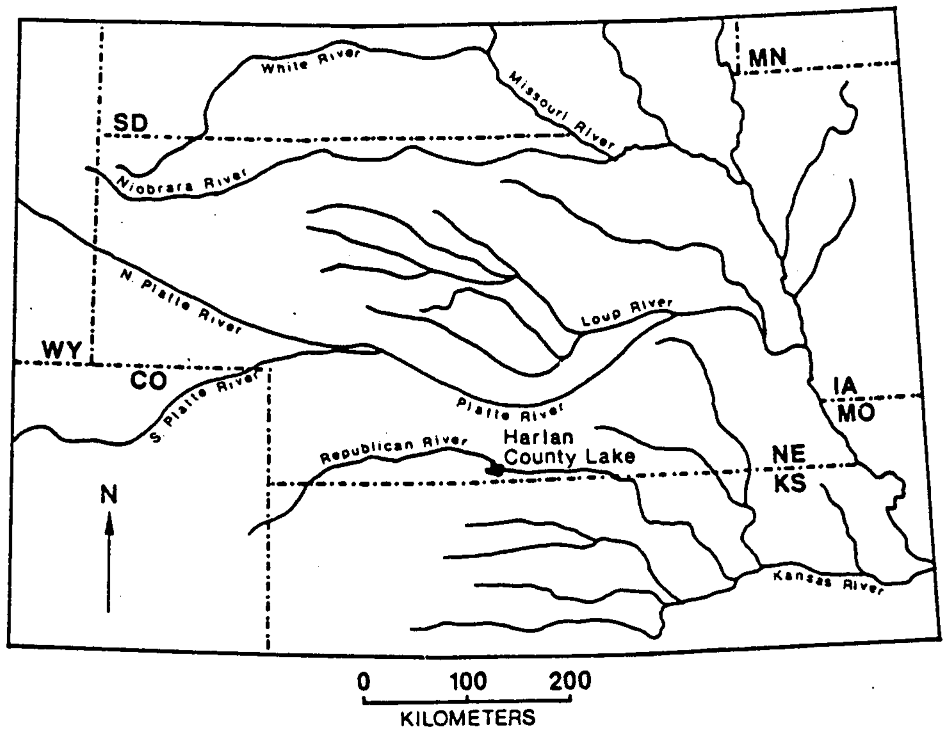

Figure 1:1 Position of Harlan Lake and Republican River in the Great Plains. Note that the Republican River is a major tributary to the Kansas River. May (1986,1989) has constructed an alluvial chronology for the Loup River system, Nebraska, and Johnson and Martin (1987) a chronology for the Kansas River system, north central and northeastern Kansas.

Plate 1:1 Harlan Lake, Nebraska. Photo shows loess bluffs (Peoria loess) on the south shore of the lake. View is looking to the west from a location just west of the Bone Cove site.

Plate 1:2 Cutbank on the Republican River, Nebraska. This is the Schoenenburg site, located about 3 km south of Orleans, Nebraska. Deposits are late Holocene in age.

Kansas River (see Figure 1:1), which has received the attention of several researchers (see Johnson and Martin 1987; Johnson and Logan, In press). That is, the results of this study provide information for a region that has not, hitherto, been studied in such a detailed fashion. Further, as Johnson and Martin (1987) noted, it is not known whether changes documented along major trunk streams in the central Great Plains occurred concurrently along major tributary streams. By reconstructing change along the Republican River, a major tributary to the Kansas River, this study should help in assessing the spatial and temporal patterns of fluvial change. Research completed in other river systems has revealed that the direction and magnitude of response in the fluvial system may vary according to position in the drainage basin (e.g., Knox et al. 1981; Van Nest and Bettis 1990).

Reconstruction of Landform Evolution

The reconstruction of landforms depends on accurate interpretation of the record of sedimentation and erosional activity. As Baker (1983,118) has noted, the fluvial record is especially critical to late Quaternary research in the central Great Plains:

> "... in many interior regions of the United States, fluvial deposits provide the only easily studied record of Quaternary paleoclimate."

Deciphering the stratigraphic record, especially the fluvial record, is often difficult, however: there are problems interpreting it, correlating it from region to region, and determining the causes of change preserved in it.

Despite considerable research, a thorough understanding of fluvial depositional environments remains elusive (Reineck and Singh 1975; Allen 1978; Collison 1986). One common approach to reconstructing change in depositional conditions is to analyze changes in grain sediment size within a deposit (e.g., Folk and Ward 1957). Unfortunately, such studies have produced a range of conclusions: sediment size in stratigraphic sequences has been linked to sediment source area (e.g., Diffendal and Corner 1983; Kochel and Baker 1988); proximity to the active channel at the time of deposition (e.g., Nanson 1980; Bown and Kraus 1987; Brakenridge 1988); and changes in flood magnitude and frequency (e.g., Starkel 1983; Knox 1985).

Although numerous studies have demonstrated that the sedimentation record contains evidence of changes in fluvial depositional and erosional activity (e.g., Schumm 1976; Baker and Penteado-Orellana 1977; Knox et al. 1981; Starkel 1983; Waters 1988), it has proven difficult to determine the cause of such changes. Much of the literature suggests that fluvial adjustment results from climate change, although there is considerable disagreement about the type of adjustment that is triggered by a given climatic regime. Indeed, some have argued that more than one type of fluvial adjustment can result from a given climatic regime (e.g., Baker and Penteado-Orellana 1977).

Early work by Bryan (1941) and Antevs (1952) in the American southwest indicated that river incision occurred during dry periods, aggradation during the transition to wet periods. According to this model, dry conditions reduce the vegetation cover, thereby resulting in larger floods having increased erosive power. Mesic conditions, on the other hand, increase the vegetation cover, thereby slowing the velocity of river flow. As a result of the decrease in flow velocity, deposition occurs. Subsequent work in the region reported similar findings (e.g., Leopold and Miller 1954; Karlstrom 1988; Waters 1988). Conversely, several studies linked mesic periods to a regime of river incision. According to this model, increased moisture fosters an expanded vegetation cover which, in turn, reduces sediment delivery to channels and prompts entrenchment. In contrast, xeric conditions cause contraction of the vegetation cover, and a concomitant increase in sediment delivery to channels; oversupplied with sediment, rivers then aggrade their beds (Huntington 1914; Love 1979; Brakenridge 1981,1984; Knox 1983).

Other studies postulated that adjustment is governed by the complex response of the drainage basin to a single event. Adjustment occurs when an internal threshold (e.g., slope, surface material, vegetation) is exceeded. Adjacent basins, depending on how close they are to exceeding a threshold, may respond differently to change in a single external variable (e.g., climate, base-level, tectonics). This hypothesis, developed to explain the lack of synchrony among erosional and depositional episodes in adjacent basins of the American southwest (Schumm 1973), is supported by findings from several studies in the western United States (e.g., Schumm and Parker 1973; Womack and Schumm 1977; Patton and Schumm 1981; Boison and Patton 1985).

Although most researchers have argued that climate is an important, if not the most important, factor in fluvial activity, some studies have suggested that conditions inherent to a particular drainage basin may be equally important in explaining activity. Trimble (1983) maintained that floodplains are fairly immobile, even during high magnitude fluvial events. He argued that vegetation armors floodplain sediments and reduces the velocity of overbank flow, thereby reducing the erosive power of overbank floods. He concluded that the erosion of deposits and the loss of sediment from storage is a slow process that is affected by factors (e.g., amount of sediment in storage, conveyance capacity of streams) in addition to climate (Trimble and Lund 1982; Trimble 1983; Trimble 1989). This model stands in contrast to that of Knox (1983), which concluded that changes in the magnitude and frequency of floods are key determinants of fluvial activity.

Recent studies of drainage basins, although producing conflicting results in several instances, generally agree that the fluvial system is inherently complex, and that adjustment within it is complicated and often non-synchronous (Butzer 1980; Bettis and Benn 1984). When regional synchrony is observed, it is generally assumed to have resulted from major change in an external variable such as climate, base-level, or tectonics (Wendland 1982; Knox 1983). In studies of the late Quaternary record of the Great Plains, tectonism can usually be eliminated as a variable (Osterkamp et al. 1987). The existence of synchrony among sedimentation records in the region is typically linked, therefore, to a climatic control (Wendland 1982; Knox 1983; May 1989).

This study will develop a model of landform evolution in an area of the central Great Plains that has not, to this point, been studied. The reach of the Republican River examined here fills a "geographical niche" between the Loup River system to the north and the Kansas River system to the south, both of which have received attention from investigators (e.g., May 1986, 1989; Johnson and Martin 1987; Johnson and Logan, In Press). Also, the Republican River falls between these two systems in terms of size (larger than the Loup River, smaller than the Kansas River), and is a major tributary of the Kansas River. The resultant radiocarbon-controlled chronology will be compared to dated sequences in the surrounding region to determine whether synchrony exists among them. Finally, the chronology will be compared to the emerging paleoenvironmental record of the region, and, where possible, the paleoclimatic conditions that appear to have triggered landform change will be discussed.

CHAPTER 2: LATE QUATERNARY LANDFORM EVOLUTION IN THE CENTRAL GREAT PLAINS

Late Pleistocene Landform Evolution

Much of the chronology of Pleistocene landform evolution for the central Great Plains, including stratigraphic relationships and implications, was compiled in the 1940s and 50s, prior to the development of radiocarbon dating (e.g., Lugn 1935; Schultz and Stout 1945; Frye and Fent 1947; Frye and Leonard 1951, 1952); subsequent revisions, as summarized in Reed and Dreeszen (1965), Bayne and O'Connor (1968), and Dreeszen (1970), afforded radiocarbon control and additional detail to the chronology. Unfortunately, as detailed in the following review, little has been done during the last twenty years to further refine the chronology. Moreover, the work completed to date varies in scope and detail. At one extreme are the single-site archaeological studies that contain little, if any, information about fills and surfaces other than those associated with excavation sites; at the other end are the basin-wide surveys that measure, describe, and correlate fills and terraces throughout a drainage basin. Additionally, erosion has removed a large part of the late Quaternary record from many drainage basins in the central Great Plains (see Knox 1983). Consequently, a comprehensive sedimentation and erosion chronology for the region is lacking (see Knox 1983; Osterkamp et al. 1987).

My review of the literature is arranged chronologically from the late Pleistocene to the late Holocene. The survey begins at 30,000 yr B.P. to ensure that it includes the Peoria loess, the major upland deposit found in the study area. The date of the Pleistocene/Holocene boundary is designated after Porter (1983), and the divisions of the Holocene are based on the tripartite climates as proposed by Antevs (1955). The temporal categories are:

Late Pleistocene	30,000 - 10,000 yr BP
Early Holocene	10,000 - 7500 yr BP
Middle Holocene	7500 - 4000 yr BP
Late Holocene	4000 yr BP - Present

Late Pleistocene (30,000 - 10,000 yr B.P.)

The most comprehensive record of late Pleistocene fluvial sedimentation comes from Nebraska, where Schultz and Stout (1945) developed, and in subsequent years revised, their T0 (floodplain) to T5 (highest terrace) sequence of terraces and fills (see Schultz and Stout 1948; Schultz and Martin 1970; Schultz and Hillerud

1978). Unfortunately, the site descriptions contained in much of their work lacked the requisite stratigraphic detail and documentation for the construction of a regional fluvial chronology.

In some of the first basin-wide, radiocarbon-controlled studies of drainage systems in the central Great Plains, Brice (1964, 1966) mapped two Pleistocene-age terraces in Nebraska, the Wellfleet terrace in the Medicine Creek valley and the Kilgore terrace in the Loup River valley. Although the ages of the Wellfleet terrace and fill were unknown, Libby (1955) reported a radiocarbon age of 10,500 yr B.P. on charcoal extracted from basal fill of the Stockville terrace, the terrace that is inset against the Wellfleet. In the Loup River valley, the entrenchment that formed the Kilgore terrace occurred prior to 10,500 yr B.P., and was followed by alluviation of the Elba terrace fill after 10,500 yr B.P. (Brice 1964). In the Middle Platte River valley, Nebraska, May (1989a) reported deposition of the Todd Valley Formation around 14,000 yr B.P., and incision to form the Todd terrace shortly thereafter; this terrace may be contemporaneous with the Wellfleet terrace of Medicine Creek. Diffendal and Corner (1983) noted three late Quaternary alluvial fills along Pumpkin Creek in western Nebraska, but, owing to a lack of datable material, did not provide a chronology of fluvial activity.

Along the Pomme de Terre River in south-central Missouri, fill beneath the highest of three terraces present in the valley was deposited, and subsequently incised, between 32,000 and 30,000 yr B.P. (Haynes 1976, 1985). Fill of the middle terrace was deposited between 26,000 and 16,500 yr B.P. Stability and soil formation occurred between 16,500 and 13,000 yr B.P., followed after 13,000 yr B.P. by the entrenchment that formed the middle terrace. Alternating aggradation and incision took place around 11,000 yr B.P., and then aggradation during the late Pleistocene and early Holocene (Haynes 1976,1985). In the same valley, Brakenridge (1981) identified a period of floodplain stability between 26,000 and 13,500 yr B.P., and an episode of aggradation between 10,500 and 8100 yr B.P.

The late Pleistocene fluvial record of Kansas is documented most completely along the Kansas River and its tributaries in the northeastern part of the state (Johnson and Logan, In press). At several locations along the Kansas River valley, Frye and Leonard

(1952,1954) noted late Pleistocene alluvium beneath the surface of a low terrace. Davis and Carlson (1952) mapped three terraces along the Kansas River, and Dufford (1958) correlated them to Pleistocene glacial advances; neither study, however, provided detailed information about fills underlying the terrace surfaces, or about the surfaces themselves. Recent radiocarbon dating of terrace fills along the Kansas River has shown that deposits assigned previously to the Pleistocene actually date to the Holocene (Holien 1982; Bowman 1985; Johnson 1985). Following a survey of radiocarbon assays from the Kansas River basin, Johnson and Martin (1987) concluded that a regionally synchronous episode of floodplain stability and strong soil formation occurred between 10,500 and 10,300 yr B.P.

Elsewhere in northeastern and north-central Kansas, Bayne and Fent (1963) reported late Pleistocene fluvial deposits buried in the Republican River valley. Schmits (1980) identified, but did not radiocarbon date, a late Pleistocene fill along the Big Blue River; the oldest Holocene fill, which is inset against the late Pleistocene fill, dated to at least 6280 yr B.P. In western Kansas, Rogers (1984) mapped three late Pleistocene terraces along the Arkansas River and one along the Smoky Hill River, but provided no radiocarbon ages on terrace fills; ages were assigned on the basis of resident faunal remains.

Research in North and South Dakota has focused on the Missouri River and its tributaries. In South Dakota, Coogan and Irving (1959) recognized three late Quaternary terraces along the Missouri River. At several locations in North Dakota, Bickley (1972) mapped, but did not radiocarbon date, the late Pleistocene Oahe Formation. Along the Knife River, the lowermost (youngest) of three Pleistocene terraces was formed by an episode of incision about 13,000 yr B.P. (Reiten 1983).

On the periphery of the central Great Plains, Holliday (1987) mapped three terraces along the South Platte River in northeastern Colorado. Fill of the Kersey terrace, the oldest of the three, was deposited until 10,000 yr B.P., after which incision formed the terrace scarp. Also in northeastern Colorado, Stanford (1979) reported an episode of soil development around 10,500 yr B.P. To the east of the central Great Plains in northeastern Iowa, undated Pleistocene terrace remnants were cited by Hallberg and others (1984) as evidence of episodic, late Pleistocene river entrenchment.

The dearth of studies having basin-wide radiocarbon control has precluded the construction of a comprehensive late Pleistocene fluvial chronology for the central Great Plains. It appears that most sediments of this age have been eroded from or buried in valleys, leaving only scattered terrace remnants to piece together fluvial activity. Evidence gathered from these remnants suggests widespread entrenchment, concurrent with a period of regional surface stability and soil development, at or shortly before 10,500 yr B.P.

Because eolian deposits are well-preserved on uplands, much more is known about late Pleistocene eolian activity. In the central Great Plains, eolian sediments dating to the late Pleistocene (30,000 to 10,000 yr B.P.) have been divided into two loess units: a lower unit, designated the Gilman Canyon Formation, and an upper unit, designated the Peoria loess (Reed and Dreeszen 1965). The Gilman Canyon Formation was deposited and a paleosol developed in its upper portion between 34,800 and 21,290 yr B.P. (Dreeszen 1970; May and Souders 1988; May 1989a; D.W. May, Pers. Comm.

1989). Recent radiocarbon ages on the Gilman Canyon Formation include 24,270 ± 750 yr B.P. on the uppermost Ab and 33,590 ± 2260 yr B.P. on the lowermost Ab (both C13-corrected) in Phillips County, north central Kansas; 25,090 ± 590 yr B.P. (C13-corrected) on the middle portion of the uppermost Ab at the Eustis site in southwestern Nebraska; 19,640 +230/-240 yr B.P. on the upper Ab near Central City, Nebraska (Fredlund and Jaumann 1987); and 20,550 ± 590 yr B.P. (C13-corrected) on the uppermost Ab at Pratt County, south central Kansas (W.C. Johnson, Pers. Comm. 1990). At the type section, which is located at Buzzard's Roost in southeastern Lincoln County, Nebraska, the Gilman Canyon Formation was described as medium dark gray, noncalcareous, slightly humic silt (Reed and Dreeszen 1965).

Peoria loess is the major eolian deposit preserved in the vicinity of Harlan Lake. It is a buff, calcareous, massive, coarse silt-to-very fine sand that attains maximum thickness adjacent to major river valleys (Frye and Leonard 1951, 1952). Leonard (1951, 1952) divided the loess into basal, lower, and upper molluscan faunal zones. Descriptions of the Peoria loess frequently noted that the basal zone is leached, an indication of relatively slow deposition (Frye and Leonard 1952), and that it has a coarser texture than do the middle and upper zones (Lugn 1935). Sand horizons found within it are postulated to have been deposited during periods of relatively higher wind velocity (Prescott 1952). The clay mineralogy of the Peoria loess comprises montmorillonite, illite, quartz, and a trace of kaolinite (Swineford and Frye 1951; Beavers 1957).

Based on radiocarbon ages from its base, the Peoria loess is believed to have been deposited in the Great Plains between 21,000 and 14,000-to-12,000 yr B.P. (Thompson and Bettis 1980; Ruhe 1983). Recently, a radiocarbon age of 19,640 yr B.P. on Picea charcoal was obtained from the upper portion of the Gilman Canyon Formation below the base of the Peoria loess near Central City, Nebraska, a locality that lies about 200 km to the northeast of the Harlan Lake area (Fredlund and Jaumann 1987). Evidence that Peoria loess deposition was episodic comes from western Iowa (Daniels et al. 1960; Ruhe et al. 1971) and southwestern Illinois (McKay 1986), where the deposit contains dark, organic-rich bands that are thought to represent incipient soils formed during periods of slower deposition; differential abundance and preservation of fossil mollusks in the loess have also been cited as evidence of episodic loess deposition (Frankel 1957). The Brady paleosol developed in the upper portion of the Peoria loess between 12,700 and 8000 yr B.P. (Frye et al. 1968; Dreeszen 1970; Lutenegger 1985; May 1989a); recently obtained radiocarbon ages of 8850 yr B.P. on the uppermost Ab and 10,050 yr B.P. on the lowermost Ab (both C13-corrected) of the Brady paleosol in Phillips County, north central Kansas, fall within this range (W.C. Johnson, Pers. Comm. 1990). The geographic extent of this episode of pedogenesis is unknown.

Studies of the loess cover around the United States have revealed that particle size and cover thickness decrease rapidly with increasing distance from the loess source area (Swineford and Frye 1951; Waggoner and Bingham 1961; Frazee et al. 1970). In the central Great Plains, the Peoria loess attains a maximum thickness of 10 meters on the flanks of the Nebraskan Sand Hills (Thorpe and Smith 1952; Ruhe 1983). To the south and southeast, the cover thins and particle size decreases (Smith 1965; Lugn 1968). These findings led Smith

(1965) and Lugn (1968) to conclude that loess in central and south-central Nebraska was derived from the Sand Hills. More recently, however, Ahlbrandt and others (1983) reported Holocene radiocarbon ages on organic-rich sand underlying large dunes in the headwaters of the Loup River system adjacent to the Sand Hills. Based on these ages, they concluded that the Sand Hills date to the Holocene rather than to the Pleistocene, which would preclude them as the source of the Peoria loess. Corroboration for this conclusion comes from Muhs (1985), who reported widespread dune activity in northeastern Colorado during the late Holocene.

Elsewhere in the Sand Hills, Wright and others (1985) cited late Pleistocene radiocarbon ages on basal sediments in interdunal depressions as evidence that dunes have been stable since the end of the Pleistocene. Because large-scale, complex dunes such as those that constitute the Sand Hills result from long-term accumulations of sand (Lancaster 1988), it seems probable, as Bradbury (1980) postulated, that the Sand Hills formed during the late Pleistocene, and then underwent minor reworking during the Holocene.

While loess was being deposited in the central Great Plains, eolian sediments were accumulating in the northern Great Plains. In North Dakota, the Mallard Island member of the Oahe Formation was deposited prior to 13,000 yr B.P., followed by the Aggie Brown member between 13,000 and 8500 yr B.P. (Clayton et al. 1976; Jorstad et al. 1986). Along the Missouri River in South Dakota, a meter-thick mantle of eolian silt was deposited during the late Pleistocene or early Holocene (Ahler et al. 1974).

In summary, loess deposition was widespread across the central Great Plains during the late Pleistocene. Although inadequately dated, the record of this deposition is well documented, especially in Kansas and Nebraska. There, deposition of the Gilman Canyon Formation, which had begun by 35,000 yr B.P. (May 1989a), ended around 21,290 yr B.P. and soil development was initiated (D.W. May, Pers. Comm. 1989). The overlying Peoria loess was deposited from about 21,290 yr B.P. until sometime between 14,000 and 12,000 yr B.P.; this loess appears to have been contemporaneous with the eolian Mallard Island member identified in North Dakota. As indicated by the presence of incipient soils in loess, deposition was likely episodic. The thick cover of Peoria loess in central and south-central

Nebraska is believed to have been derived from the Sand Hills, although a minority of researchers have recently questioned the age of these hills. Following cessation of loess deposition, the Brady paleosol developed in the upper portion of the Peoria loess; radiocarbon ages on this soil range from 12,700 to 8000 yr B.P.

Holocene Landform Evolution

Early Holocene (10,000 - 7500 yr B.P.)

As might be expected, the Holocene sedimentation record, owing to its relative youthfulness and therefore better preservation, is documented more completely than that of the late Pleistocene. The fluvial record shows evidence of both synchronous and non-synchronous activity; the former is usually linked to major climate change (Wendland 1982; Knox 1983), whereas the latter is typically attributed to the complex response of individual basins to changes in external or internal variables (Schumm 1973.

Most research has suggested that an interval of widespread soil formation at the Pleistocene/Holocene boundary ended after 10,500 yr B.P. with a long-term, and relatively uninterrupted, episode of aggradation (Knox 1983); there is evidence from some basins that this alluviation was punctuated by brief episodes of incision (Hallberg et al. 1984). In drainage basins of Nebraska, Davis (1962), Brice (1964,1966), May and Holen (1985), and May (1989) also noted aggradation after 10,500 yr B.P. Along the Pomme de Terre River in Missouri and the Des Moines River in central Iowa, aggradation occurred from 11,000 to 8000 yr B.P. (Ahler 1976; Brakenridge 1981; Haynes 1985; Bettis and Hoyer 1986). Buchanan Drainage, a tributary to the Skunk River in central Iowa, underwent aggradation between 10,230 and 6300 yr B.P. (Van Nest and Bettis 1990). Along the Kansas River in Kansas, floodplain alluviation began shortly after 10,430 yr B.P. (Holien 1982; Johnson and Martin 1987), but appears to have slowed by 8300 yr B.P. (Johnson and Martin 1987).

Although most studies noted river aggradation during the early Holocene, research in Wyoming and South Dakota has uncovered evidence of widespread entrenchment during this period. At the Agate Basin site in eastern Wyoming, Albanese (1982) reported cutting and filling after 10,000 yr B.P. Along the Missouri River in South Dakota, Reiten (1983) noted incision until about 8000 yr B.P. Climatic conditions different than those that affected the central Great Plains during this time may have caused these disparate adjustments.

At the same time as early Holocene fluvial activity was taking place, the Bignell loess, which overlies the Brady paleosol, is postulated to have been blown from the floodplains of major rivers in the central Great Plains and deposited on uplands (Condra et al. 1947; Dreeszen 1970). This loess is described at the type section in Lincoln County, Nebraska, as gray silt about 2.5 to 3.0 m thick (Schultz and Stout 1945; Schultz and Martin 1970). It differs in color and molluscan faunal assemblage from the underlying Peoria loess (Leonard 1952), and has been identified in counties of north-central Kansas that are contiguous to Harlan County (Frye and Leonard 1949; Frye and Leonard 1952; W.C. Johnson, Pers. Comm. 1989). Research has suggested that deposition of the Bignell loess began shortly after 8000 yr B.P. (Lutenegger 1985; May 1989a). If so, it probably correlates with the early Holocene loess reported along the Missouri River in South Dakota (Ahler et al. 1974; McFaul 1985), and with the Aggie Brown member of the Oahe Formation in North Dakota (Jorstad et al. 1986). It may also correlate with early Holocene eolian activity documented in the Sand Hills of Nebraska and sand fields of neighboring states (Agenbroad 1977; Ahlbrandt et al. 1983; Muhs 1985), and with eolian sediments deposited in northern Texas around 9000 yr B.P. (Holliday 1987).

Middle Holocene (7500 - 4000 yr B.P.)

Although interrupted in some locations by a brief episode of incision, the aggradation that dominated the early Holocene continued during the middle Holocene. On the Loup River in Nebraska, May and Holen (1985) and May (1989) reported continued alluviation until sometime between 4780 and 3030 yr B.P. Along Medicine Creek and the Loup River of Nebraska, Brice (1964, 1966) noted aggradation until about 5000 yr B.P. Radiocarbon ages of 7800 and 7400 yr B.P. on paleosols developed in terrace fill along the Republican River, Nebraska, furnished additional evidence of middle Holocene floodplain aggradation (Libby 1955).

Along the Big Blue River in Kansas, Schmits (1980) identified two inset fills, the older radiocarbon dating to at least 6300 yr B.P., and the younger at 5800 to 4800 yr B.P. In the Pawnee River watershed, western Kansas, Mandel (1988) noted alluviation prior to 7000 yr B.P., and an episode of soil formation around 7000 yr

B.P. Rapid aggradation occurred between 7000 and 5000 yr B.P., followed by soil development around 5000 yr B.P. Entrenchment after 5000 yr B.P. then formed the scarp of the highest terrace. Aggradation was the dominant process along the Kansas River after 7250 yr B.P. (Bowman 1985). Except for a short period of floodplain stability and soil formation that occurred in some basins between 6500 and 6000 yr B.P. (Bettis and Hoyer 1986), rivers in central and western Iowa were aggrading until approximately 4000 yr B.P. (Hallberg et al. 1984; Bettis and Hoyer 1986; Bettis and Littke 1987).

In the Pomme de Terre River valley, Missouri, Haynes (1985) reconstructed a detailed middle Holocene fluvial record comprising several cut and fill cycles that dated between 11,000 and 1400 yr B.P. He reported alluviation between 8000 and 6300 yr B.P., followed by a short period of soil formation from 6300 to 6000 yr B.P. and renewed aggradation to 5000 yr B.P. Erosion then occurred between 5000 and 4600 yr B.P., followed by alluviation until 2900 yr B.P. In the same valley, Ahler (1976) noted erosion between 8000 and 7500 yr B.P., surface stability between 7500 and 6300 yr B.P., and rapid alluviation after 6300 yr B.P. Also in the same valley, Brakenridge (1980, 1981) reported incision from 8100 to 7500 yr B.P., followed by a period of surface stability to 5000 yr B.P., and then rapid aggradation after 5000 yr B.P.

Along the South Platte River in northeastern Colorado, Holliday (1987) reported deposition of the Kuner terrace fill prior to 3000 yr B.P. Along the Missouri River in North Dakota, Reiten (1983) documented aggradation from 8000 to 4500 yr B.P. Also along the Missouri River in North Dakota, Kornbrath (1975) noted a radiocarbon age of 5995 yr B.P. on charcoal buried in terrace fill.

In summary, following a brief and local episode of incision around 8000 yr B.P., aggradation dominated fluvial activity in the central Great Plains until approximately 6000 to 5000 yr B.P. In most basins, a short episode of erosion took place after 5000 yr B.P., followed by alluviation. At several locations in the central Great Plains, aggradation during the middle Holocene was interrupted by up to five intervals of floodplain stability and soil formation.

Recent studies have suggested that multiphase eolian activity took place in parts of the central Great

Plains during the middle Holocene. In Nebraska, Bradbury (1980) and Wright and others (1985) concluded that tracts of the Sand Hills were reworked during the middle Holocene, but rejected the conclusions of Ahlbrandt and others (1983) that the hills formed during the Holocene. Clayton and others (1976), Reiten (1983), and Jorstad and others (1986) postulated that eolian silt (the Pick City member of the Oahe Formation) was deposited between 8500 and 5000 yr B.P. on terraces and uplands in North Dakota. Paleosols developed in middle and late Holocene eolian silt along the Missouri River, North Dakota, provided strong evidence that eolian deposition was episodic (Jorstad et al. 1986). Near Great Bend and Pratt, south central Kansas, two paleosols developed in and buried by eolian silt have been radiocarbon dated at ca. 6500 yr B.P. At the Great Bend sand prairie between these two localities, a paleosol developed in loess and overlain by dune sand has been radiocarbon dated at ca. 5000 yr B.P. (W.C. Johnson, Pers. Comm. 1989). The age of 5000 yr B.P. fits closely with evidence of eolian activity documented in northern Texas between 5500 and 4500 yr B.P. (Holliday 1989).

Late Holocene (4000 yr B.P - Present)

The chronology of late Holocene fluvial activity appears to be more complex than chronologies of earlier fluvial activity. The cause of this greater complexity is uncertain, however. It may result from the relative youth, and therefore better preservation, of late Holocene deposits, or possibly it can be attributed to a greater frequency of climatic fluctuations during the late Holocene than during earlier periods. Certainly the relative youth of deposits affords a finer dating resolution, thereby adding detail and complexity to chronologies. That resolution notwithstanding, many late Holocene fluvial chronologies feature several episodes of aggradation alternating with floodplain stability and soil formation. This record stands in marked contrast to that of the early Holocene, which is characterized by evidence of aggradation interrupted by only one or two soil-forming intervals.

The late Holocene sedimentation record of the Kansas River basin contains abundant evidence of change in fluvial activity. In the Walnut Creek valley, Artz (1983) noted aggradation between 4500 and 2000 yr B.P., entrenchment at 2000 yr B.P., and finally aggradation and soil formation. Along Stranger Creek in northeastern Kansas, Logan (1985) documented alluviation between 4300 and 1300 yr B.P. Schmits (1980) mapped two alluvial fills, one dating to about 2400, the other to pre-1000 yr B.P., along the Big Blue River in northeastern Kansas. Research on the Kansas River has produced radiocarbon ages of 4290, 2600, 2395, 1600, and 1200 yr B.P. on paleosols buried in Holliday terrace fill (Bender et al. 1980; Holien 1982; Johnson 1985), and alluvium underlying the Kansas River floodplain has yielded a radiocarbon age of 785 yr B.P. (W. Dort, Jr., Pers. Comm. 1984). These ages demonstrate that fill of the Holliday terrace had begun to accumulate by at least 4300 yr B.P. Incision to form the terrace scarp then occurred sometime between 1200 and 785 yr B.P. The presence of paleosols is also evidence that late Holocene aggradation in the Kansas River basin was episodic (Johnson and Martin 1987).

In the Pawnee River basin of western Kansas, Mandel (1988) reported aggradation between 5000 and 2600 yr B.P., followed by an episode of soil formation between 2600 and 2000 yr B.P. Incision followed by aggradation took place between 2000 and 1300 yr B.P., and then soil formation occurred until approximately 1000 yr B.P.;

alternating incision and aggradation have dominated fluvial activity during the last 1000 years.

In northern Colorado along the South Platte River, Holliday (1987) reported incision after 3000 yr B.P., followed by alluviation during the last 1000 years. Along Medicine Creek in Nebraska, Brice (1966) noted aggradation of the Mousel terrace fill between 4000 and 1000 yr B.P., incision after 1000 yr B.P., and then deposition of the fill that underlies the lowermost terrace. In the South Loup River valley, Nebraska, May and Holen (1985) and May (1989) documented incision prior to 3030 yr B.P., followed by aggradation, and finally another episode of incision sometime before 2080 yr B.P. Aggradation then took place until around 300 yr B.P., at which time incision occurred. Along the Calamus River, a small tributary in the Loup River drainage, Falk and Pepperl (1980) reported an episode of soil formation between 1330 and 1200 yr B.P., an interval that matches closely the 1660 to 930 yr B.P. period of soil development noted by May (1989).

In North and South Dakota, work has been concentrated on the Missouri River and its tributaries. In South Dakota, Coogan and Irving (1959) found late Prehistorical artifacts in fill of the lowermost terrace along the Missouri River. Along Bear Creek in southwestern South Dakota, Harksen (1974) reported radiocarbon ages of 2350 and 780 yr B.P. on two inset fills. In North Dakota on the Knife River, Reiten (1983) documented an episode of incision concurrent with a period of soil formation between 4500 and 2500 yr B.P. Aggradation then occurred from 2500 to 500 yr B.P. During the last 500 years, the river has been alternately cutting and filling its channel. Also in North Dakota, Hamilton (1967, 1967a) reported the deposition of five units of fill along Jones Creek, a tributary to the Missouri River, since approximately 250 yr B.P.

Along the Pomme de Terre River in Missouri, Haynes (1985) documented four episodes of late Holocene cutting and filling, each spanning 300 to 600 years; periods of cutting, which were brief in contrast to intervals of filling, occurred about 2900, 1800, 1000, and 430 yr B.P. Deposition has dominated the basin for the last 250 years (Haynes 1985). At Delaware Canyon, Oklahoma, Ferring (1986), Hall (1987), and Hall and Ferring (1987) reported rapid aggradation between 2700 and 1900 yr B.P. and 700 and 600 yr B.P.; this alluviation was interrupted by entrenchment around 1000 yr B.P. and during the last 100 years. Ferring (1986), Hall (1987), and Hall and Ferring (1987) also identified episodes of soil formation between 1900 and 1000 yr B.P. (Caddo paleosol) and 600 and 385 yr B.P. (Delaware Canyon paleosol). In Iowa, the Buchanan Drainage experienced alluviation between 3000 and 2000 yr B.P. (Van Nest and Bettis 1990). It has been noted that valley fill immediately adjacent to many streams in western and central Iowa has accumulated during the past 2000 years (Ruhe 1969; Bettis and Thompson 1981). This late Holocene fill has buried middle and early Holocene fluvial deposits (Van Nest and Bettis 1990).

In summary, the fluvial activity of the late Holocene, which consisted of several episodes each of alluviation, entrenchment, and soil formation, stands in marked contrast to the relatively uninterrupted aggradation that characterized the early Holocene. Although synchrony is not evident among all events, it is present among some of them, most notably those of the early part of the late Holocene. With the exception of a few basins where incision was recorded, aggradation appears to have been the dominant fluvial activity from about 4500 to 3000 yr B.P. in several basins, and from

4500 to 2000 yr B.P. in others. Most studies noted a short episode of entrenchment following this aggradation, and then renewed deposition, although the timing of these two events varied across the central Great Plains. It is impossible to identify much synchrony among fluvial activity of the last 2500 years. This lack of synchrony indicates that factors in addition to regional climate change (e.g., water table level fluctuations, internal variables) must be considered when attempting to explain some of the late Holocene fluvial activity documented in the central Great Plains. It should also serve to caution those workers who seek to derive regional climatic conditions from the alluvial record of one basin. Episodes of soil formation, which occurred at different places around 2600 to 2400, 2100 to 1600, and 1200 yr B.P., correspond to periods when alluviation slowed and surface stability took place. Several short-term erosional and depositional events occurred in some basins between approximately 1000 yr B.P. and the present.

Episodic eolian sedimentation that occurred during the middle Holocene apparently continued into the late Holocene. In North Dakota, Jorstad and others (1986) reported eolian deposition, interrupted by short periods of soil formation, during the last 5000 years; these deposits constitute the Riverdale member of the Oahe Formation. Reiten (1983) documented eolian sedimentation along the Knife River in North Dakota between 2500 and 500 yr B.P. Eolian silt caps containing cultural artifacts that date between 900 and 600 yr B.P. were mapped by Coogan and Irving (1959) on the middle terrace of the Missouri River in South Dakota; elsewhere in South Dakota, McFaul (1985) identified similar eolian deposits mantling terraces on the Missouri River.

There is also strong evidence of late Holocene dune activity at several locations in the central Great Plains. Muhs (1985) postulated sand dune movement between 3000 and 1500 yr B.P in northeastern Colorado, activity that appears to have been contemporaneous with the late Holocene dune formation that Ahlbrandt and Fryberger (1980) and Ahlbrandt and others (1983) noted in Nebraska. In Lincoln County, Nebraska, cedar stumps, which radiocarbon dated to 400 yr B.P., were uncovered in eolian sand (Weakley 1962). A similar find is noted by Martin (1985), who found recently excavated cedar stumps that dated to "modern" in a small gully in the

Medicine Lodge River basin, south-central Kansas.

In summary, it appears that eolian activity was an important component of late Holocene upland landform evolution in those areas of the central Great Plains that have sandy surface materials. Sand dunes appear to have been active in northeastern Colorado and Nebraska between 3000 and 1500 yr B.P. Findings from North Dakota suggest that eolian sedimentation, interrupted periodically by soil formation, occurred after 5000 yr B.P. In south-central Nebraska, eolian sand deposition took place around 400 yr B.P. Based on the limited record of upland activity in the central Great Plains, it appears that late Holocene eolian sedimentation on uplands was contemporaneous with fluvial aggradation in valleys.

Colluvial Deposits

In addition to fluvial and eolian deposits, several studies have noted the existence of colluvium on river valley sides in the central Great Plains. Colluvial deposits are believed to be transported by hillslope wash from upland deposits to the valley floor (Leopold et al. 1966). In Iowa, valley side deposits comprising redeposited loess were radiocarbon dated at 8000 to 3000 yr B.P. (Thompson and Bettis 1980; Bettis et al. 1984; Bettis and Hoyer 1986; Bettis and Littke 1987). Moving away from the valley wall across the floodplain towards the active channel, these deposits graded into fluvial deposits (Bettis and Littke 1987). Accumulation of valley side deposits was apparently episodic, as indicated by the presence within them of multiple weakly-to-moderately developed paleosols (Bettis et al. 1984; Bettis and Hoyer 1986). The episode of hillslope instability during which valley side deposits accumulated is believed to have been caused by a period of enhanced westerly flow that affected the central Great Plains between 9000 to 6000 yr B.P. (Bettis and Hoyer 1986); such flow is postulated to have extended the rainshadow of the Rocky Mountains eastward, spreading dry conditions to the interior of North America (Wendland 1978). In valleys during this period, the poorly-dated fluvial record indicates generally continuous aggradation (Knox 1983). After 3000 yr B.P., an increase in the vegetation cover resulting from a return to more mesic conditions likely restored hillslope stability. An episode of incision noted about 3000 yr B.P. in several drainage basins of the central

Great Plains truncated the valley side deposits, forming a terrace that trends smoothly from valley margins to uplands.

Summary

Although research of the last twenty years has measurably increased knowledge of late Quaternary sedimentation and erosional activity in the central Great Plains, sizable gaps remain in the chronology of landform evolution for this period. For example, owing to a paucity of radiocarbon ages, little is known, about the late Pleistocene and early Holocene fluvial record (Knox 1983). Evidence uncovered to date suggests river entrenchment at the end of the Pleistocene, followed by alluviation, and finally an episode of soil development at the Pleistocene/Holocene boundary. On uplands, deposition of the eolian Gilman Canyon Formation ended, and an episode of soil development was initiated, about 21,000 yr B.P. Beginning shortly after 21,000 yr B.P., the Peoria loess was deposited, burying the Gilman Canyon Formation. Deposition continued until 14,000-to-12,000 yr B.P., at which time the Brady Paleosol developed in the Peoria loess; radiocarbon ages on this paleosol range from 12,700 to 8000 yr B.P. The Peoria loess in the central Great Plains is generally assumed to have been blown out of the Nebraska Sand Hills, although a recent study has postulated that the hills formed in the Holocene, not in the Pleistocene (Ahlbrandt and others 1983). While acknowledging that some minor reworking of dunes likely occurred during the Holocene, most researchers continue to support the theory that the Sand Hills formed during the Pleistocene (e.g., Bradbury 1980; Wright et al. 1985).

Fluvial activity during the early Holocene was dominated by aggradation, although in some basins a short episode of incision preceeded deposition. On uplands, deposition of the Bignell loess, and its eolian counterparts in the Dakotas and northern Texas, is thought to have begun about 8000 yr B.P. There is also evidence from northwestern and central Nebraska of sand dune formation during the early Holocene. Around 8000 yr B.P., an episode of floodplain stability occurred in several drainage basins; other basins, however, experienced incision at this time. Following this, alluviation, alternating with short periods of surface stability, took place; valley side deposits accumulated in some localities between 8000 and 3000 yr B.P. Aggradation slowed, or ceased, between 6000 and 5000 yr

B.P., and incision occurred. Eolian sedimentation during the middle Holocene has been recorded at scattered locations around the central Great Plains.

In contrast to the record of the middle and early Holocene, which contains evidence of relatively uninterrupted aggradation, the late Holocene fluvial record is characterized by several erosional and depositional events, most of which are non-synchronous. The absence of synchrony suggests that factors in addition to regional climate change must be considered when attempting to explain late Holocene fluvial activity. The eolian record contains evidence of late Holocene dune formation in northeastern Colorado and northwestern Nebraska, and eolian sedimentation in North Dakota; the activity in Colorado and Nebraska appears to have been contemporaneous with fluvial aggradation.

CHAPTER 3: LATE QUATERNARY PALEOENVIRONMENTAL RECORD OF THE CENTRAL GREAT PLAINS

Methods of Paleoenvironmental Reconstruction

Current knowledge of late Quaternary climatic conditions in the central Great Plains is based on a limited, but rapidly expanding, paleoenvironmental record. This record consists of evidence from surrogate indicators such as micro- and macrobotanical fossils, landsnails, opal phytoliths, and vertebrate and invertebrate remains. Because single indicators generally fail to produce a clear signal of paleoenvironmental change, researchers frequently use more than one surrogate indicator in reconstructions. Although more time-consuming and expensive, such a procedure usually affords a broader paleoecological spectrum from which to gauge environmental and climatic change.

Notwithstanding its limitations, palynology remains the technique used most widely in the reconstruction of late Quaternary environmental conditions (e.g., Davis 1963; Delcourt and Delcourt 1980; Jacobson and Bradshaw 1981; Bryant and Holloway 1985; Fall 1987). Typically, bogs and lakes serve as ideal preservational sites for pollen; unfortunately, both are rare in the subhumid central Great Plains. In eolian and fluvial deposits, the materials that mantle much of the central Great Plains, pollen preservation is generally poor. Further, because pollen does not deteriorate uniformly (Sangster and Dale 1964; Cushing 1967; Havinga 1967), certain taxa may be misrepresented in the record. Well-preserved pollen is present in and has been recovered from fluvial deposits. It is unclear, however, whether pollen assemblages extracted from such deposits are representative of the regional vegetation cover (Fall 1987; Hall 1989). Consequently, most of our knowledge about paleoenvironmental conditions in the central Great Plains derives from work completed in contiguous states (e.g., Missouri, Iowa, and Minnesota). In these states, which are located on the humid, eastern periphery of the region, sites conducive to pollen preservation are more common.

In lieu of poorly preserved pollen and other floral remains, researchers in the central Great Plains have turned to other surrogate indicators in their reconstruction of paleoenvironmental conditions. Some have advocated the use of inorganic biogenetic plant particles (phytoliths) to document vegetation change

(Twiss et al. 1969; Rovner 1971; Brown 1984). Phytoliths, which are composed of silica that precipitated around plant cells (Rovner 1971), accumulate in eolian and fluvial deposits, and are highly resistant to dissolution. They have proven useful in detecting the presence of grasses (Rovner 1971; Kurmann 1985), and in differentiating between C4 (warm and dry region) and C3 (cool and moist region) grasses (Twiss 1987). Because their hard shell renders them resistant to the aeration present in loess deposits, terrestrial landsnails have also proven to be a useful surrogate indicator of late Quaternary environmental conditions in the central Great Plains. Approximately thirty species, most of which are extinct today in the central Great Plains, have been identified in the Peoria loess of the region (Leonard 1952; Wells and Stewart 1987), often in great abundance (Leonard 1952).

The central Great Plains has also produced among the largest quantities of fossil faunal remains (Schultz and Martin 1970). As noted by many researchers (e.g., Hibbard 1970; Hoffman and Jones 1970; Graham 1987; Martin and Hoffman 1987), the character of a faunal province is determined by regional climatic conditions.

In locating actual boundaries between late Quaternary faunal provinces, however, one must consider additional variables such as topography, soils, frequency of precipitation, the occurrence of fire, and the impact of human activity (Martin and Klein 1984; Martin and Hoffman 1987). Moreover, faunal sequences do not always reflect short-term fluctuations in climate (see Graham 1987). In spite of these limitations, researchers, by comparing fossil faunal remains to modern analogues, have reconstructed paleoenvironmental conditions in several parts of the central Great Plains (e.g., Hibbard 1970; Hoffman and Jones 1970; Lundelius et al. 1983; Graham 1987; Stewart 1989).

Finally, several scholars have used the characteristics of paleosols to make some inferences about paleoenvironmental conditions (e.g., Simonson 1954; Ruhe 1969, 1974; Morrison 1978; and Mahaney 1981). The degree of soil development has been linked to the warmth of the climate (Morrison 1978); the amounts of kaolinite and halloysite contained in a paleosol have been related to leaching intensity (Mahaney 1981); the color of a paleosol has been associated with a particular vegetation community (Simonson 1954; Ruhe

1969, 1974); and the weathering ratios of heavy-to-light minerals have been used as an indicator of weathering intensities (Ruhe 1969). There are some problems, however, associated with the derivation of paleoclimatic conditions from paleosols, including secondary enrichment of buried paleosols (e.g., Ruhe 1956; Mausbach et al. 1982); the postburial alteration of chemical properties (e.g., Stevenson 1969; Valentine and Dalrymple 1976; Schaetzl and Sorenson 1987); and the difficulty in controlling for the impact of non-climatic factors (i.e., parent material, time, topographic position) on pedogenesis and the resultant soil profile (e.g., Scholtes et al. 1951).

Late Quaternary Paleoenvironmental and Climatic Conditions

Much of our knowledge about late Quaternary paleoenvironmental conditions in the central Great Plains derives from work completed in the last two decades. We can now reconstruct generalized late Pleistocene vegetation patterns for the region, but our knowledge of Holocene patterns is limited. Even less is known about the nature of the climatic fluctuations that dictated these patterns. My objective in the following review is to provide a summary of paleoenvironmental and climatic conditions in the central Great Plains during the last 30,000 years. The text is arranged chronologically using the temporal categories proposed in Chapter 2.

Late Pleistocene (30,000 - 10,000 yr B.P.)

As detailed in Chapter 2, the Gilman Canyon Formation accumulated and then the Peoria loess was deposited over much of the central Great Plains between about 35,000 and 14,000-to-12,000 yr B.P. Although early studies often postulated that late Pleistocene loess sheets accumulated during drought conditions (e.g., Schultz and Stout 1948), evidence from modern depositional environments suggests that loess can accumulate only on well-vegetated surfaces (Yaalon and Dan 1974). Moreover, the landsnail fauna contained in the Peoria loess of the central Great Plains implies deposition on a well-vegetated surface (Shimek 1930; Leonard 1952; Wells and Stewart 1987a).

The floral record also suggests that mesic conditions existed in the central Great Plains during loess deposition. In loess deposits at two locations in Kansas, Wells and Stewart (1987, 1987a) found needle leaves of _Pinus flexilis_ (limber pine) and _Picea_ (spruce) charcoal dated at 14,450 yr B.P. At the North Cove paleoindian site in south-central Nebraska, Wells and Stewart (1987) recovered cones, needles, and wood of _Picea glauca_ (white spruce) from spring deposits; ages on the wood ranged from 14,700 to 13,100 yr B.P. (Johnson 1989). A preliminary grass phytolith analysis of loess exposed at the Eustis Pit, Frontier County, southwestern Nebraska, indicated that Peoria loess was deposited during periods of more effective moisture, whereas soil formation occurred during periods of less effective moisture (Fredlund et al. 1985). Floral data from Sander's well, an upland site in southeastern Kansas, and from the North Cove site on Harlan Lake, south central Nebraska, implied that a _Populus_ (aspen) parkland existed on uplands between 24,000 and 12,800 yr B.P. (Fredlund and Jaumann 1987; Fredlund 1989). Ruhe (1983), on the basis of conifer wood found in loess deposits in Iowa, concluded that trees were present during deposition of the Peoria loess. Finally, in a survey of the pollen record, Webb and others (1983) and Baker and Waln (1985) postulated that treeless openings, perhaps a _Pinus_ parkland, existed on the central Great Plains during the late Pleistocene.

Faunal remains also suggest that steppe or parkland vegetation existed on the central Great Plains during the time of loess deposition. In northeastern Colorado, the Selby and Dutton fauna, which were extracted from Peoria loess that dates to approximately 16,000 yr B.P., are characteristic of an open grassland environment with few, if any, trees; in the same fauna, the appearance of Mammuthus columbi (Columbian mammouth) in lacustrian deposits that date to between 16,000 and 12,000 yr B.P. implies an increase in effective moisture, and perhaps additional tree cover (Graham 1987). Loess deposits in southwestern Iowa that date between 24,000 and 14,000 yr B.P. have yielded small mammal fauna typical of grassland areas (Rhodes 1984). Prairie and steppe fauna such as Cynomys (both white- and black-tailed prairie dog) have been recovered from Peoria loess deposits across the central Great Plains (Hoffman and Jones 1970). A late Pleistocene grassland fauna comprising Mammuthus sp. (mammouth), Equus (horse), and Bison antiquus (bison) was excavated from deposits along the Republican River, Nebraska (Corner 1977). In Peoria loess at the North Cove paleoindian site, Nebraska, Stewart (1989) identified Thomomys talpoides (northern pocket gopher), Spermophilus *kimballensis* (kimballs ground squirrel), and *Bootherium bombifrons* (woodland muskox), fauna that typically reside in an environment that is cooler and more mesic than that which is extant at present in Nebraska.

The most detailed account of late Pleistocene environmental conditions in the central Great Plains comes from palynological studies undertaken along the eastern and northern periphery of the region. *Picea* was the dominant component of the vegetation cover in the Sand Hills of South Dakota until 12,500 yr B.P., at which time *Pinus* and herbaceous pollen increased (Watts and Wright 1966). At Muscotah and Arlington Marshes in northeastern Kansas, Gruger (1973) documented a *Picea* forest from 23,000 to 15,000 yr B.P., followed at 11,500 yr B.P. by a slow decline in *Picea* and a rise in mixed prairie and deciduous tree species such as *Ostrya* (ironwood), *Carylus* (hazlenut), and *Quercus* (oak). To the south in southeastern Kansas, the *Picea* forest was transitional to a mosaic of prairie and *Quercus*/*Carya* (hickory) forest that existed until about 22,000 yr B.P. Around 22,000 yr B.P., this mosaic evolved to a mixed *Picea*/deciduous forest in river valleys and a *Populus* parkland on uplands. The *Picea* population declined in the central Great Plains between 12,000 and 10,500 yr B.P., and by 10,500 yr B.P. *Picea* had disappeared from the region. After 10,500 yr B.P., deciduous species such as *Salix* (willow), *Ulmus* (elm), *Fraxinus* (ash), and *Quercus* dominated the vegetation cover of the central Great Plains (Fredlund and Jaumann 1987).

Paleoenvironmental studies in Iowa, although portraying vegetation under more mesic climatic conditions than those that likely existed in the central Great Plains, have produced a similar picture of late Quaternary vegetation and climatic change. Much of northern Iowa apparently supported a *Picea*-dominated forest during the late Pleistocene (Brush 1967; Durkee 1971). At Lake West Okoboji in northwestern Iowa, a *Picea* and *Larix* (Tamarack) dominated forest existed from 14,000 to 11,800 yr B.P. After 11,800 yr B.P., there was a sharp decrease in *Picea* and *Larix*, and an equally abrupt rise in *Quercus* and *Ulmus* (Van Zant 1979). In southwestern Iowa, *Picea* was present in high concentrations until approximately 11,800 yr B.P. (Baker et al. 1980).

Finally, several researchers have used soil characteristics in attempting to derive late Pleistocene paleoenvironmental conditions. Based on the nature and sequence of horizons in buried soils in the upper Mississippi River valley, Simonson (1954) concluded that the Yarmouth soil (pre-Illinoian in age) developed under a forest vegetation in a climatic regime similar to that of the present. This conclusion was based on a comparison of the horizon thickness, clay mineral types, and clay content of the buried soil with those same characteristics of the surface soil. The podzolic nature of the Sangamon paleosol has been attributed to its formation under a forest vegetation cover (Thorpe et al. 1951). Humus forms present in paleosols developed in western European loess deposits were used to conclude that moist conditions existed in the late Pleistocene (Dalyrymple 1958). Ruhe (1969) suggested that the red-yellow color of the Sangamon and late Sangamon paleosols may have resulted from development under conditions warmer and wetter than those of the present.

In summary, evidence indicates that the Peoria loess accumulated on a well-vegetated surface rather than on a barren one. Many landsnail species preserved in the Peoria loess are extant today in litter-covered forested areas of the Rocky Mountains. Faunal remains characteristic of grasslands have been recovered from loess in Iowa, Nebraska, and elsewhere in the central Great Plains. A preliminary grass phytolith analysis of Pleistocene loess at the Eustis Pit, southwestern Nebraska, indicates that Peoria loess accumulated under cool and mesic conditions. Macro- and microfloral evidence suggests that a parkland vegetation community may have existed on uplands of the central Great Plains during loess deposition, 24,000 to 12,000 yr B.P. Limited data from southeastern Kansas indicates that a more continuous Picea/deciduous cover was present during this time in some river valleys. East of the central Great Plains, a more continuous Picea-dominated forest apparently existed from about 24,000 to 12,000 yr B.P. In response to warming climatic conditions after 12,000 yr B.P., there was a sharp decline in Picea and a concommitant rise in deciduous species across the central Great Plains.

In addition to being cooler, there is evidence that the late Pleistocene was characterized by reduced seasonal temperature extremes. The disappearance of the late Pleistocene landsnail fauna from the central Great Plains around 12,500 yr B.P. is generally ascribed to the onset of a climate characterized by high temperatures and more xeric conditions; that is, climatic conditions similar to those present in the region today (Leonard 1952; Wells and Stewart 1987). Other aspects of the faunal record have led several investigators to conclude that late Pleistocene biological communitites were more complex and diverse than those of the Holocene (Taylor 1965; Martin and Neuner 1978; Martin and Hoffman 1987; Martin and Martin 1987). A more equable climate featuring cool summers and warm winters is postulated to have fostered this complexity and diversity (Martin and Neuner 1978; Martin and Hoffman 1987; Martin and Martin 1987). The transition to a less equable climate (i.e., increased seasonality) at the close of the Pleistocene destroyed these complex biological communities, and brought about the extinction of many mega-fauna (Martin and Neuman 1978; Martin and Hoffman 1987).

Recent climatic models based on atmospheric variables and orbital parameters have largely agreed with conclusions drawn from the paleoenvironmental record. For the central Great Plains, Kutzbach and Wright (1985) postulated drier conditions coupled with slightly cooler January and significantly cooler July temperatures (i.e., narrowed seasonal temperature range) during the late Pleistocene than exist in the region today. In another climatic reconstruction, Kutzbach and Guetter (1986) argued that reduced seasonal temperature extremes and cold, dry conditions dominated the climatic pattern of the interior of North America during the late Pleistocene. Their model concluded that the seasonal temperature range began to widen around 12,000 yr B.P. because of an increase in annual radiation extremes. Although omitting a discussion of temperature seasonality, a third model of atmospheric activity, the COHMAP (Cooperative Holocene Mapping Project) model, reported warming of the climate in North America around 12,000 yr B.P. as summertime solar radiation began to increase (COHMAP Members 1988).

Early Holocene (10,000 - 7500 yr B.P.)

The paleoenvironmental record, albeit limited, suggests that the transition from the Pleistocene to the Holocene was a period of major climatic and vegetational change across the central Great Plains. At Muscotah Marsh in northeastern Kansas, Gruger (1973) recognized a sharp decline in *Picea* after 12,000 yr B.P., and a concurrent rise in *Quercus*, *Ulmus*, *Fraxinus*, and *Salix*. By 10,500 yr B.P., *Picea* had been replaced by deciduous species. The deciduous forest persisted until grasslands expanded about 9000 yr B.P. At the Rosebud site in the Sand Hills of South Dakota, Watts and Wright (1966) noted an increase in Pinus and herbaceous pollen during the early Holocene. The pollen record from Swan Lake in the Nebraska Sand Hills showed marsh vegetation during the early Holocene, an indication of relatively xeric conditions (Wright et al. 1985). At the Siebold Slough in North Dakota, Cvancara and others (1971) reported a decline in Picea pollen and a concomitant increase in grass pollen about 9700 yr B.P. And in central Oklahoma, prairie indicators such as Poaceae and Asteraceae increased in abundance around 11,000 yr B.P. (Wilson 1966).

The vertebrate and invertebrate records of the early Holocene also contain evidence of major climatic change. Fauna such as Blarina brevicauda (short-tailed shrew), which had been present in the region as recently as the Sangamon interglacial, reappeared in the central Great Plains between 13,000 and 10,500 yr B.P., an indication of climatic warming (Hoffman and Jones 1970). There was also a change to the modern landsnail assemblage between the end of Peoria loess deposition (about 13,000 yr B.P.) and the start of Bignell loess deposition (about 8000 yr B.P.), a change attributed to the onset of a less equable climate at the end of the Pleistocene (Leonard 1952).

A more extensive paleoenvironmental record of the Pleistocene/Holocene transition is available along the eastern periphery of the central Great Plains. Deciduous forest comprising predominantly *Quercus* and *Ulmus* covered the area around Lake West Okoboji, northwestern Iowa, from 11,800 to 9000 yr B.P. Probably in response to post-glacial warming and increased aridity, this forest was gradually replaced around 9000 yr B.P. by prairie grasses and shrubs, most notably Gramineae, *Artemesia*, and *Ambrosia* (Van Zant 1979). In north-central Iowa, increasingly xeric conditions after 8000 yr B.P. likely caused the decline in *Quercus* and *Ulmus* and the increase in *Ambrosia* (Brush 1967; Durkee 1971). In Iowa, micromammal fossils testify to increased aridity during the early Holocene and establishment of the prairie by 8400 yr B.P. (Semken 1980; Rhodes and Semken 1986).

In summary, there is conclusive evidence of vegetational change during the early Holocene on the eastern periphery of the central Great Plains. Although the paleoenvironmental record for this period in the central Great Plains is much more limited, the available data do suggest some change in vegetational communities, specifically a shift to a grassland vegetational community. On the eastern periphery of the central Great Plains, floral and faunal records show that prairie moved eastward across the central Great Plains between 9000 and 8000 yr B.P., eventually extending into Iowa by 8000 yr B.P. (Leonard 1952; Gruger 1973; Van Zant 1979; Semken 1980; Wright et al. 1985; Rhodes and Semken 1986). These changes in vegetational and faunal assemblages are thought to reflect a shift to warmer temperatures and increased aridity (Webb et al. 1983), a change that likely resulted from increased summer insolation (COHMAP Members 1988). As documented in the previous chapter, this climatic and vegetational shift is marked in the early Holocene sedimentation record by widespread fluvial aggradation. The coupling of fluvial deposition to the onset of xeric conditions has been postulated elsewhere (e.g., Huntington 1914; Love 1979; Brakenridge 1981, 1984; Knox 1983).

Middle Holocene (7500 - 4000 yr B.P.)

The most widely reported event of the middle

Holocene is the Altithermal, a period when climatic conditions were apparently drier than those of the present. First noted by Antevs (1955), the Altithermal is believed to have resulted from the dominance of zonal atmospheric flow, a pattern that increased the frequency of dry Pacific air over the central Great Plains (Wendland 1978). According to some studies, this period of dry climatic conditions affected much of the central Great Plains from approximately 7000 to 5000 yr B.P., and may have been severe enough to reduce human populations in the region (Hurt 1966; Frison 1975; Benedict 1979; Wedel 1986). Conversely, others have argued that the cultural hiatus in the archaeological record is not the result of a depopulation of the central Great Plains, but rather the product of insufficient research in the region (see Reeves 1973).

According to the paleoenvironmental record, the tall grass prairie (apparently in response to the drought) migrated eastward during the period to areas that at present support mixed deciduous-prairie vegetation. At Lake West Okoboji in northwestern Iowa, Van Zant (1979) reported a rapid decline in Quercus/Ulmus after 9000 yr B.P., concomitant with a rise in, and eventual dominance of, prairie species such as Gramineae, <u>Artemesia</u>, and <u>Ambrosia</u>. Prairie vegetation covered the area between 7700 and 3200 yr B.P. Van Zant (1979) also identified an influx of coarse sediment to Lake West Okoboji between 7730 and 6200 yr B.P., activity that he interpreted as evidence of a dry, or nearly-dry, lake. Micromammal remains from the Cherokee Sewer site in northwestern Iowa also indicate increased aridity between 8400 and 6300 yr B.P. (Semken 1980), followed by more mesic conditions after 5000 yr B.P. (Rhodes and Semken 1986). In the pollen record at the Siebold Slough, North Dakota, Cvancara and others (1971) reported dry conditions between 8500 and 4500 yr B.P. Fossil diatom assemblages and pollen profiles from southeastern Minnesota showed dry conditions between 8000 and 4000 yr B.P., and maximum eastward extension of the prairie about 7200 yr B.P. (Wright 1976; Brugam 1980; Brugam et al. 1988). In northeastern Kansas, Gruger (1973) noted the dominance of prairie species <u>Ambrosia</u> and <u>Artemesia</u> between 9900 and 5000 yr B.P.; about 5000 yr B.P., the present mixed deciduous/prairie mosaic began to develop.

In contrast to the middle Holocene paleoenvironmental record of areas peripheral to the central Great Plains, a record that reveals an expansion of prairie vegetation, the middle Holocene record from the central Great Plains lacks evidence of major climatic and vegetational change. Either the record is too limited to reveal the climatic and vegetational changes, or the changes did not affect the surrogate indicators of vegetational change that have been studied to date. At Swan Lake in the Nebraska Sand Hills, Wright and others (1985) noted relatively low water levels from 8900 to 3700 yr B.P., but could not discern any significant fluctuations in pollen frequencies. At Hackberry Lake in the Nebraska Sand Hills, Sears (1961) cited an increase in arboreal pollen around 5000 yr B.P. as evidence of more mesic conditions during the middle Holocene.

Evidence of arid conditions is, however, preserved in the eolian sedimentation record of the middle Holocene. Parts of the Nebraska Sand Hills and adjacent areas experienced eolian activity during the period (Agenbroad 1977; Ahlbrandt and Fryberger 1980; Bradbury 1980; Ahlbrandt et al. 1983; Muhs 1985; Wright et al. 1985). Findings from northern Texas indicated eolian sedimentation between 5500 and 4500 yr B.P. (Holliday 1989). In central Kansas, paleosols developed in and buried by eolian silt have been radiocarbon dated at about 6500 yr B.P., and at the Great Bend sand prairie, a paleosol developed in eolian silt and buried by dune sand has been radiocarbon dated at approximately 5000 yr B.P. (W.C. Johnson, Pers. Comm. 1989). This middle Holocene eolian activity likely resulted from increased aridity and an attendant reduction in the vegetation cover.

The fluvial record of the middle Holocene reveals aggradation in the central Great Plains between 8000 and 6000-to-5000 yr B.P. (Johnson and Logan, In press). This episode of aggradation may have resulted from drought conditions that reduced the vegetation cover and increased sediment delivery to rivers (Knox 1983). As documented in studies of the prairie during the 1930s, arid conditions can rapidly and dramatically diminish grassland cover (Albertson and Weaver 1942; Weaver and Albertson 1943; Weaver and Bruner 1948; Tomanek and Hulett 1970), leaving the surface exposed to raindrop impact and the effects of overland flow; with the return of plentiful moisture, the grassland cover is rapidly reestablished (Weaver and Albertson 1943; Tomanek and Hulett 1970).

Atmospheric models have also generated drought conditions during the middle Holocene. The AGCM (Atmospheric General Circulation Model) developed by Kutzbach and Guetter (1986) and the COHMAP model (COHMAP Members 1988) postulated maximum summer warmth in the northern hemisphere around 6000 yr B.P.; a strengthening of westerly flow across North America is thought to have caused these warm and dry conditions (COHMAP Members 1988). Moisture availability was further reduced by the increased evaporation that accompanied such warmth.

In summary, paleoenvironmental records from across the central Great Plains show a dominance of prairie vegetation and xeric conditions during the middle Holocene. The prairie apparently expanded eastward, reaching its limit in eastern Minnesota and western Wisconsin around 7200 yr B.P. The dry climatic conditions may have been of sufficient severity to reduce human populations in the region. Maximum aridity occurred between 6000 and 5000 yr B.P., followed by a return to more mesic conditions after 5000 yr B.P. The dry conditions, by reducing the vegetation cover, likely accelerated eolian activity in the region.

Late Holocene (4000 yr B.P. - Present)

As noted in several studies, frequent changes in flood magnitude and frequency are preserved in the late Holocene sedimentation record (e.g., Knox et al. 1981; Haynes 1985; May 1986). Dominance of meridional atmospheric circulation, which results in high intensity frontal precipitation, apparently gave rise to this fluvial regime (Knox et al. 1981; Brakenridge 1981). In the Loup River system, Nebraska, May (1986) cited thick late Holocene alluvial fills and frequent episodes of incision as evidence of highly variable flood magnitudes; Knox and others (1981) and Knox (1985) reported similar findings from drainage basins in southwestern Wisconsin. Such short-term change in fluvial regime does not appear to have perisited long enough to affect regional faunal and floral patterns, as it is imperceptible in the paleoecological record of the period (Fredlund and Jaumann 1987).

The change noted most frequently in the late Holocene paleoenvironmental record of the central Great Plains is the replacement of prairie vegetation by the mixed deciduous forest and prairie. At Muscotah Marsh in northeastern Kansas, this shift, which is indicated by an increase in _Quercus_ and _Carya_, occurred about 5000 yr B.P. (Gruger 1973); at Lake West Okoboji in northwestern Iowa, it took place about 3200 yr B.P. (Van Zant 1979). The remainder of the Holocene record to the time of anthropogenic vegetational changes reveals only minor fluctuations in the vegetation cover (see Gruger 1973; Van Zant 1979), none of which can be construed as a signal of regional climatic shifts.

In areas contiguous to the central Great Plains, however, there is evidence of significant climatic fluctuations during the late Holocene. Hall (1982), in an analysis of pollen and landsnails from rockshelters in northeastern Oklahoma, proposed that moist conditions existed between 2000 and 1000 yr B.P., followed by increasing aridity after 1000 yr B.P. At Carnegie Canyon in west-central Oklahoma, Hall and Lintz (1984) and Hall (1987) documented a high water table between 3200 and 1000 yr B.P. About 900 yr B.P. the water table declined, resulting in trenching of the canyon fill.

Although generally indiscernible in the paleoecological record, evidence of short-term climatic fluctuations does appear in the cultural record of the central Great Plains. For example, the record of the Upper Republican culture in south-central Nebraska contains evidence of mesic conditions between 1250 and 850 yr B.P., followed by increasingly xeric conditions around 700 yr B.P. (Wedel 1986). The xeric conditions apparently forced the Upper Republican culture to migrate southward (Bryson and Wendland 1967; Wedel 1986), where they remained until 450 yr B.P. when amelioration of the climate permitted a return to the central Great Plains (Wedel 1986). Records from the Mill Creek culture in northwestern Iowa also show evidence of environmental stress (apparently an acute drought) about 800 yr B.P. (Bryson and Baerreis 1968; Bryson et al. 1970). At Ash Hollow in western Nebraska, the dendrochronological record contains evidence of dry conditions at 730 yr B.P. and between 700 and 650 yr B.P. (Weakly 1962).

In summary, the late Holocene began with an increase in moisture availability between 5000 and 4000 yr B.P., signaling an end to the dry climatic conditions of the middle Holocene. Although floral and faunal records afford no conclusive evidence of regional climatic fluctuations in the central Great Plains after 4000 yr B.P., evidence of change is preserved in late Holocene sedimentation and archaeological records. Numerous cut and fill episodes indicate variable flood magnitudes, and archaeological evidence suggests that more xeric conditions, probably caused by an increase in zonal atmospheric flow, affected most of the central Great Plains from 800 to 700 yr B.P. Mesic conditions returned to the region about 450 yr B.P.

Summary and Conclusions

During the last 30,000 years, climatic conditions on the central Great Plains have changed from cool and mesic during the late Pleistocene to relatively warm and dry during the Holocene. The limited paleoecological record of the central Great Plains suggests that Peoria loess accumulated on a vegetated surface, possibly in a parkland vegetational community. Analysis of phytoliths extracted from loess deposits at the Eustis site in southwestern Nebraska indicates that loess was deposited under cool conditions with more effective moisture. In river valleys, a more continuous *Picea*/deciduous forest cover probably existed. On the eastern periphery of the region, where more data are available, it is generally accepted that a continuous *Picea* forest was present during the late Pleistocene. In response to increasing temperatures after 12,000 yr B.P., *Picea* moved rapidly northward and deciduous species became established in the central Great Plains; Picea disappeared from the region by 10,500 yr B.P. In addition to rapid warming, less equable climatic conditions appear to have developed in the region between 12,000 and 10,500 yr B.P. These conditions may have destroyed the complex biological communities that existed during the Pleistocene, resulting in either the extinction of mega-fauna or migration of those having a boreal affinity.

The limited record from the central Great Plains suggests that a mixed deciduous forest/prairie covered at least the eastern portion of the region from 10,500 to between 9000 and 8000 yr B.P. In response to warmer and drier conditions, the prairie moved eastward across the central Great Plains after 9000 yr B.P.; by 8000 yr B.P., it covered the central Great Plains, and by 7200 yr B.P. it extended to eastern Minnesota and western Wisconsin. Mesic conditions returned around 5000 yr B.P., resulting in establishment of the deciduous forest at its present position on the eastern periphery of the central Great Plains. The remainder of the Holocene paleoecological record displays only minor fluctuations in the vegetation cover, none of which provide a clear signal of climatic fluctuations.

Evidence of late Holocene climatic change does appear, however, in the archaeological and sedimentation records. The archaeological record implies that dry conditions existed in the central Great Plains between 800 and 700 yr B.P., followed by a return of more mesic conditions around 450 yr B.P. Numerous cut and fill episodes and thick alluvial fills in the sedimentation record attest to a fluvial regime dominated by highly variable flood magnitudes; such a regime contrasts markedly with the regimes of the middle and early Holocene, which appear to have been characterized by long-term aggradation, punctuated by relatively brief episodes of incision.

Finally, while it is generally accepted that dramatic and frequent climatic change has occurred in the central Great Plains over the last 30,000 years, there is less agreement as to whether such change occurred in an episodic or a gradual fashion. Based on the temporal distribution of several thousand radiocarbon determinations, Wendland and Bryson (1974) argued that climatic change occurred in an abrupt, step-wise fashion that could be linked to perturbations in atmospheric circulation. In contrast, following an analysis of the palynological record from various locations in Minnesota, Wright (1976) rejected the stepwise theory of climate change, concluding instead that change occurred gradually.

CHAPTER 4: STUDY SITE AND METHODOLOGY

Study Site

The Republican River basin has its headwaters in the dissected plains of eastern Colorado. Elevations in the portion of the Republican River basin under study here range from 1798 m above mean sea level (amsl) at the headwaters of the Arikaree River northwest of Limon, Colorado, to 593 m amsl at Harlan Lake. At its mouth on the western shore of Harlan Lake, the river drains 18,500 km^2 of eastern Colorado and southwestern Nebraska (see Figure 1:1).

In the vicinity of Harlan Lake, the Republican River has incised through Quaternary and Tertiary deposits to expose the underlying Niobrara formation and Pierre shale of Cretaceous age. The Niobrara formation is a lead gray-to-yellowish cream chalk, and the Pierre shale a carbonaceous, clayey yellowish-gray to black shale that is interbedded with gypsum (Bradley and Johnson 1957). The Pierre shale serves as an important aquiclude in the central Great Plains. The Ogallala formation of Tertiary age, which consists of unconsolidated, cross-bedded sand and gravel, overlies the Cretaceous bedrock, and is exposed on upland flanks in the study area.

Pre-Quaternary fluvial dissection cut drainage ways in the Pierre shale and Niobrara formation (Lugn 1935). These drainage ways were subsequently filled with early Pleistocene fluvial deposits that comprise the Holdrege, Grand Island, Sappa, and Crete formations (Bradley and Johnson 1957). Of these, the most widespread in the study area is the Crete formation, a Pleistocene deposit of fluvial origin that is composed of cross-bedded, clean sands and gravels. Erosion just prior to or concurrent with late Pleistocene loess deposition removed much of the Crete formation from the south shore of the Republican River valley, although extensive exposures remain on the north side of the river (Zakrzewska 1963). This asymmetrical pattern of erosion is believed to have resulted from Quaternary faulting of the Ogallala and overlying early Pleistocene formations (Zakrzewska 1963).

Both the Loveland and Peoria loess have been recognized in the study area, although exposures of the latter are far more prevalent. Following deposition of the Loveland loess, widespread landform dissection occurred. The Peoria loess was subsequently deposited, obscuring most of the small valleys formed by the earlier erosion (Lugn 1935).

With the exception of reworked loess that appears as colluvium on valley sides and Holocene fluvial deposits that are found in valleys, the Peoria loess constitutes most of the surface material in the study area (see Plate 1:1). Generally, upland soils are developed in Peoria loess, and lowland soils in silty and loamy alluvium. Upland soils, which are silty in texture with a small component (ca. 10%) of fine sand, are classified as Typic Argiustolls or Typic Haplustolls. Lowland soils close to the river are classified as Typic Ustipsamments and Typic Ustifluvents, whereas those developed on distal floodplains and terraces are clasified as Cumulic Haplustolls (Mitchell et al. 1974).

The Harlan County area receives an average of 566 mm of precipitation each year, although the annual total varies significantly (Meyers 1974). Frontal activity in the spring and convectional activity in the early summer account for the bulk of the precipitation (Figure 4:1). The vegetation cover of the study area consists of a mixed prairie-short grass cover (Figure 4:2). Short grass types are generally restricted to steep slopes and

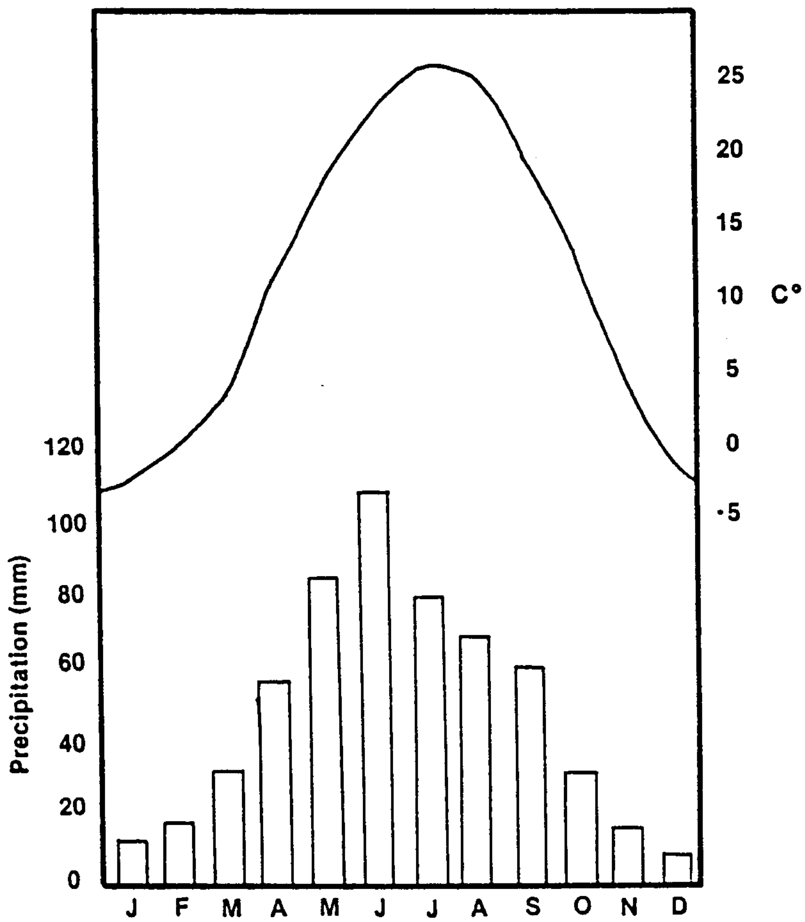

Figure 4:1 Representative climograph for study area. Data are for Alma, Nebraska, for the period 1937 to 1966. Note the summer maximum of precipitation and the winter minimum. The average annual precipitation total during the period 1937 to 1966 was 566 mm. Data are from Mitchell et al. (1974, 62).

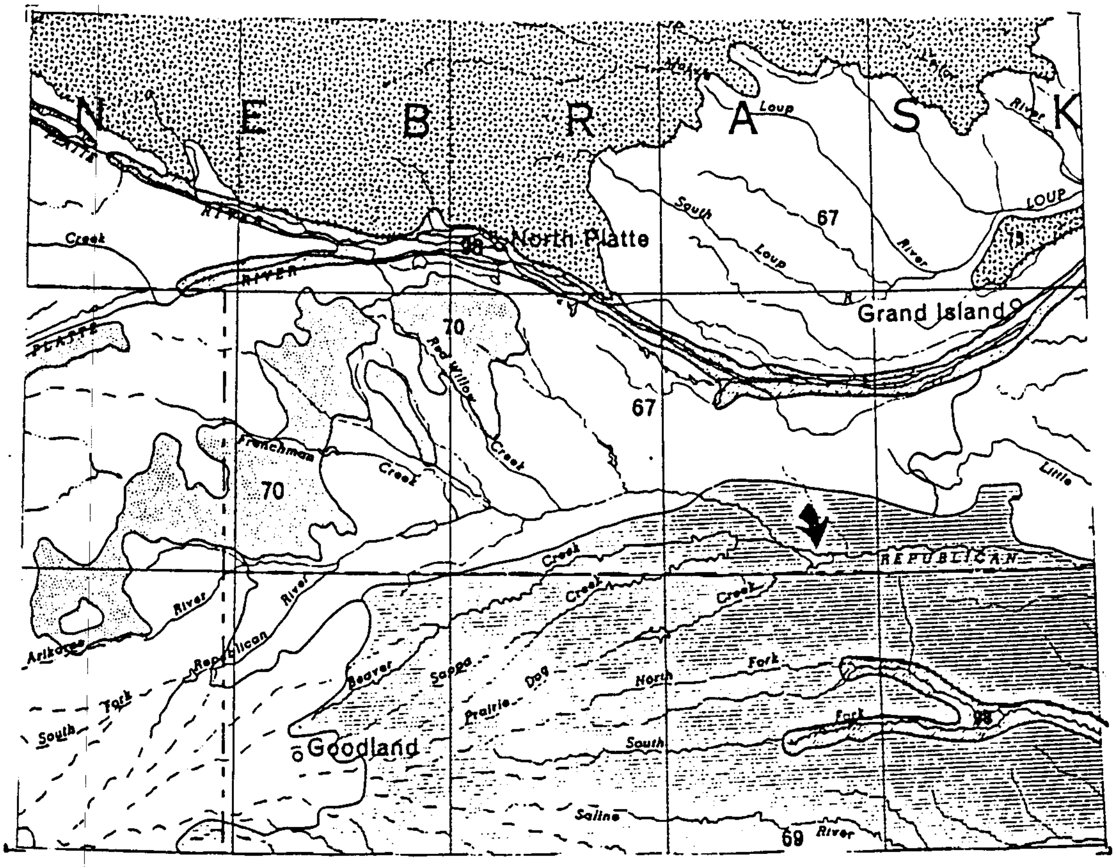

Figure 4:2 Portion of Kuchler's (1964) map of potential natural vegetation of the United States. Harlan Lake indicated with an arrow. Study site lies within vegetation community #69, the bluestem-grama prairie (<u>Andropogon-Bouteloua</u>). Map scale is 1:13,168,000. Top of the map is north. Vegatational communities depicted are:

- #67: wheatgrasss-bluestem-needlegrass (<u>Agropyron-Bouteloua-Buchloe</u>)
- #69: bluestem-grama prairie (<u>Andropogon-Bouteloua</u>)
- #70: sandsage-bluestem prairie (<u>Artemesia-Andropogon</u>)
- #75: Nebraskan Sandhills prairie (<u>Andropogon-Calamovilfa</u>)
- #98: northern floodplain forest (<u>Populus-Salix-Ulmus</u>)

dry uplands, whereas mid- and tall grasses are found on gentler slopes and in valleys (Weaver and Bruner 1948). The dominant short grasses are _Bouteloua gracilis_ (blue grama), _Buchloe dactyloides_ (buffalo grass), _Bromus commutatus_ (hairy chess), and _Andropogon scoparius_ (little bluestem). Common mid- and tall grasses include _Andropogon gerardii_ (big bluestem), _Bouteloua curtipendula_ (side-oats grama), _Angropyron smithii_ (western wheat grass), and _Panicum virgatum_ (switchgrass) (Weaver and Bruner 1948; Weaver 1954).

The extensive grasslands of the central Great Plains occupy a region of North America that receives little winter precipitation. It is a region that is subjected frequently to drought, especially when spring and summer rains fail. In contrast to the eastern half of the United States, where the midlatitude forest is the dominant vegetation type, areas supporting grassland have few days of precipitation, little cloud cover, and low relative humidities. Drought conditions occur when meridional atmospheric flow from the south is replaced by zonal flow from the west. The extent and severity of drought depends on the strength of such flow (Borchert 1950; Bryson and Wendland 1967; Bryson and Baerreis 1968; Kalnicky 1974; Wendland 1978).

The drought of the 1930s may serve as a modern analogue to the dry conditions that affected the Great Plains from approximately 7000 to 5000 yr B.P. Research conducted during and following this drought revealed significant changes in grassland cover as a consequence of the dry conditions. For example, at Hays, Kansas, the grassland cover was reduced from 85% in 1932 to 20% in 1940. After 1940, with the return of more abundant precipitation, the cover increased, although the speed of recovery varied by species and according to topographic position (Weaver and Bruner 1945; Tomanek and Hulett 1970). For example, hillslopes in parts of the central Great Plains that had been rendered barren by the drought of the 1930s supported only 50% of their pre-drought grass cover by the late 1940s (Weaver and Bruner 1948); in contrast, the pre-drought vegetation cover in many areas of Nebraska was entirely restored by the late 1940s (Weaver and Bruner 1945). After an analysis of the historical period (i.e., post-European settlement), Kuchler (1972) concluded that prairie boundaries in the central Great Plains are responsive to variations in the reliability of precipitation.

In detailed studies of the prairie in western Iowa, eastern Nebraska, and north-central Kansas, Weaver and Albertson (1943) noted that _Andropogon scoparius_ suffered the most serious losses from the drought of the 1930s. In contrast, _Andropogon gerardii_, probably because of its deeper root system, successfully weathered the dry conditions. With the loss of much of the _Andropogon scoparius_ cover, species such as _Bouteloua curtipendula_ and _Agropyron smithii_ invaded the central Great Plains (Weaver and Albertson 1943); many continue to prosper there today.

Prior to the arrival of white settlers in 1871, the Republican River valley supported numerous Native American cultures who valued the region for its plentiful game and good water supply. Perhaps the most notable of these was the Upper Republican complex, a culture that inhabited the region between 900 and 600 yr B.P. (Wedel 1986). These original inhabitants yielded eventually to the European settlers who founded Harlan County in June 1871 and later selected Alma as the county seat. That same year, Republican City was established, followed by Orleans the year after (Figure 4:3). In 1879 the railroad arrived in Harlan County, precipitating an economic boom and an increase in population (Andreas 1882). Harlan County, because of

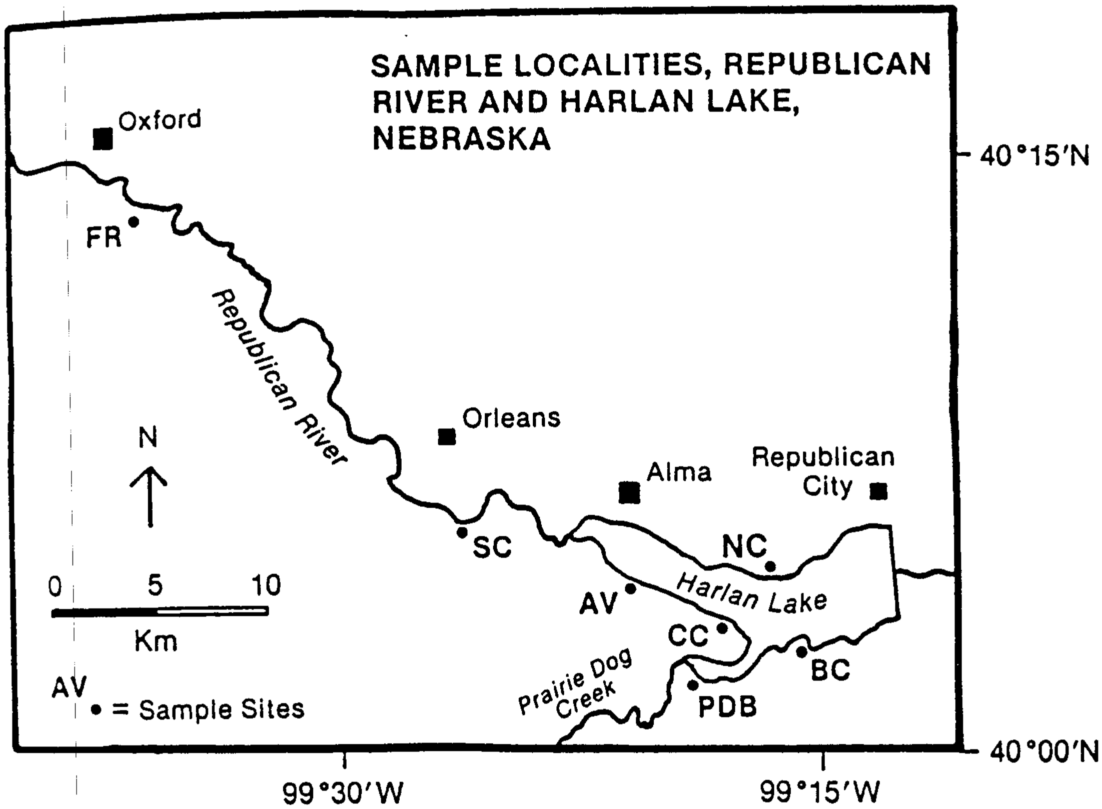

Figure 4:3 Sample localities, Republican River and Harlan Lake, Nebraska. Map scale is 1:250,000. Top of map is north. Sample site abbreviations are as follows:

AV = Alma Vista
BC = Bone Cove
CC = Coyote Canyon
FR = Freer
NC = North Cove
PDB = Prairie Dog Bay
SC = Schoenenburg

its link to agriculture, has suffered through several boom and bust cycles. It has also been affected by the 20th century consolidation of agricultural activity, and the resultant decline in the number of family farms. This decline is reflected in the population trends of the county: after reaching a peak of 9578 in 1910, the population has declined steadily to the present figure of 4292 (Mitchell et al. 1974; Bureau of the Census 1980,29-16).

Harlan Dam was planned by the United States Army Corps of Engineers in the late 1930s, primarily in response to a deadly flood in June 1935 that struck along the Republican River from the Colorado/Nebraska border to the Nebraska/Kansas border. Eyewitness accounts of this flood reported that a wall of water 3 to 16 feet high and 1 to 2 miles wide descended on Orleans at 3:00 A.M. on June 1. The floodwaters then continued downstream, striking Alma at 5:00 A.M. and Republican City just after daybreak. In all, 112 people perished across southern Nebraska, one-fourth of them in the Oxford-Orleans-Alma reach of the river (Hurlbert 1980; Pfeifer 1985; Harlan County Journal 1985). Spurred by the devastation, construction of the Harlan Project began in the late 1940s. The gates that hold

Harlan Lake were closed in 1952. Ostensibly built to afford flood control to the valley, the lake serves increasingly as a recreational facility and a reliable supply of irrigation water to farmers downriver in Nebraska and Kansas.

Methodology

Reconstruction of the late Quaternary sedimentation history of the study area required extensive sampling of and drilling in upland, valley side, and lowland deposits. Because of the outstanding exposures present on Harlan Lake, sampling focused on the lake shoreline and a reach of the Republican River from Republican City to Oxford, Nebraska (see Figure 4:3). With the aid of 7.5 minute, 1:24,000 United States Geological Survey topographic maps, I first undertook a walking survey of the river to identify accessible river and wave-cut exposures that appeared to feature eolian and fluvial deposits. Because of their potential for radiocarbon dating, those exposures that contained paleosols were assigned a higher priority for description and sampling. On the basis of this survey, the following wave-cut exposures on Harlan Lake were selected for detailed description and sampling: Alma Vista, North Cove, Bone

Cove, and Prairie Dog Bay. Unfortunately, only one non-vegetated cutbank exposure, the Schoenenburg site, was found on the reach of the river under study here (see Figure 4:3). In order to augment investigation of the stratigraphic relationships between upland and lowland deposits, I used a trailer-mounted drill rig to collect cores along two transects plotted across lowland alluvial and upland loess deposits (Plate 4:1). Cores were packaged and transported to the laboratory for description, measurement, sampling, and analysis.

As a compromise between a time-consuming, random sampling scheme and a potentially biased systematic scheme, I sampled randomly within each depositional unit of exposures and cores. At the same time, I was careful that at least one sample, and usually two, were obtained from each unit. Where stratigraphy was more complex (e.g., fluvial deposits), additional samples were collected. To prevent contamination from above, exposures were cut back at least 0.20 m prior to sampling. Depth to units was determined from exposure banktops. Moist Munsell colors of all samples were also determined in the field under natural light. For a detailed description of cores and exposures and the

Plate 4:1 Trailer-mounted drill rig (Giddings) used to collect cores from valley and valley side fills.

samples collected from them, see Appendix A.

At several locations, 2 kg bulk samples of paleosols were collected for radiocarbon dating of humates. To guard against contamination, faces were cut back at least 0.50 m prior to sampling (Plate 4:2). Then, the sediment overhanging the top of the Ab was removed to form a shelf from which samples were collected (Plate 4:3). Although I attempted to sample within 0.05 m of the top of each Ab, bioturbation of some paleosols made it necessary to sample at a lower level of the soil. Rootlets were handpicked from bulk paleosol samples in the laboratory, and then the samples were submitted to the University of Texas Radiocarbon Laboratory for radiocarbon dating of soil humates (base-soluble fraction). Where noted, ages were corrected for the stable carbon isotope (C13) ratio (see Stuiver and Polach 1977; Taylor 1987).

Because paleosols represent episodes of soil formation, an age from the upper portion of an unstripped paleosol provides a maximum date for the burial of that soil. Conversely, an age from the lower portion of a soil Ab horizon furnishes a minimum date for the beginning of soil formation (Haas et al. 1986). By affording stratigraphy a chronometric framework,

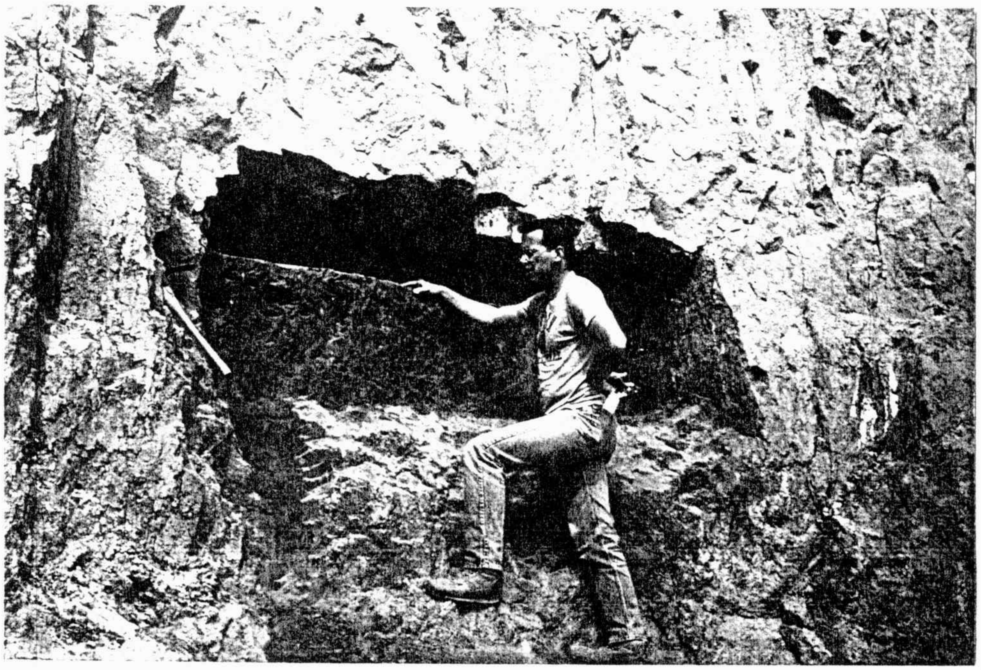

Plate 4:2 Paleosol prepared for bulk radiocarbon sample. Exposures were cut back at least 0.50 m prior to sampling to guard against contamination by humates from above. Photo is the Bone Cove site.

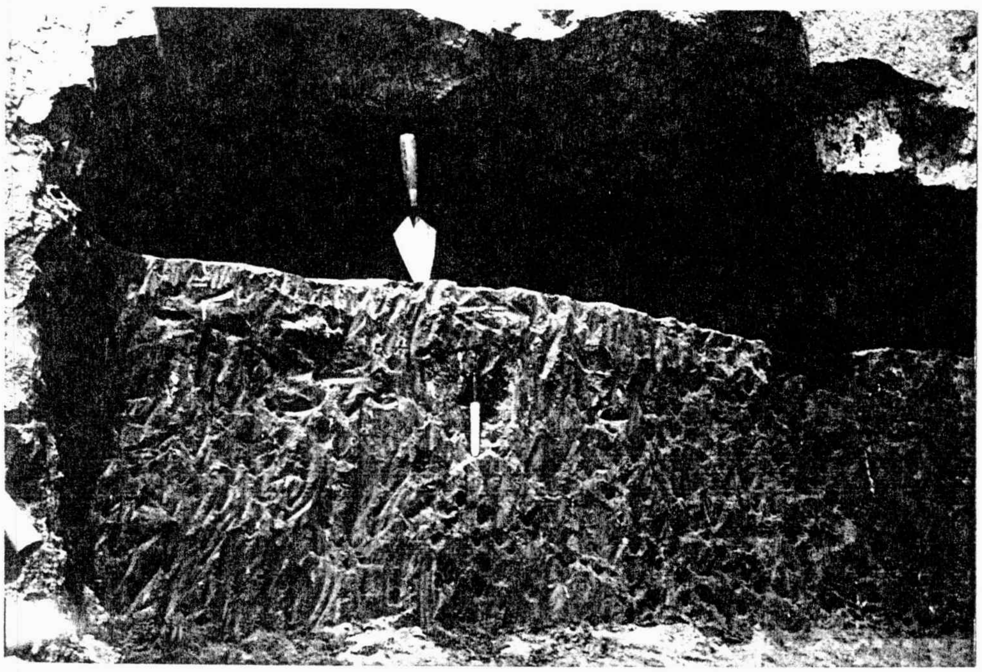

Plate 4:3 Close-up of paleosol in Plate 4:2 prepared for bulk radiocarbon sample. Sample was collected from shelf in which trowel is sticking. Pen length is 13.5 cm. Photo is the Bone Cove site.

radiocarbon ages facilitated the construction of a chronology of sedimentation and erosional activity. A list of all radiocarbon ages obtained in this study and a discussion of the process of radiocarbon dating of soils are contained in Appendix B.

In the laboratory, several basic soil analyses were run to determine the physical and chemical characteristics of deposits. I used the pipette method to determine the particle size distribution of samples (Day 1965; Gee and Bauder 1985). This method has been shown to produce results that are consistent with those obtained using the Coulter counter/multisizer (Pennington and Lewis 1979). Particle size data were converted to the logarithmic phi scale using a program developed by M.C. Prante of the Department of Geography, University of Kansas. This program constructs a cumulative frequency curve for the particle distribution, and then reads the desired phi value from it. Using this technique, the fifth, fiftieth, and eighty-fourth percentile phi values were determined. The fifth percentile value represents the coarse tail of the sample, the fiftieth percentile value the median value of the sample, and the eighty-fourth percentile value the fine tail of the sample (see Knox 1985). The mean phi value and degree of sorting were also calculated using equations devised by Folk and Ward (1957). For a detailed summary of calculations and statistical tests, see Appendix C.

The calculated phi values were plotted against depth to ascertain trends down a profile. Comparison of these trends, in turn, facilitated differentiation of eolian, fluvial, and reworked deposits, and, for fluvial deposits, permitted assessment of the relative strengths of fluvial sedimentation events. Calcium carbonate contents were measured using the Chittick method of digestion with hydrochloric acid (Dreimanis 1962). Organic carbon content was determined by the dichromate method (Allison 1965; Nelson and Sommers 1982). The organic carbon content of paleosols developed in late Holocene deposits was assumed to be proportional to the duration of surface stability and soil development (see Birkeland 1984; Holliday 1988). That is, the organic carbon content was used as an indicator of the relative duration of floodplain stability. This assumption is valid only when the paleosols being compared have approximately the same ages; if their ages differ substantially, then the loss of organic matter through postburial alteration will likely bias results (see Stevenson 1969; Gerasimov 1971; Hallberg et al. 1978). Finally, I used clay mineralogy, as determined by x-ray diffraction, to help differentiate among deposits. The clay fraction (< 2 microns) was slurry-mounted on ceramic tiles and x-rayed between 2° and 37° 2θ using Cu K-alpha radiation with Ni-filtration. To assist in the recognition of clays, both air-dried and oven-dried (600° C for 1 hour) samples were run. I identified diffraction peaks by comparison to x-ray diffraction traces provided in Carroll (1970).

The range of laboratory analyses discussed above permitted determination of the physical character of the various sediments in the study area. The resultant data were then used to differentiate among fluvial, eolian, and reworked deposits, and, for fluvial deposits, to gauge the relative strengths of fluvial sedimentation events. Radiocarbon ages on paleosols afforded chronometric control for the chronology, and, in the case of late Holocene fluvial deposits, provided an indicator of the relative duration of floodplain stability. Radiocarbon determinations also facilitated correlations with chronologies of landform evolution that have been produced elsewhere in the Great Plains.

CHAPTER 5: RESULTS AND DISCUSSION

As a result of descriptions completed and sampling undertaken at the six study localities, I have identified loess, colluvial, and alluvial deposits; at least one type of deposit is present at each site, and two sites feature all three deposits (see Figure 4:3 for site locations). A more precise location of study sites is provided by annotated topographic maps included in the description of each site. As noted earlier, detailed descriptions of exposures and cores are contained in Appendix A. The following abbreviations are used in descriptions of the sample locations:

BC = Bone Cove
PDB = Prairie Dog Bay
NC = North Cove
AV = Alma Vista

Late Pleistocene Study Sites

Bone Cove

Bone Cove is located on the south shore of Harlan Lake, approximately nine km southeast of Alma, Nebraska (Figure 5:1). Sampling was completed on the east-facing exposure of the cove (Plates 5:1 and 5:2). At the base of the exposure, 1.5 m of cross-bedded, coarse and medium fluvial sands are overlain by yellowish-brown

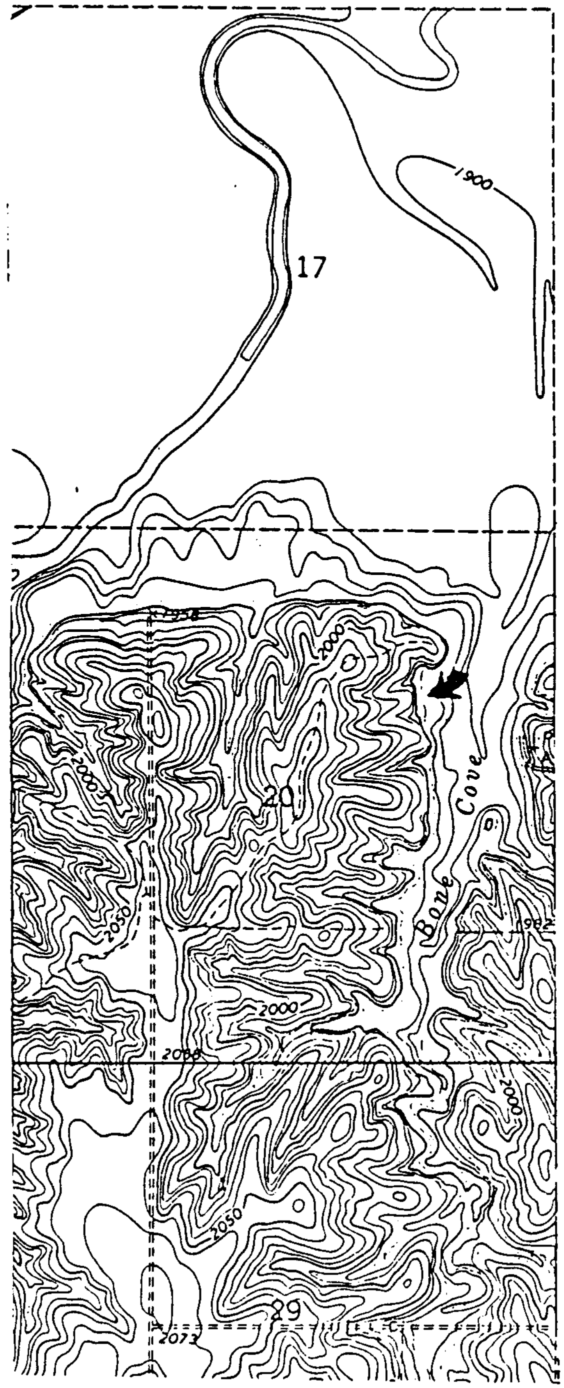

Figure 5:1 Section of topographic map showing location of Bone Cove sample site (arrow). Site located in: SW 1/4, NE 1/4, Sec. 20, T1N, R17W, Alma, NE/KS 7.5 minute quadrangle. Top of the map is north, contour interval is 10 feet.

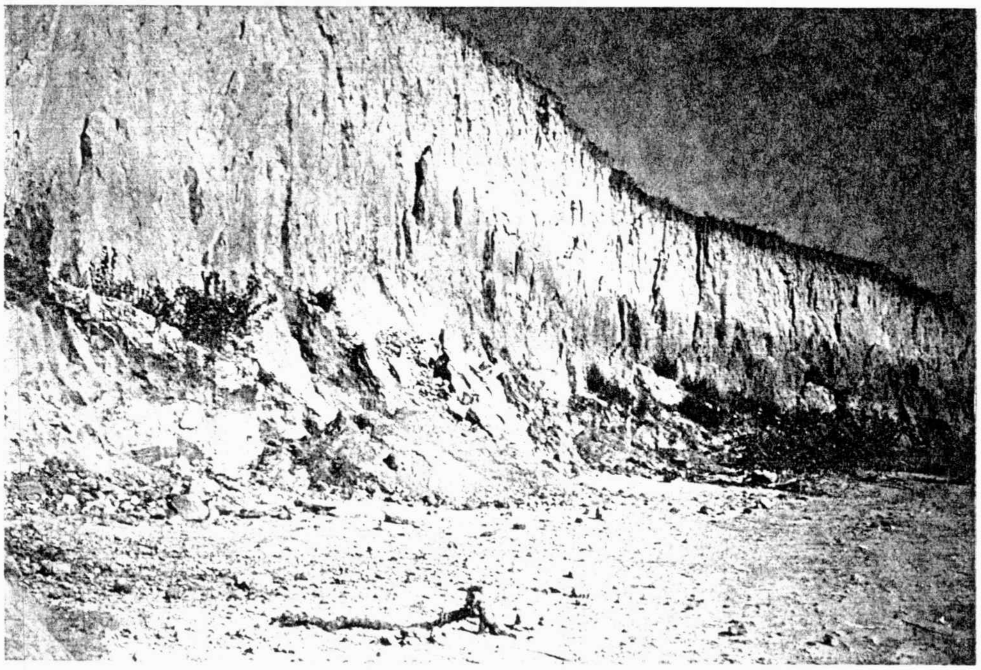

Plate 5:1 Bone Cove exposure. Exposure comprises a basal unit of the temporal equivalent of the Gilman Canyon Formation and overlying Peoria loess. Paleosol, which divides the two units, was radiocarbon dated at 26,260 ± 680 yr B.P. (Tx-5910). Note diffuse upper boundary of paleosol, an indication that it is a cumulic soil.

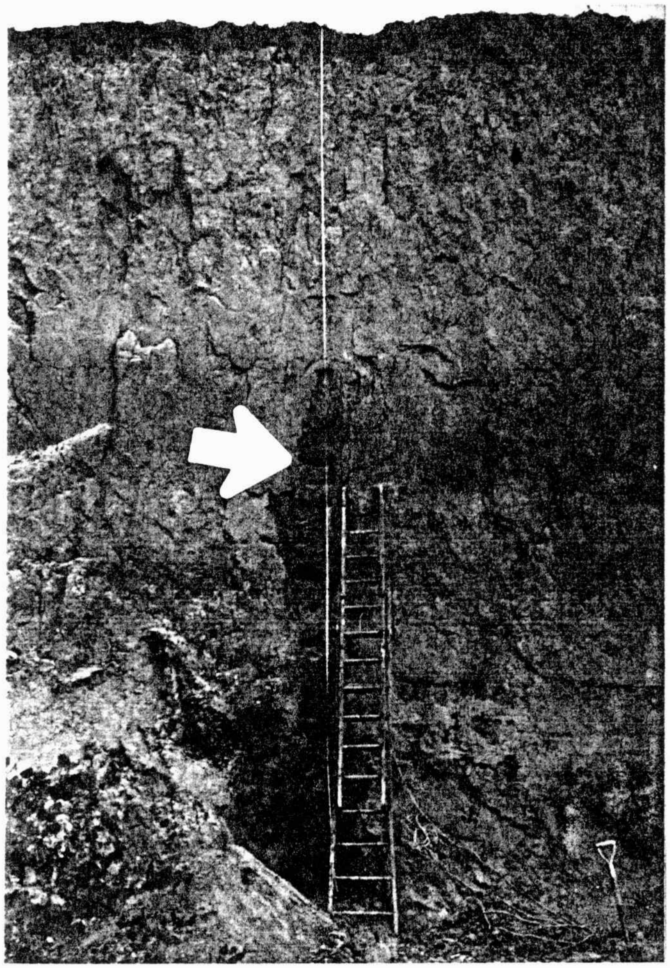

Plate 5:2 Sample section on Bone Cove exposure. From bottom to top, exposure comprises the temporal equivalent of the Gilman Canyon Formation, a paleosol radiocarbon dated at 26,260 ± 680 yr B.P. (Tx-5910), and Peoria loess. Radiocarbon sample indicated with arrow. Note the diffuse upper boundary of paleosol, an indication that it is a cumulic soil.

(10YR 6/4), massive fine sand (Figures 5:2 and 5:3). A paleosol featuring dark gray (10YR 3/1) Ab and Bw horizons is developed in this fine sand; the diffuse upper boundary of the Ab horizon indicates that pedogenesis was able to keep pace for a short time with deposition of the overlying loess (see Plate 5:2), thereby forming a cumulic soil (see Riecken and Poetsch 1960). Soil humates from the upper 0.17 m (5.20 - 5.27 m) of the paleosol were radiocarbon dated (uncorrected for C13 ratio) at 26,260 ± 680 yr B.P. (Tx-5910), an age that correlates with dates of 35,000 to 21,290 yr B.P. on the upper portion of the Gilman Canyon Formation as reported elsewhere in Nebraska (Dreeszen 1970; May and Souders 1988; May 1989a; D.W. May, Pers. Comm. 1989).

The paleosol is overlain by up to 5 m of slightly calcareous, massive buff (10YR 5/3 to 10YR 4/4) silt that was deposited sometime after 26,260 yr B.P. (see Figures 5:2 and 5:3). Aside from a layer of slightly coarser silt between 4 and 5 m that may correspond to deposition by higher wind velocities, the texture is homogeneous to the top of the exposure (see Figure 5:3). Landsnails are scattered throughout the deposit; a high concentration between 2 and 3 m may denote deposition under conditions of increased moisture. The exposure is

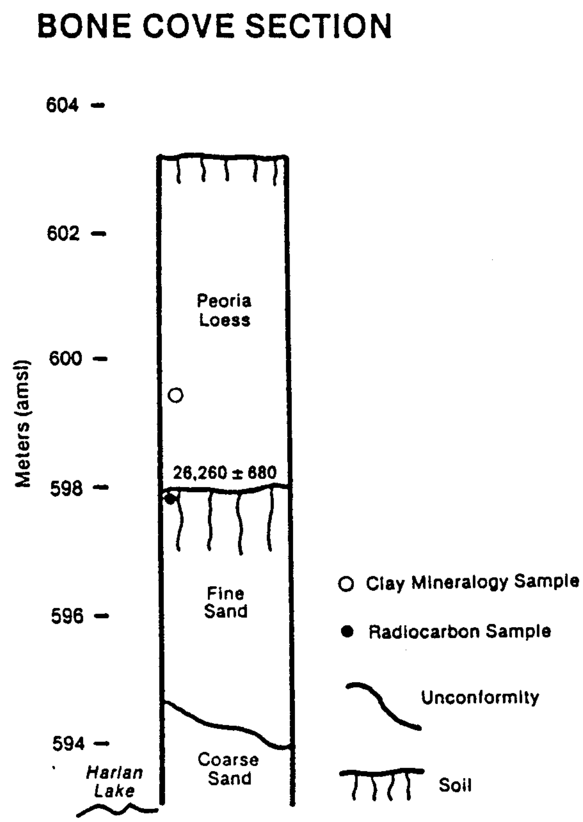

Figure 5:2 Diagrammatic section of the Bone Cove exposure. Depth is measured from the banktop. Banktop height is 603 m above mean sea level. Locations of clay mineral and radiocarbon samples are indicated. Note the unconformity between basal sands and the overlying fine sands. Fine sand unit is thought to be the temporal equivalent of the Gilman Canyon Formation.

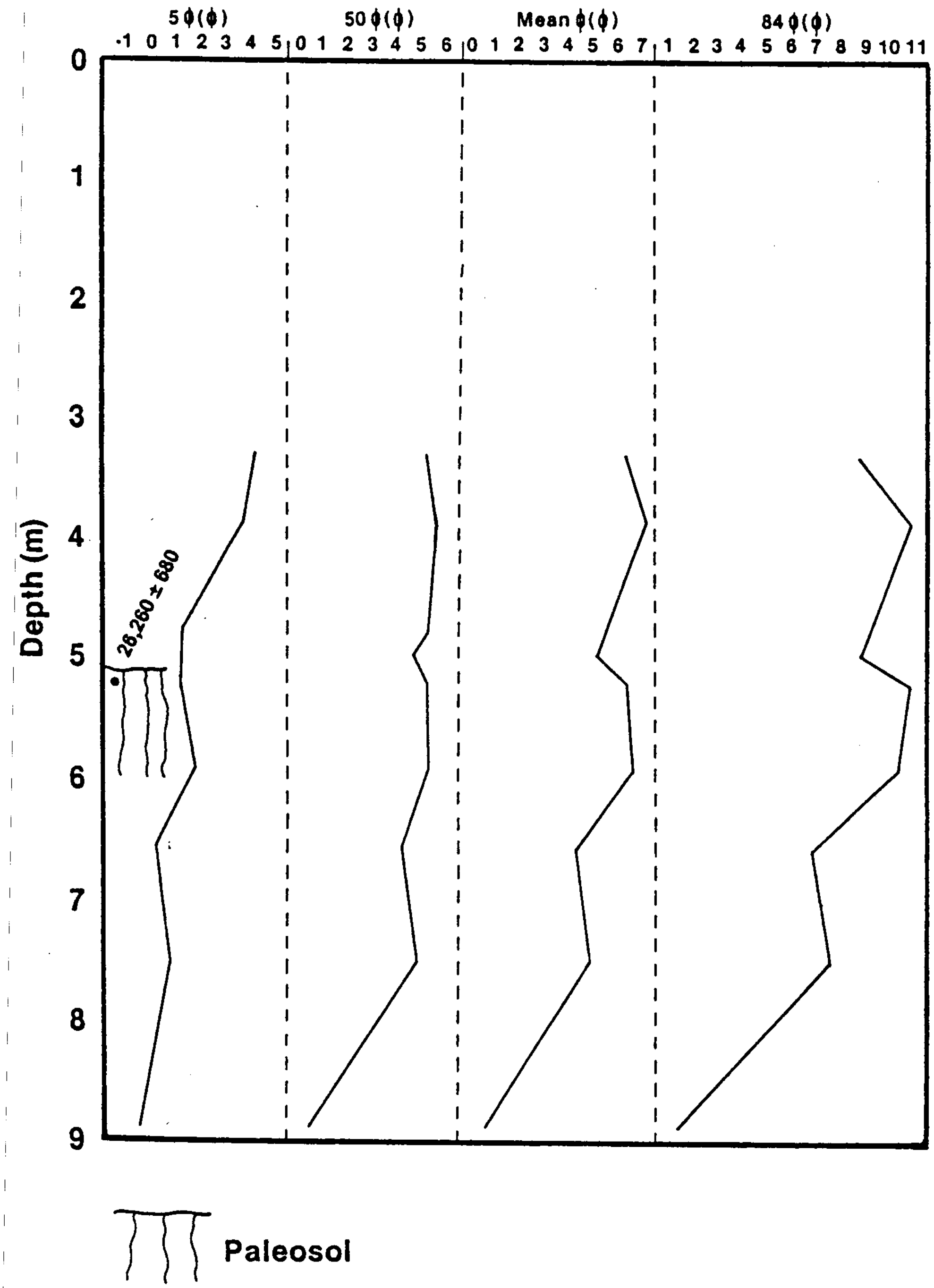

Figure 5:3 Particle size trends for the Bone Cove sample section. Depth is measured from the banktop. Banktop height is 603 m above mean sea level.

calcareous throughout, with highest values present in the upper portion of the paleosol (Figure 5:4). An x-ray diffraction sample collected between 3.30 and 3.35 m (see Figure 5:2) revealed montmorillonite, illite, quartz, and kaolinite clays (Figure 5:5), a mineral assemblage that is typical of the Peoria loess (Swineford and Frye 1951; Beavers 1957). In age (deposition post-26,260 yr B.P.), appearance, texture, and clay mineralogy, this unit is identical to the Peoria loess that mantles much of the central Great Plains.

In the classical loess stratigraphy of the Great Plains, as reported by C.B. Schultz and associates of the University of Nebraska-Lincoln, the Brady paleosol is developed in the top of the Peoria loess. At the Bone Cove site, however, the Brady is absent, at least as a buried soil; indeed, there is no evidence in the Peoria loess that deposition ceased, or even slowed (e.g., organic enrichment), and there does not appear to be any unconformity in the deposit. It is possible that the Brady paleosol was eroded from the Bone Cove exposure sometime during the Holocene, although one would expect to see evidence of such an event preserved in the stratigraphy. It is also possible that the

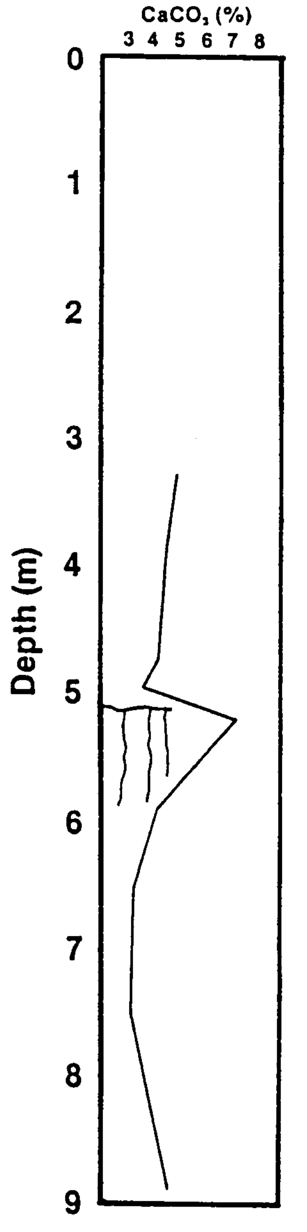

Figure 5:4 Calcium carbonate trend for the Bone Cove sample section. Note the accumulation of calcium carbonate in the paleosol. Depth is measured from the banktop. Banktop height is 603 m above mean sea level.

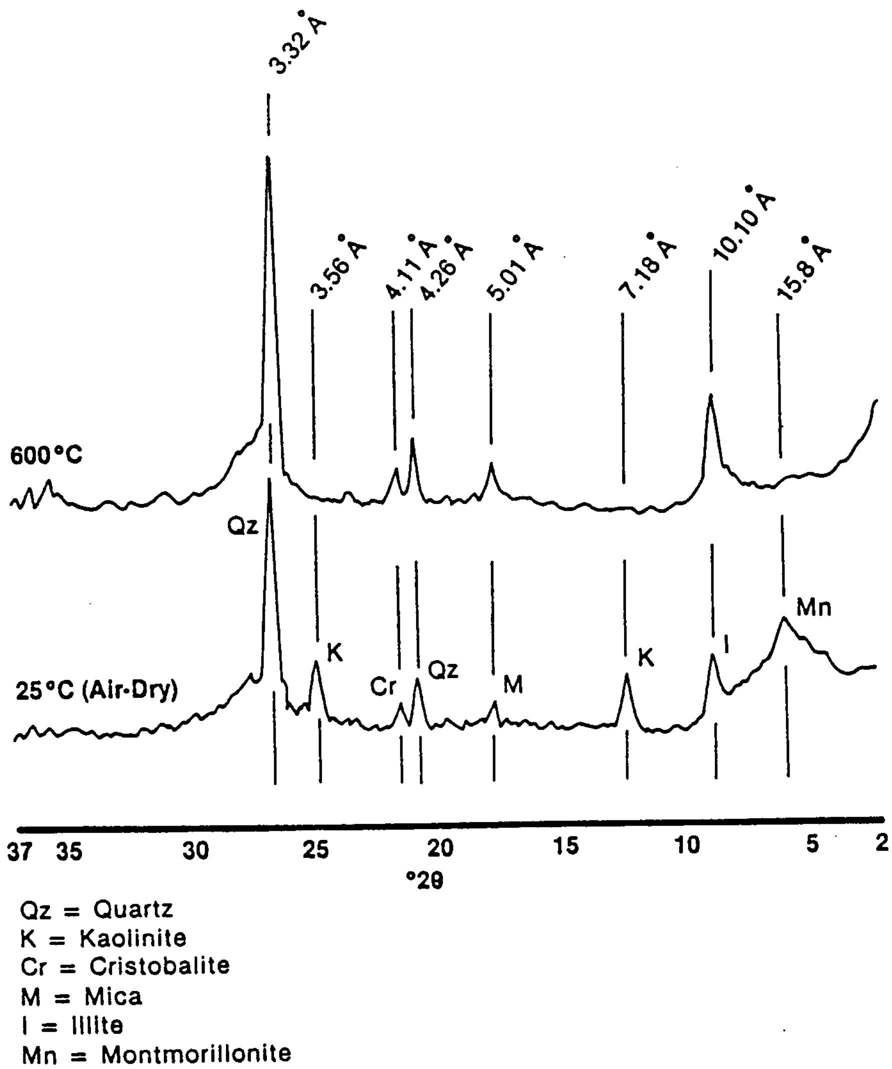

Figure 5:5 X-ray diffraction traces for clays (<2 microns) from sample collected at Bone Cove section, Peoria loess (depth of 3.30 - 3.35 m). See Figure 5:2 for sample location. Material sampled was deposited sometime after 26,000 yr B.P. X-ray traces are for air-dried (25° C) and oven-dried (600° C) samples. Mineral assemblage is typical of the Peoria loess, as reported by Swineford and Frye (1951)

surface soil at the Bone Cove exposure is the stratigraphic and temporal equivalent of the Brady paleosol. A radiocarbon age on humates from the base of the surface soil would answer this question.

In summary, the Bone Cove site features what appears to be part of the Gilman Canyon Formation overlain by up to 5 m of massive Peoria loess. The cumulic paleosol at the base of the loess, which marks the start of Peoria loess deposition, was radiocarbon dated at 26,260 yr B.P. Although the Brady soil does not appear as a buried soil at the site, the surface soil may be the stratigraphic and temporal equivalent of it.

Prairie Dog Bay

The Prairie Dog Bay site, which is located at the Prairie Dog Bay inlet on the south shore of Harlan Lake, comprises four exposures: PDB1, PDB4, PDB3, and PDB5 (Figures 5:6 and 5:7, Plate 5:3). The extensive (3 km^2), nearly level Prairie Dog Bay surface extends from the shore of Harlan Lake to the uplands (Plate 5:4). The PDB1 exposure is composed of buff (10YR 5/3), calcareous, massive coarse silt that displays little variation in texture from the base of the exposure to

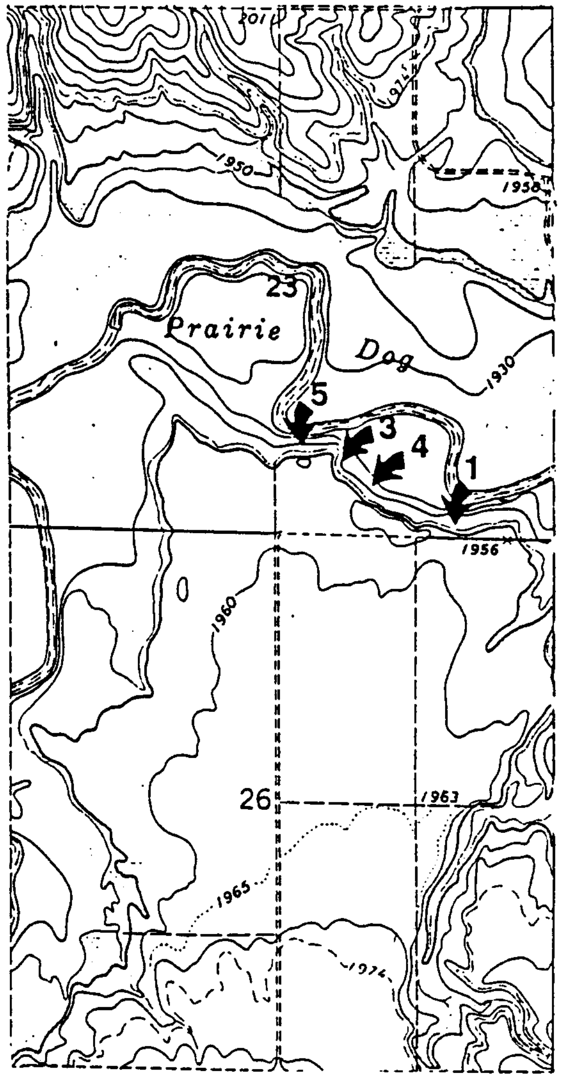

Figure 5:6 Section of topographic map showing location of Prairie Dog Bay sample sites. Numbers correspond to the individual sample sites: 1 = PDB1, 3 = PDB3, 4 = PDB4, 5 = PDB5. Sites located as follows: PDB1 - SE 1/4, SE 1/4, Sec. 23, T1N, R18W; PDB4 - SW 1/4, SE 1/4, Sec. 23, T1N, R18W; PDB3 - SW 1/4, SE 1/4, Sec. 23, T1N, R18W; and PDB5 - SE 1/4, SW 1/4, Sec. 23, T1N, R18W. Alma, NE/KS 7.5 minute quadrangle. Top of map is north, contour interval is 10 feet. The pre-dam channel of Prairie Dog Creek is visible in the middle of Sec. 23. The extensive and flat Prairie Dog Bay surface stretches about 2 km to the south of the sample locality.

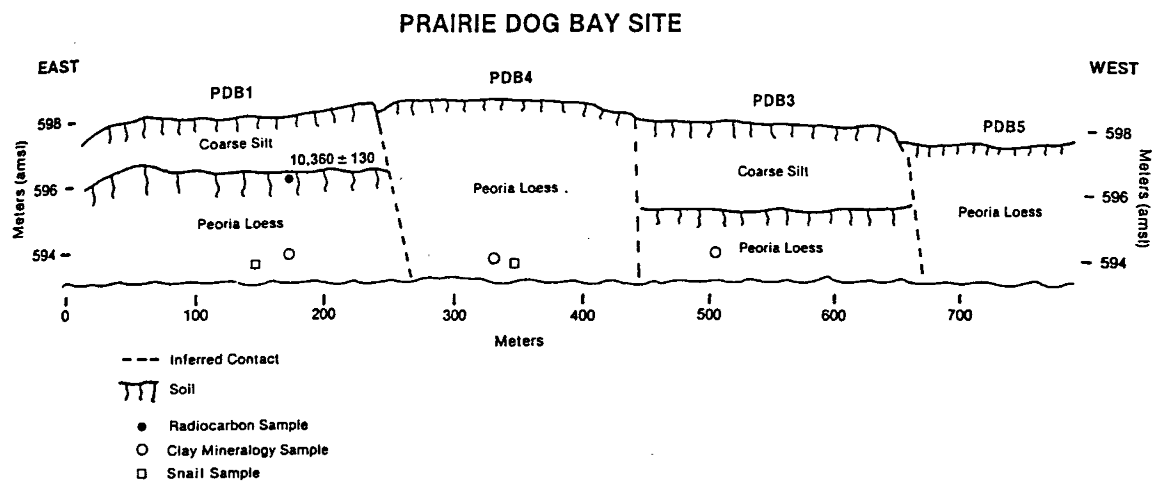

Figure 5:7 Idealized profile of the Prairie Dog Bay site and sample localities. View is from Prairie Dog Bay facing south, with east on the left and west on the right of the profile. Surface height is given in meters above mean sea level (amsl). Note locations of snail, radiocarbon, and clay mineral samples. See Figure 5:15 for proposed late Quaternary evolution of the site.

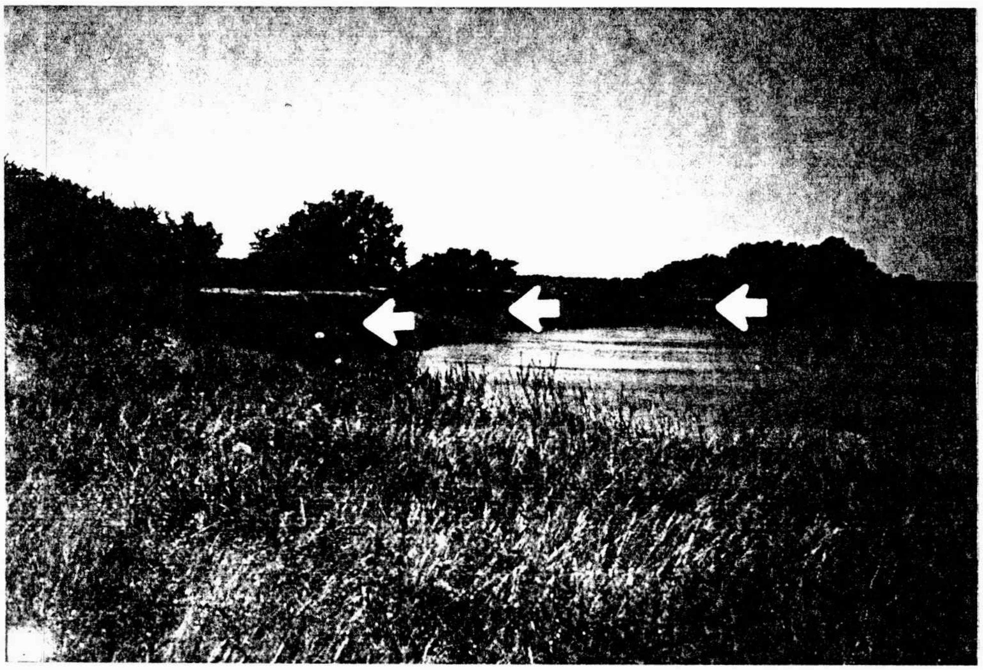

Plate 5:3 Prairie Dog Bay site. Arrows mark, from left to right, PDB1, PDB4 and PDB3 sample localities. Note that PDB1 and PDB3 feature paleosols, but PDB4 does not. Paleosol in PDB1 (foreground) was radiocarbon dated at 10,360 ± 130 yr B.P. (Tx-5909). Deposits are Peoria loess (ca. 25,000 to 11,500 years old) and reworked Peoria loess (early Holocene in age).

Plate 5:4 Prairie Dog Bay surface. Photo taken from surface looking towards Harlan Lake and the edge of the PDB4 exposure. Surface extends approximately 2 km back from the lake.

the banktop; this implies that deposition was uninterrupted (Figure 5:8). Humates extracted from the upper 0.20 m (2.10 - 2.28 m) of the paleosol that is prominent in the middle of the exposure produced a radiocarbon age (uncorrected for C13 ratio) of 10,360 ± 130 yr B.P. (Tx-5909). A small (ca. 1 kg) bulk sample of sediment from the base of the exposure yielded numerous landsnails, including *Vallonia gracilicosta*, *Pupilla blandi*, *Pupilla muscorum*, and *Gastrocopta armifera*; these species were present in upland settings of the central Great Plains between 13,000 and 10,500 yr B.P. (J.D. Stewart, Pers. Comm. 1986). With the exception of *Gastrocopta*, which is still present in topographically sheltered settings of the central Great Plains, all of these landsnails are extant today in litter-covered forest areas of the Rocky Mountains and along the United States/Canadian border, that is, in climates having more effective moisture and a lower annual temperature than the climate that exists today in the central Great Plains (Leonard 1952; Wells and Stewart 1987a). The x-ray diffraction traces on a sample collected just below the paleosol (see Figure 5:7) reveal the dominance of montmorillonite, illite, quartz, and kaolinite (Figure 5:9). Based on the

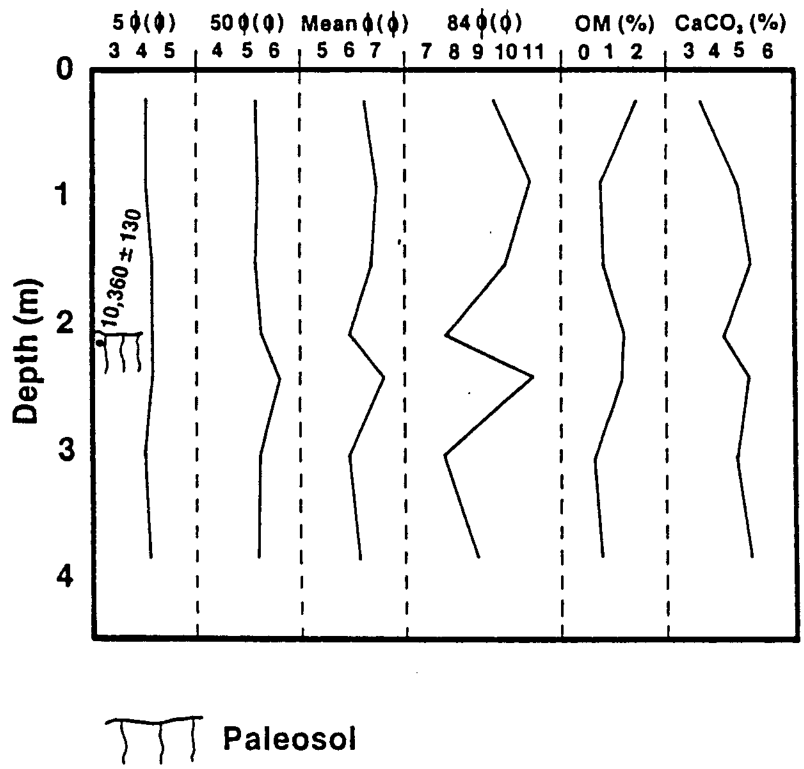

Figure 5:8 Particle size, organic matter, and calcium carbonate trends for PDB1 sample site. Depth is measured from banktop. Banktop height is 598 m above mean sea level.

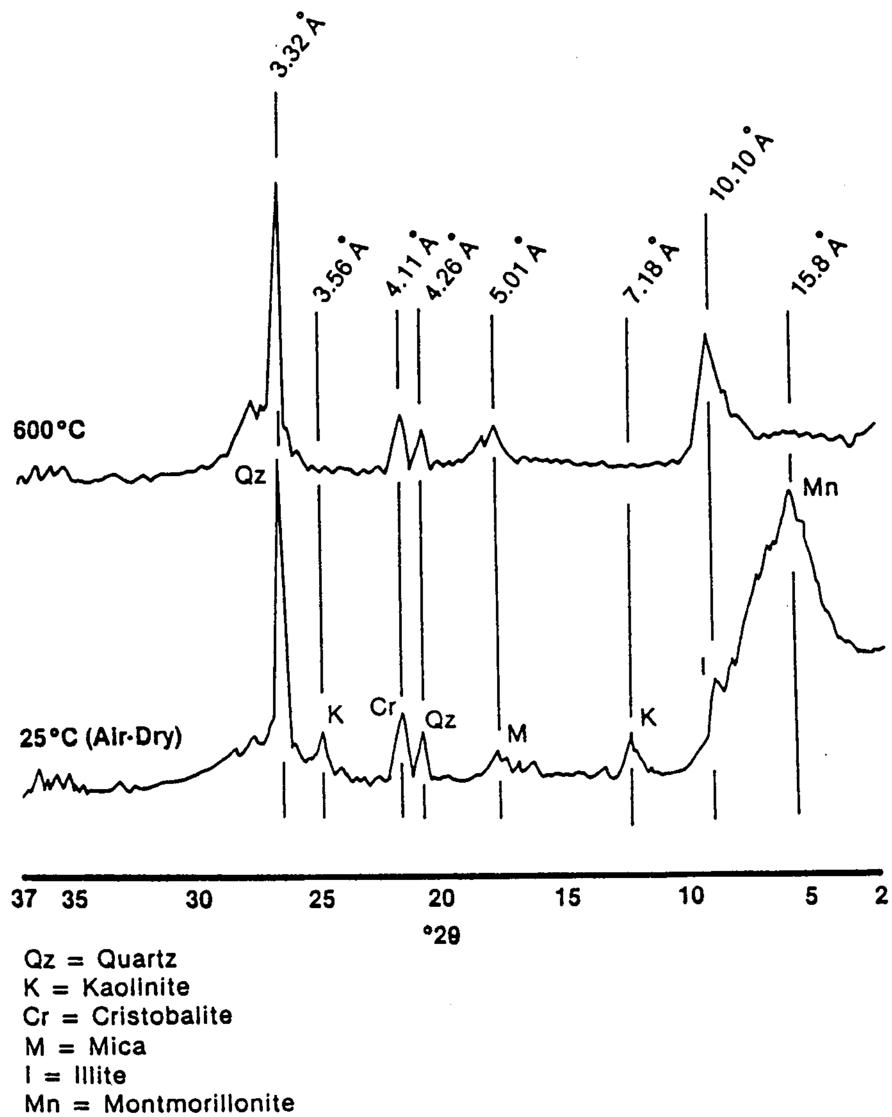

Figure 5:9 X-ray diffraction traces for clays (<2 microns) from sample collected at PDB1 section, Peoria loess (depth of 3.85 - 4.15 m). See Figure 5:7 for sample location. Material sampled was deposited sometime between 13,000 and 11,500 yr B.P. X-ray diffraction traces are for air-dried (25° C) and oven-dried (600° C) samples. Mineral assemblage is typical of the Peoria loess, as reported by Swineford and Frye (1951).

radiocarbon age on the paleosol, the silty texture and clay mineralogy of the basal 3 m of the deposit, and the presence of late Pleistocene terrestrial landsnail fauna, it appears that the sediment below the paleosol is *in situ* Peoria loess.

The paleosol just above the middle of the exposure features a dark brown (10YR 3/2) Ab horizon and a prominent Btb horizon. Calcium carbonate has been leached from the Ab horizon (see Figure 5:8). The paleosol can be traced along the length of the PDB1 exposure, and hand-auger and drill cores established that a paleosol at roughly the same depth as the radiocarbon dated paleosol extends at least 1 km to the south of the exposure. The 2 m of massive, buff (10YR 5/3) coarse silt that overlies the paleosol is identical in texture and color to the Peoria loess in which the paleosol is developed (see Figure 5:7). This implies that the massive coarse silt was derived from the Peoria loess. The lack of structure in this upper silt unit suggests that it is an eolian deposit, possibly reworked from Peoria loess deposits during the early Holocene. A 0.40 m-thick surface soil with Ab and Bw horizons is formed in this upper material.

The PDB4 exposure lies to west of the PDB1 exposure (see Figures 5:6 and 5:7). Although both consist of buff (10YR 5/3), coarse silt (Figure 5:10, see Figure 5:8), and have similar clay mineral assemblages (Figure 5:11, see Figure 5:9), PDB4 lacks the paleosol that is so conspicuous in the PDB1 face (see Plate 5:3). Moreover, hand-augering to depths of 4.0 m in the surface to the south of the PDB4 exposure failed to reveal a paleosol; this is approximately twice the depth to the paleosol in the PDB1 exposure. Further, a bulk sample (ca. 1 kg) of material from the base of the PDB4 exposure yielded a landsnail assemblage consisting of <u>Discus</u> <u>shimeki</u>, <u>Discus</u> <u>cronkitei</u>, <u>Vallonia</u> <u>gracilicosta</u>, and <u>Pupilla</u> <u>blandi</u>. These species, all of which are extinct today in Kansas, are presently found in litter-covered forest areas of the Rocky Mountains and along the United States/Canadian border (Leonard 1952; Wells and Stewart 1987a). They thrive under climatic conditions of more effective moisture and a lower annual temperature than are found today on the central Great Plains. The presence of <u>D.</u> <u>Shimeki</u>, which was not noted at the PDB1 site, is particularly noteworthy in that it restricts deposition of the sediment to the period between 20,000 and 13,000 yr B.P. (J.D. Stewart Pers.

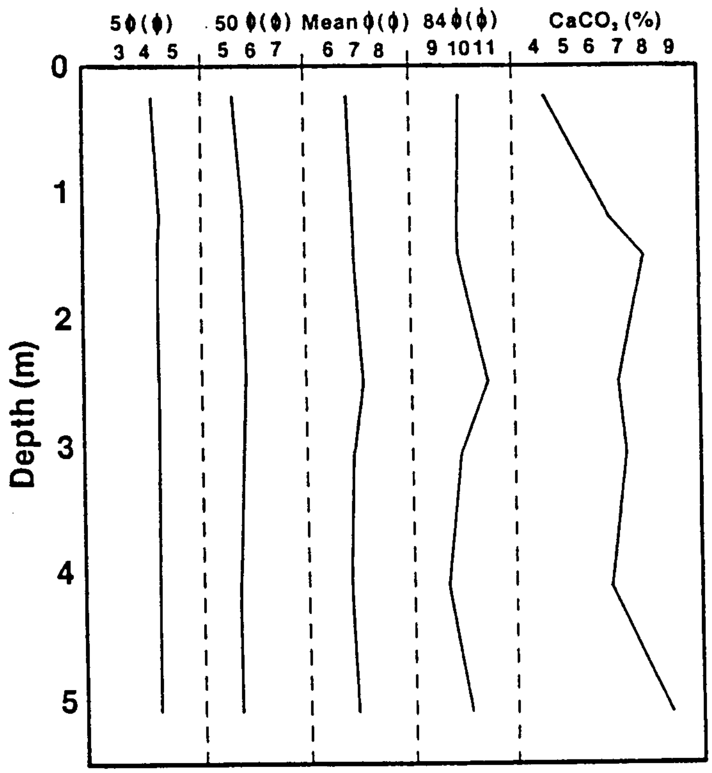

Figure 5:10 Particle size and calcium carbonate trends for PDB4 sample. Depth is measured from the banktop. Banktop height is 598 m above mean sea level.

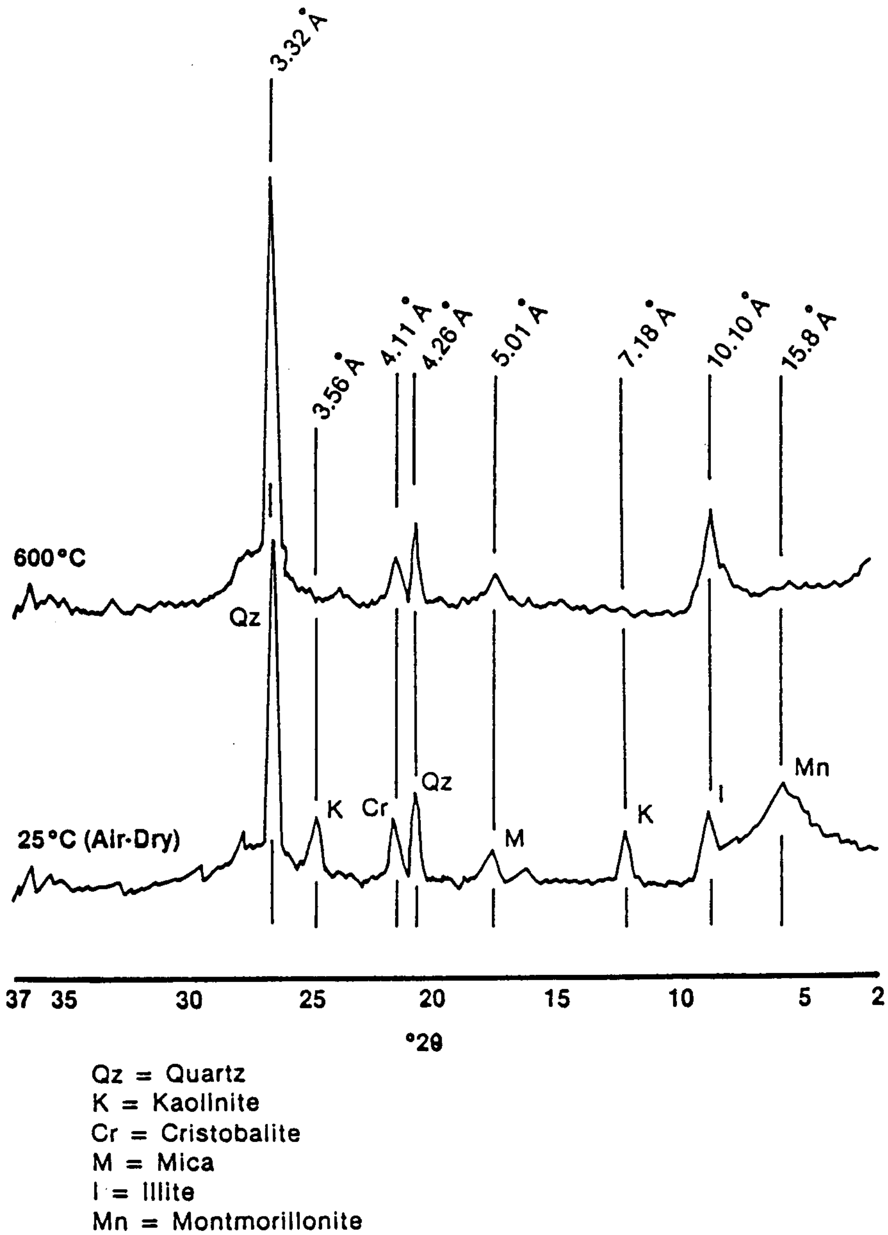

Figure 5:11 X-ray diffraction traces for clays (<2 microns) from sample collected at PDB4 section, Peoria loess (depth of 4.09 - 4.16 m). See Figure 5:7 for sample location. Material sampled was deposited sometime between 25,000 and 13,000 yr B.P. X-ray traces are for air-dried (25° C) and oven-dried (600° C) samples. Mineral assemblage is typical of the Peoria loess, as reported by Swineford and Frye (1951).

Comm. 1986), that is, before the fill at PDB1 was apparently deposited.

Similar to the PDB1 exposure, the PDB3 exposure features a paleosol (Ab/Bwb profile) developed in and buried by buff, massive coarse silt (Plate 5:5). Unlike the paleosol at PDB1, which is dark brown (10YR 5/3) and has crumb structure, the paleosol at PDB3 is dark gray (10YR2/2) and has blocky structure. It is also more clay-rich and calcareous than the paleosol at PDB1 (see Figure 5:8, Figure 5:12). Auger holes in the PDB3 surface disclosed that a paleosol is present at approximately the same depth as the one in the exposure for a distance of about 1 km back from the face. As revealed by the eighty-fourth phi percentile trend, samples collected from the PDB3 face have consistently higher clay contents (see Figure 5:12) than did samples taken from PDB1 and PDB4 (see Figures 5:8 and 5:10). X-ray diffraction traces for samples from PDB3 display the typical kaolinite-quartz-illite mineral assemblage of the Peoria loess (Figure 5:13), with slightly more kaolinite than was present in PDB1 and PDB4 (see Figures 5:9 and 5:11). Curiously, the PDB3 sample lacks cristobalite, a mineral that is present in PDB1. Although by no means conclusive, these data indicate

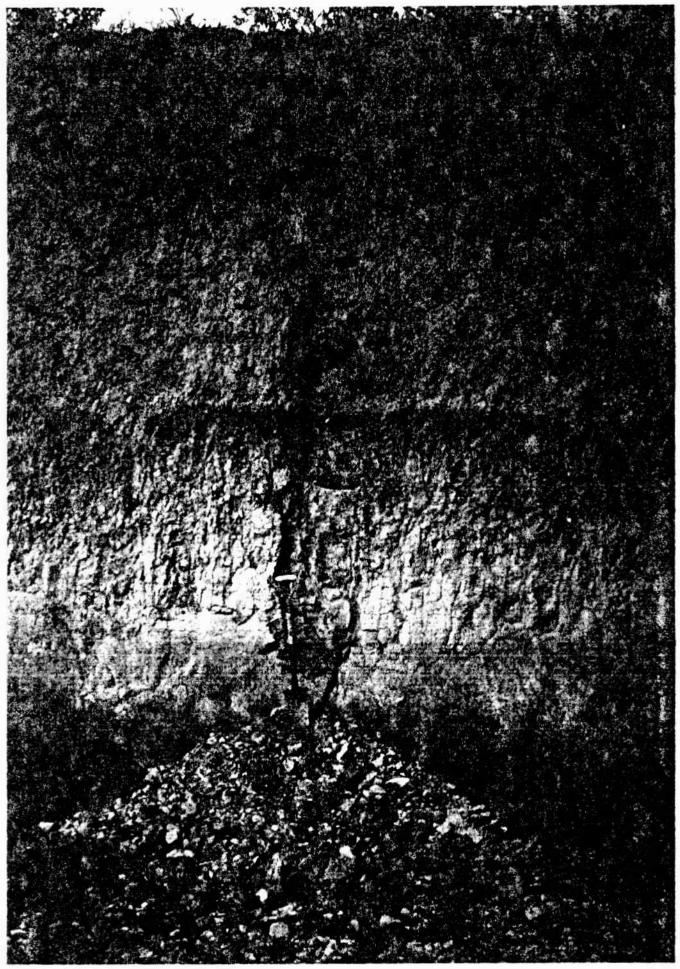

Plate 5:5 PDB3 sample section. Fill below paleosol is thought to be Peoria loess deposited sometime between 13,000 and 11,500 yr B.P. Paleosol may correlate with soil radiocarbon dated at 10,360 yr B.P. in nearby PDB1 exposure. Sediment overlying paleosol is eolian coarse silt probably reworked from Peoria loess during the early Holocene.

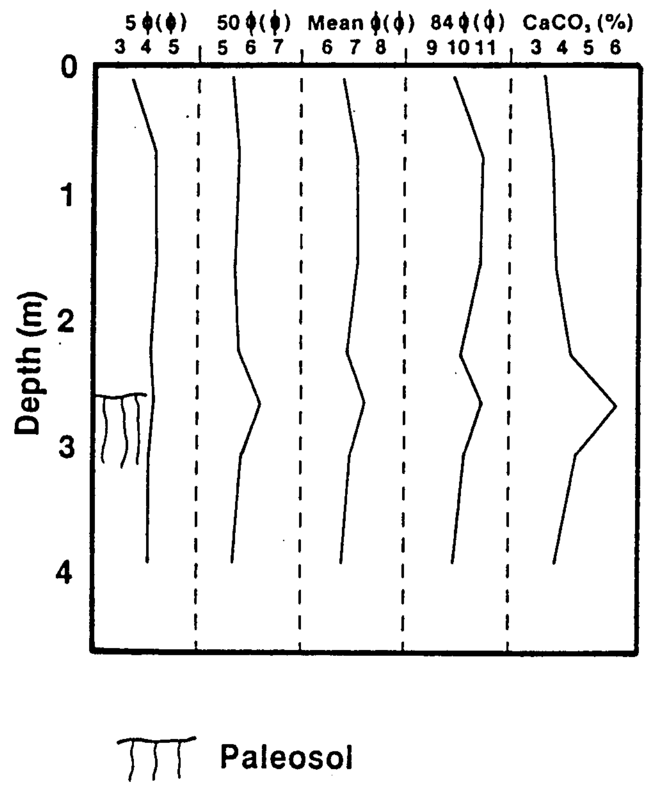

Figure 5:12 Particle size and calcium carbonate trends for the PDB3 sample section. Depth is measured from the banktop. Banktop height is 598 m above mean sea level.

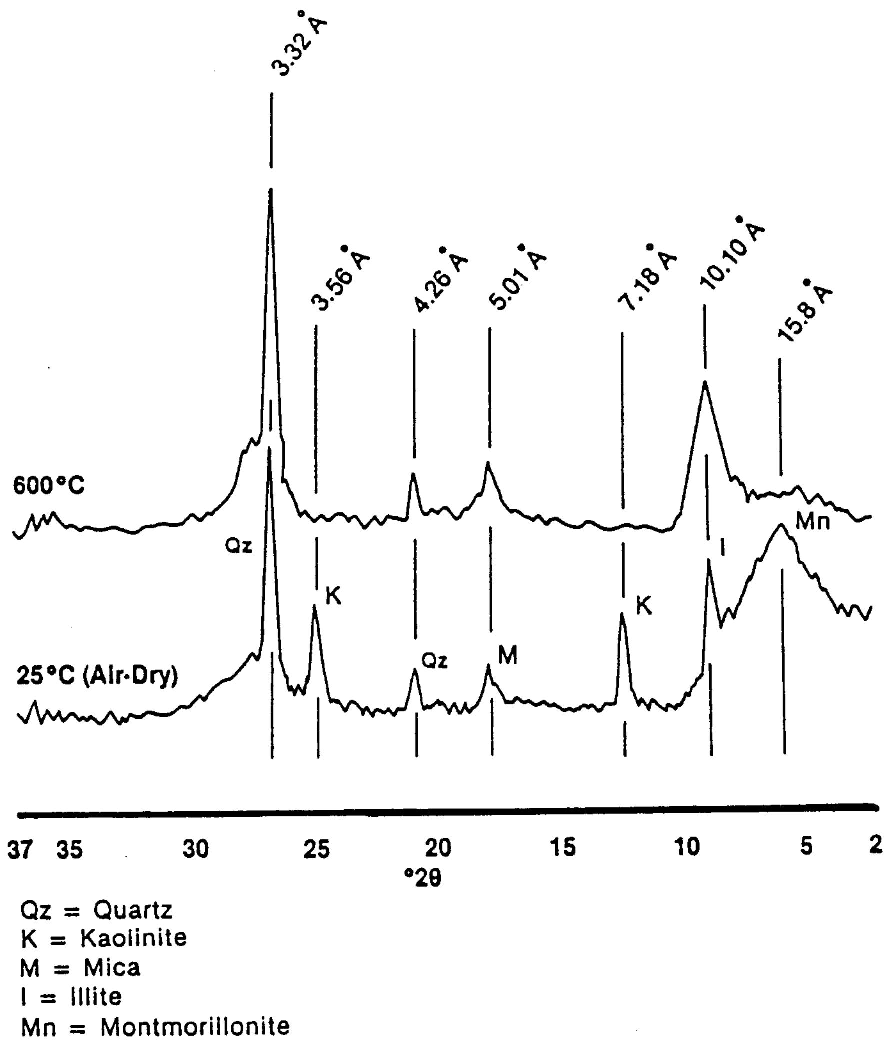

Figure 5:13 X-ray diffraction traces for clays (<2 microns) from sample collected at PDB3 exposure (depth of 3.90 - 3.96 m) below the paleosol. See Figure 5:7 for sample location. Material sampled is thought to be the Peoria loess deposited between 13,000 and 11,500 yr B.P. X-ray traces are for air-dried (25° C) and oven-dried (600° C) samples. Mineral assemblage is typical of the Peoria loess, as reported by Swineford and Frye (1951).

that the fill of PDB3 may be different than that of PDB1. It is equally possible that the PDB3 and PDB1 fills are the same age, but that the PDB3 fill has been more highly weathered. This would account for the higher clay and kaolinite contents of PDB3 relative to PDB1.

The PDB5 exposure, which lies to the west of the PDB3 exposure (see Figures 5:6 and 5:7), comprises buff-to-light brown (10YR 5/2-to-10YR 4/4) coarse silt. The fill lacks a paleosol, giving it the appearance of PDB4, and, similar to PDB4, it is highly calcareous in the upper 2 m (Figure 5:14, see Figure 5:10). Its texture is similar to that of the other three exposures.

The evidence from the four exposures at the Prairie Dog Bay site implies a complex sedimentation history between approximately 20,000 and 10,000 yr B.P. To summarize my observations at the site: 1) A radiocarbon date of 10,360 yr B.P. on the paleosol in the PDB1 exposure indicates that deposits at that site date to the late Pleistocene; 2) Although the banktop heights of the four exposures are different (PDB4 is the highest, PDB5 the lowest), the relief between the highest surface and the lowest one is only 1.5 m (see Figure 5:7); 3) PDB1 and PDB3 contain a paleosol that differs in color,

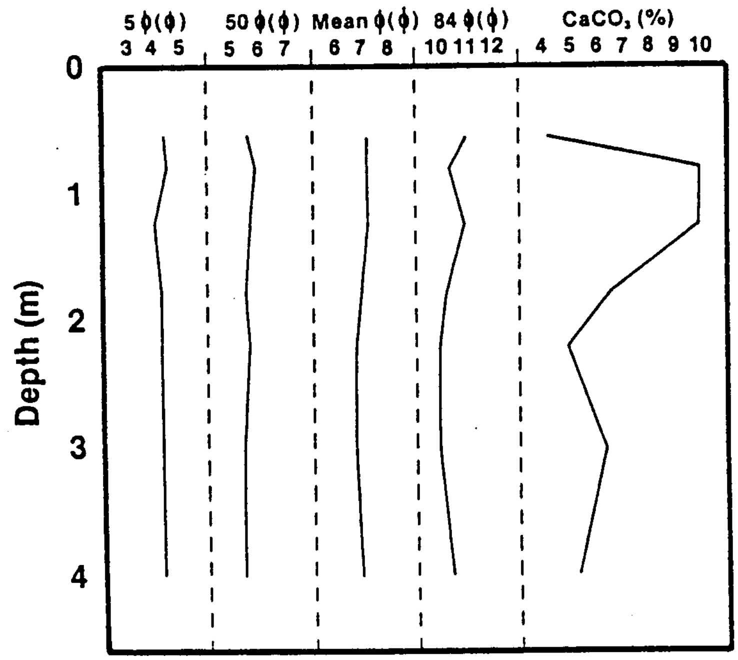

Figure 5:14 Particle size and calcium carbonate trends for PDB5 sample section. Depth is measured from banktop. Banktop height is 598 m above mean sea level.

structure, and texture, whereas PDB4 and PDB5 do not contain a paleosol; 4) Landsnails found near the base of the PDB1 exposure date to the period 13,000 to 10,500 yr B.P., whereas landsnails found at the base of PDB4 date to the period 20,000 to 13,000 yr B.P.; 5) Sedimentary structures are absent from all fills at the site, suggesting that material is wind-deposited; and 6) Together, the fills constitute the extensive (3 km^2), nearly level Prairie Dog Bay surface.

The radiocarbon age from the middle of the PDB1 exposure and the landsnail assemblages from the lower half of it confirm that the exposure consists of *in situ* Peoria loess that was likely deposited prior to 10,360 yr B.P. This terminal age for deposition of the Peoria loess is close to the accepted range (14,000 to 12,000 yr B.P.) for the end of Peoria loess deposition in the Great Plains (Thompson and Bettis 1980; Ruhe 1983). Landsnails from the contiguous PDB4 exposure place deposition of that fill to between 20,000 and 13,000 yr B.P., in other words, immediately prior to deposition of the PDB1 fill. Based on this evidence, I postulate that PDB4 is a remnant of an extensive loess surface that was deposited between 20,000 and 13,000 yr B.P., and then entrenched before 13,000 yr B.P. Following this incision, the loess in the lower half of PDB1 was deposited.

The relationship of the PDB3 and PDB5 exposures to this sequence of events is less clear, however. The small (up to 1.5 m) differences among banktop heights of the four sites cannot be construed as proof that four distinct fills exist; the differences in elevation may have resulted from post-depositional erosion, and indeed they become more difficult to detect as one moves from the exposures across the Prairie Dog Bay surface to the upland. The absence of a paleosol and the similarities in texture (see Figures 5:10 and 5:14) and color suggest that PDB4 and PDB5 comprise fill of the same age. Based on the landsnail fauna from the base of PDB4, that fill (Peoria loess) was deposited between 20,000 and 13,000 yr B.P., and therefore predates the PDB1 fill which was deposited between 13,000 and 10,500 yr B.P. Fitting the PDB3 fill into the sequence is more problematic. The presence of a paleosol in PDB3 demonstrates that the fill is not contemporaneous with the PDB4 and PDB5 fills. On the other hand, the PDB3 fill does not match the PDB1 fill. The PDB3 fill is more clay-rich than the PDB1 fill, and the paleosol in PDB3 is more clay-rich and more gray in color than the paleosol in PDB1 (see Figures 5:8 and 5:12). These differences suggest that either the fills are different, or that PDB3 is more weathered than is PDB1. Assigning the two fills to different depositional events further complicates the chronology of events by introducing a third period of aggradation. Moreover, it is not clear that the aforementioned differences between PDB1 and PDB3 fill are significant. These differences notwithstanding, I propose that PDB3 correlates with PDB1.

On the basis of the stratigraphic relationship among the four exposures, I have constructed the following chronology of events (Figure 5:15). First, deposition of Peoria loess before 13,000 yr B.P. formed a broad loess surface. Second, prior to 13,000 yr B.P., this loess surface was entrenched and partially removed, leaving PDB4 and PDB5 as remnants; a similar episode of entrenchment along the middle Platte River, Nebraska, sometime after 14,000 yr B.P. was noted by May (1989a). Third, Peoria loess accumulated until about 13,000 yr B.P., filling the recently-entrenched valleys and forming the loess units at PDB1 and PDB3; some loess probably accumulated on PDB4 and PDB5, as well. Fourth, the paleosols at PDB1 and PDB3 then formed during a

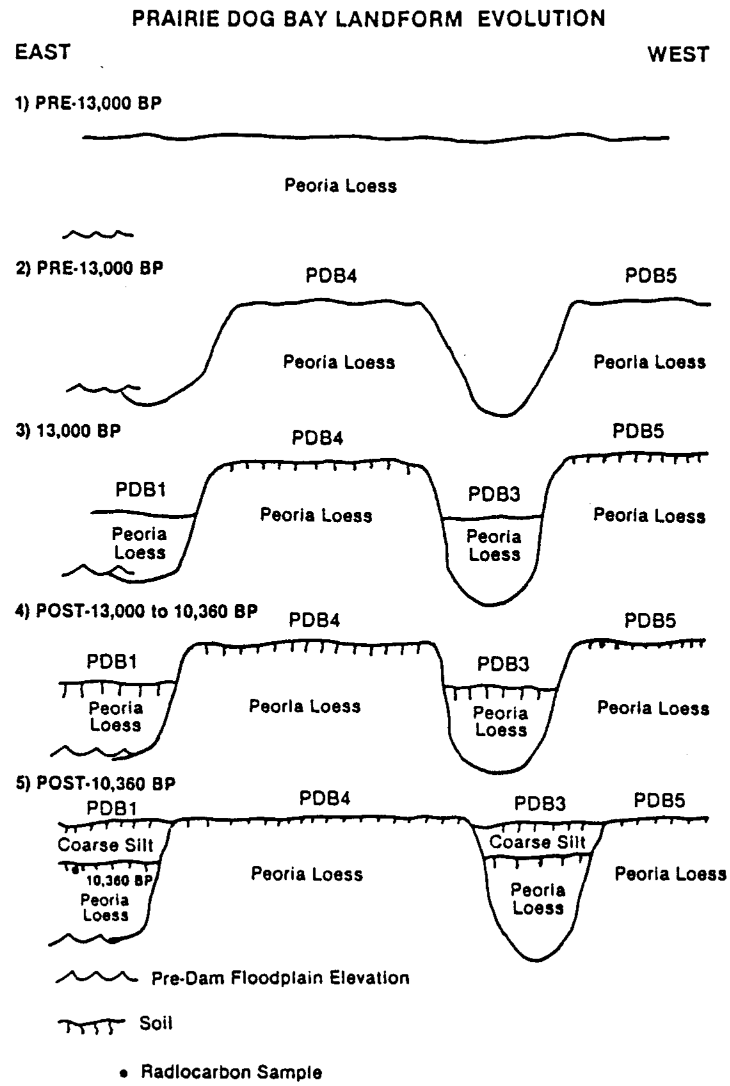

Figure 5:15 Chronology of events at Prairie Dog Bay site, pre-13,000 yr B.P. to early Holocene. View is from Prairie Dog Bay looking south, with east on the left and west on the right of the profile. Height and width of exposures are not to scale. Chronology is as follows: 1) Pre-13,000 yr B.P: Deposition of Peoria loess to form broad loess surface; 2) Pre-13,000 yr B.P.: Entrenchment of loess surface and removal of loess; 3) 13,000 yr B.P.: Renewed deposition of Peoria loess in recently-cut valleys; 4) 11,500 to 10,360 yr B.P.: Soil development on the recently-deposited Peoria loess; and 5) Post-13,000 yr B.P.: Eolian deposition of coarse silt to form the extensive Prairie Dog Bay surface.

period of surface stability and soil formation that lasted until 10,360 yr B.P.; it was during this same interval of stability that the Brady paleosol developed across at least part of the central Great Plains. Finally, surface stability ended after 10,360 yr B.P. with eolian deposition of coarse silt, probably reworked locally from deposits of Peoria loess. A local source for this coarse silt would explain why it is nearly identical in texture and color to the Peoria loess in which the radiocarbon dated paleosol is developed. The absence from the coarse silt of any fluvial structures indicates that it is an eolian rather than a fluvial deposit.

North Cove

The North Cove site, located on the north shore of Harlan Lake about 6 km southwest of Republican City, has yielded a wealth of paleoenvironmental and stratigraphic information about late Pleistocene and early Holocene conditions in the study area (Plate 5:6, Figure 5:16). Numerous late Pleistocene vertebrate and invertebrate aquatic and terrestrial fossils, many of a boreal affinity, were recovered from spring deposits near the base of the north end of the North Cove exposure

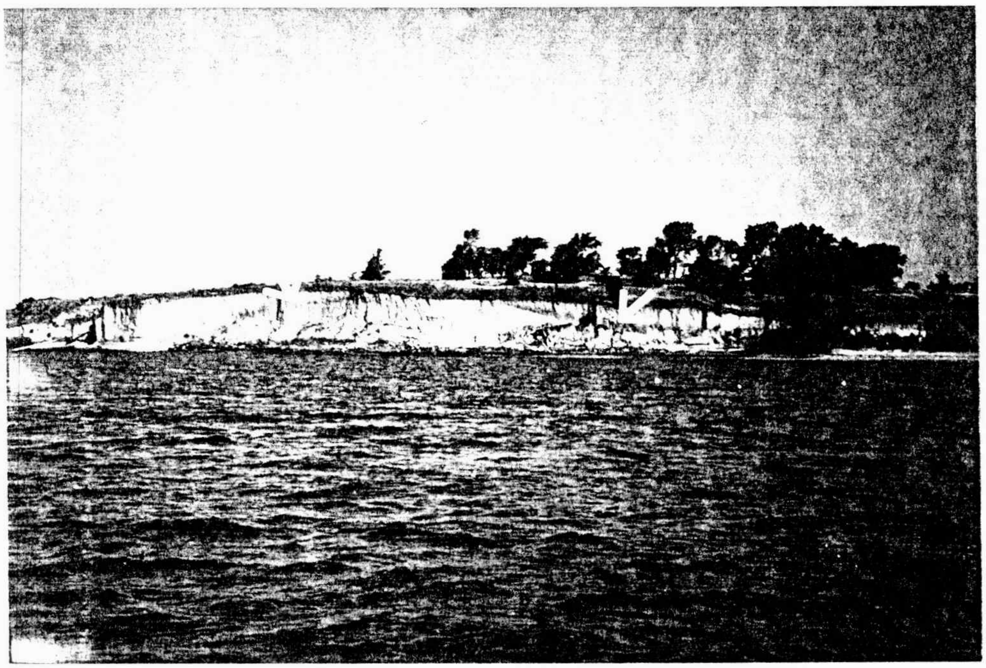

Plate 5:6 North Cove site. Upper Ab of paleosol on the left (arrow) was radiocarbon dated at 16,130 ± 130 yr B.P. (DIC-3358), and upper Ab of paleosol on the right (arrow) at 10,550 ± 160 yr B.P. (Tx-6319).

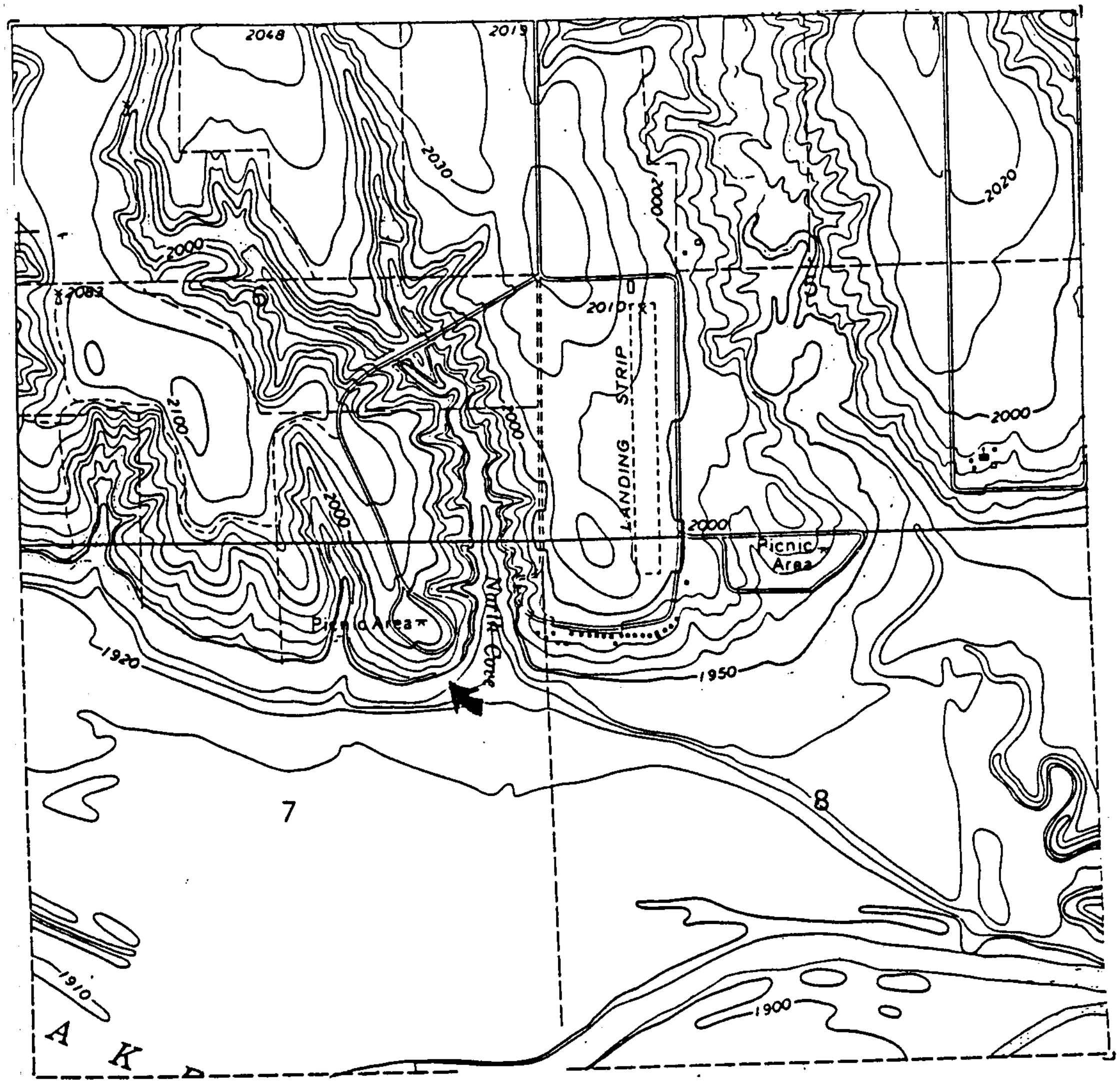

Figure 5:16 Section of topographic map showing location of North Cove sample site (arrow). Site located in: N 1/2, NE 1/4, Sec. 5., T1N, R18W Alma NE/KS 7.5 minute quadrangle. Top of the map is north, contour interval is 10 feet. The pre-dam channel of the Republican River can be seen in the lower portion of section 8.

(Figure 5:17). Among the extracted fauna were _Bootherium bombifrons_ (woodland muskox), _Sorex cinerus_ (masked shrew), _Sorex articus_ (arctic shrew), _Lepus americanus_ (snowshoe hare), _Microtus xanthognathus_ (yellow-cheeked vole), _Tamias minimus_ (least chipmunk), _Phenacomys intermedius_ (heather vole), and _Dendragapus canadensis_ (spruce goose); prior to their collection at the North Cove site, several of these species had been unrecorded in Nebraska (Stewart 1989). In addition to faunal remains, _Picea glauca_ (white spruce) wood, cones, and needles were recovered from an organic-rich, clayey lens lying adjacent to the spring deposits. Radiocarbon ages of 12,965 ± 135 yr B.P. (UGa-5476), 13,100 ± 140 yr B.P. (UGa-5477), and 12,650 ± 250 yr B.P. (Beta-18,188) were obtained on the _Picea sp._ wood (Johnson 1989). The site was also excavated for paleoindian artifacts in September, 1987, although no conclusive evidence of a cultural presence was unearthed.

Although pollen preservation in the deposits is poor, counts from three samples of sediment dating between 14,500 and 12,500 yr B.P. revealed approximately 68% arboreal pollen; the arboreal component was dominated by _Picea_ (35%), with lesser amounts of _Pinus_ (5%), _Juniperus_ (juniper) (3%), _Populus tremuloides_

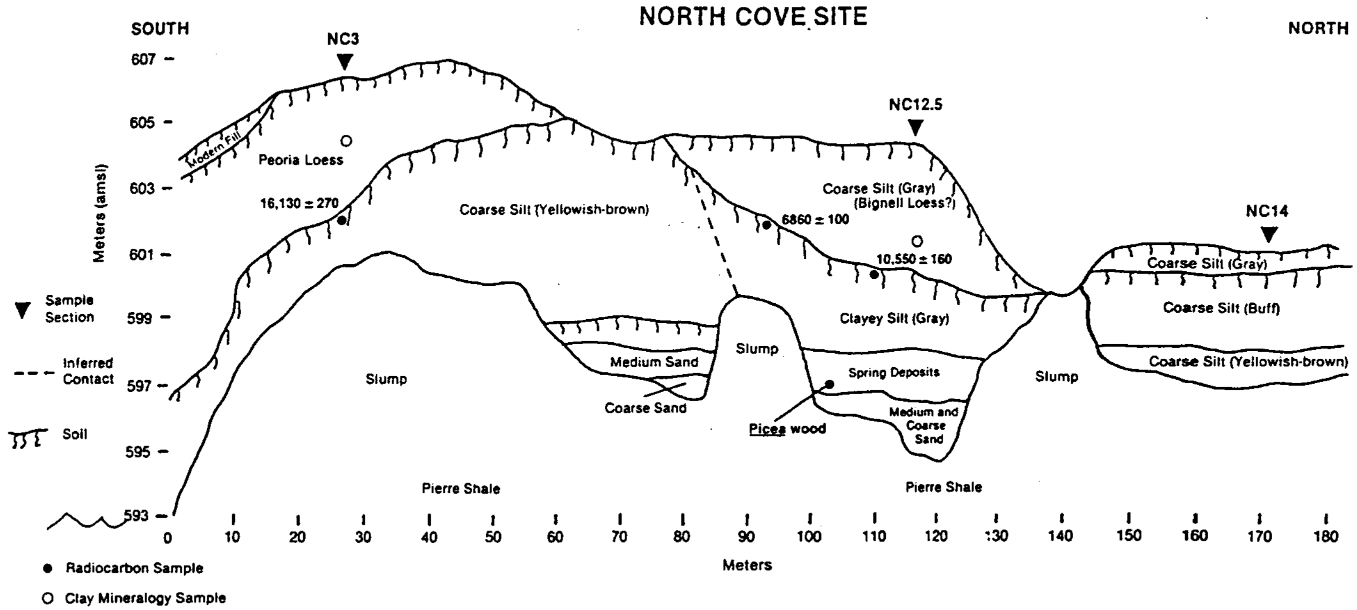

Figure 5:17 Profile of North Cove exposure and sample localities NC3, NC12.5, and NC14. View is from Harlan Lake facing northwest, with north on the right and south on the left of the profile. Surface height is given in meters above sea level (amsl). Spring deposits from which the late Pleistocene floral and faunal remains were extracted lie in the center of the exposure at sample section NC12.5. Note location of radiocarbon and clay mineral samples.

(aspen) (2.6%), and _Quercus_ (1.8%). The most common constituent of the nonarboreal component was Poaceae (grass) (9%), followed by various species of Cyperaceae (sedge) (8.9%), _Artemesia_ (big sagebrush) (2.9%), and _Ambrosia_ (giant ragweed) (2.3%) (Fredlund 1989). Finally, a distal humerus, skull, and vertebra of _Bison occidentalis_ were recovered from an area just to the north of the spring; the humerus was radiocarbon dated at 11,365 ± 865 yr B.P. (UGa-5480), and the vertebra at 11,020 ± 635 yr B.P. (UGa-5475) (Johnson et al. 1986).

On the basis of these floral and faunal remains, a preliminary picture of late Pleistocene environmental conditions in and around the North Cove site has begun to emerge. It appears that the area immediately adjacent to the spring may have supported a mixed _Picea_/deciduous parkland (Fredlund 1989), although the vegetative composition of this parkland on uplands is uncertain; the presence of _Populus_ pollen implies, however, that at least some aspen was present on uplands (Fredlund 1989). _Populus_ pollen had previously been recovered from the Sanders Well upland site in southeastern Kansas (Fredlund and Jaumann 1987). The recovery of a _Mammut americanum_ (mastadon) vertebra from the shore of the North Cove site also suggests that some woodlands were present on uplands during the late Pleistocene (Stewart 1989). The vertebrate remains from the spring deposit indicate permanent, cool water, and a cover comprising brush, and possibly coniferous, vegetation (Stewart 1989).

Sediments at the site range in age from pre-20,000 yr B.P. at the southern end of the exposure to middle Holocene at the northern end. At sample site NC3 (southern end of the exposure), basal cross-bedded river sands are overlain by yellowish-brown (10YR 5/4), calcareous, hard, fine sand-to-coarse silt (Figure 5:18). At the time sampling was conducted, the lower half of the exposure was masked by slump (see Figure 5:17); subsequent removal of this slump has revealed at least two faint paleosols developed in the yellowish-brown unit. A sample collected from the upper 0.10 m (3.81 - 3.88 m) of a brown (10YR 4/4) paleosol developed in the upper portion of the yellowish-brown unit yielded a radiocarbon age on humates (uncorrected for C13 ratio) of 16,130 ± 270 yr B.P. (DIC-3358). To my knowledge, an episode of soil formation at this time has not been reported elsewhere in the central Great Plains.

It seems likely that the date is several thousand years

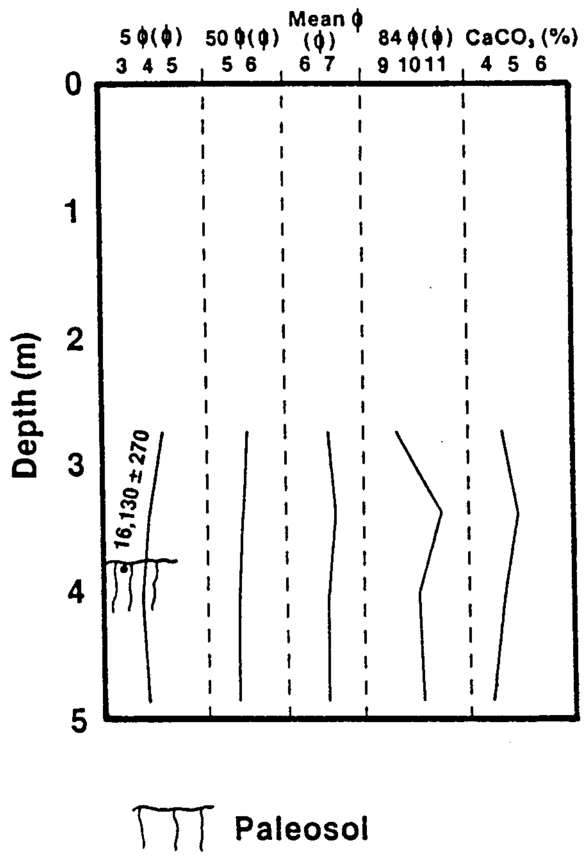

Figure 5:18 Particle size and calcium carbonate trends for NC3 sample section. Depth is measured from the banktop. Banktop height is 606 m above mean sea level.

young, possibly because the sample was contaminated from above by youthful humates. More likely, the paleosol dates to around 20,000 yr B.P., which would correlate it to the upper portion of the Gilman Canyon Formation. Buff (10YR 5/3), calcareous, soft silt that has yielded landsnails and vertebrate remains characteristic of the Peoria loess overlies this paleosol (Johnson 1989).

More germane to this research is the age of a paleosol that divides Pleistocene deposits on the southern end of the exposure from Holocene deposits on the northern end (see Figure 5:17). This paleosol, which stretches across the northern half of the exposure (see Plate 5:6), consists of a dark brown (10YR 3/1) Ab and clay-rich Btb horizons (Plate 5:7). It is darker (10YR 3/1) than the brown paleosol (10YR 4/4) radiocarbon dated at 16,130 yr B.P. at the southern end of the exposure, and the material in which it is developed is a gray (5Y 5/1), clayey silt in contrast to the yellowish-brown (10YR 5/4) fine sand-to-coarse silt in which the 16,130 yr B.P. paleosol is formed. Where the paleosol rises steeply towards the surface (see Figure 5:17), a sample collected from the upper 0.28 m (2.65 - 2.73 m) of the Ab produced a radiocarbon age on humates (uncorrected for C13 ratio) of 6860 ± 100 yr

Plate 5:7 North Cove spring deposits. Sample section NC12.5 is located in the center of photo. Basal fill (arrow) consisting of spring deposits and <u>Picea</u> wood dated at ca. 13,000 to 11,400 yr B.P. Paleosol radiocarbon dated at 11,530 $\pm$ 150 yr B.P. (Tx-6321) at base and 10,550 $\pm$ 160 (Tx-6319) at top. Material overlying paleosol is thought to be the Bignell loess, deposited after 10,500 yr B.P.

B.P. (DIC-3357); on an archaeological test pit face that had been freshly dug into the more level portion of the same soil (see Figure 5:17), a sample from the upper 0.04 m of the same Ab yielded a C13-corrected radiocarbon age on humates of 10,550 ± 160 yr B.P. (Tx-6319) (Johnson 1989).

The discrepancy between the ages probably resulted from contamination of the 6860 yr B.P. sample by youthful humates washed from above; indeed, Haas and others (1986) reported that radiocarbon ages from freshly excavated faces were up to 1000 years older than ages obtained from faces that had been exposed for several years. A paleosol uncovered on the east side of North Cove appears to be contemporaneous with the paleosol radiocarbon dated in the northern half of the North Cove site; it produced a C13-corrected radiocarbon age on humates of 10,270 ± 160 yr B.P. (Tx-6320) from the upper 0.30 m, and a C13-corrected radiocarbon age on humates of 11,530 ± 150 yr B.P. (Tx-6321) from the lower 0.20 m (Johnson 1989). Assuming these two soils are the same, as is indicated by the radiocarbon ages, then pedogenesis at the site was underway by at least 11,400 yr B.P., and continued until about 10,300 yr B.P.

Sample section NC12.5, which extended from the surface of the exposure to the late Pleistocene spring deposits (see Figure 5:17), reveals that clayey silt underlies the paleosol (Figure 5:19). This clayey silt unit, which is gray in color (5Y 5/1), features organic-rich laminations that likely developed in a wet, well-vegetated environment such as a backswamp, swale, or the edge of a spring or stream; the reduced nature of the sediment and its low calcium carbonate content (see Figure 5:19) are further evidence of deposition in a saturated environment.

The stratigraphic relationship between these lowland spring or streamside deposits and the upland late Pleistocene loess is not fully understood. Assuming that the gray clayey silt unit and its rich faunal and floral assemblages accumulated in a lowland stream environment sometime between 14,500 and 12,500 yr B.P., as is suggested by radiocarbon ages and floral and faunal remains, then there must be a channel cut in the loess somewhere to the south of the spring or streamside deposits; based on the radiocarbon ages of ca. 13,000 yr B.P. from the spring deposits, it would appear that this incision was contemporaneous with late Pleistocene incision noted at the Prairie Dog Bay site as well as

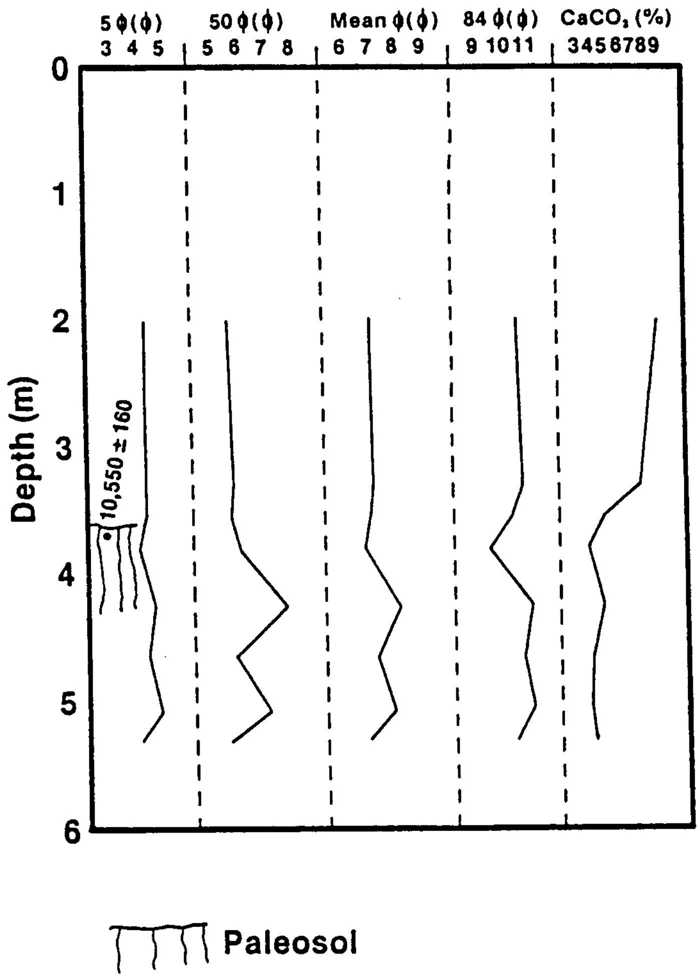

Figure 5:19 Particle size and calcium carbonate trends for NC12.5 sample section. Depth is measured from the banktop. Banktop height is 605 m above mean sea level.

elsewhere in Nebraska (May 1989a). Despite careful searching, hindered in many places by slump, this cut has not yet been found. The contact may be obscured because the inset and *in situ* fills both have a coarse silt texture, or it may be hidden by micro-slumping on the exposure face. I depict the inferred position of the contact with a dashed line on Figure 5:17.

The material overlying the paleosol is dark brown (10YR 4/3), homogeneous, calcareous coarse silt whose phi values (see Figure 5:19) and x-ray diffraction traces (Figure 5:20) are identical to those of the Peoria loess at the south end of the exposure (Figure 5:21, see Figure 5:18); the material that overlies the paleosol at NC12.5 is more calcareous and darker, however, than the Peoria loess at NC3, suggesting that it is a different depositional unit. The surface of the deposit at NC 12.5 grades smoothly to upland loess deposits, an indication that the material is not a fluvial fill. Rather, it appears that it is either colluvium that was blown or washed off upland deposits sometime after 10,500 yr B.P., or the Bignell loess.

Several observations support the hypothesis that it is the Bignell loess. First, in contiguous Phillips County, Kansas, Leonard (1952a) identified a light gray

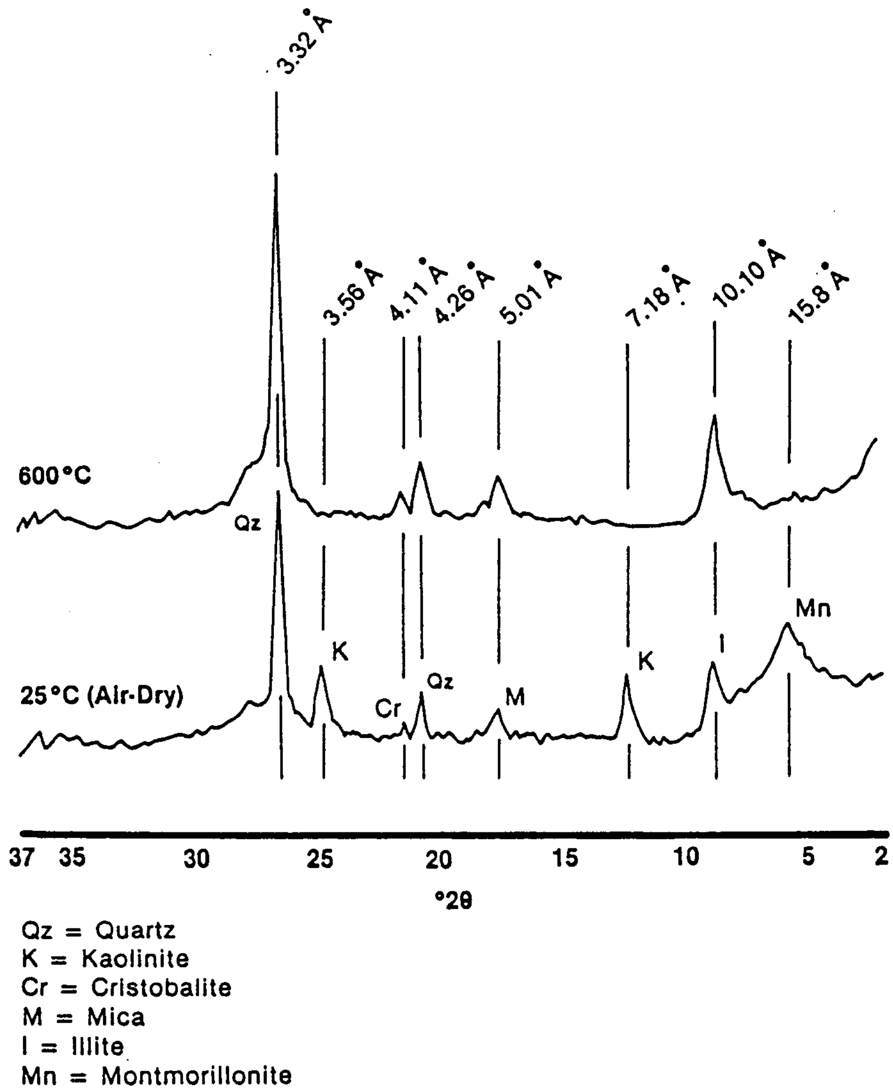

Figure 5:20 X-ray diffraction traces for clays (<2 microns) from sample collected at NC12.5 section, Bignell loess (?) (depth of 3.30 - 3.35 m). See Figure 5:17 for sample location. Material sampled was deposited shortly after 10,300 yr B.P. X-ray traces are for air-dried (25^{o} C) and oven-dried (600^{o} C) samples. Mineral assemblage is identical to that of the Peoria loess (see Figure 5:21) as described by Swineford and Frye (1951), which suggests that the Bignell loess at the site was derived from Peoria loess.

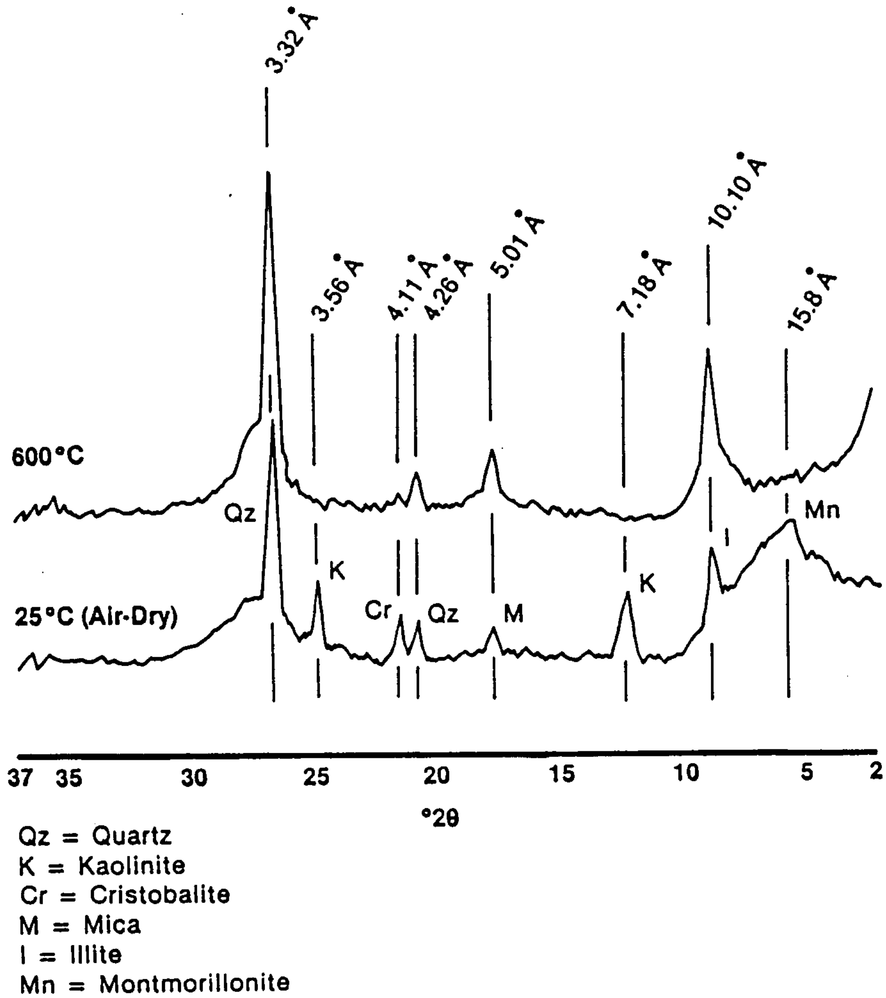

Figure 5:21 X-ray diffraction traces for clays (<2 microns) from sample collected at NC3 section, Peoria loess (depth of 2.75 - 2.80 m). See Figure 5:17 for sample location. Material sampled was deposited sometime between ca. 20,000 and 13,000 yr B.P. X-ray traces are for air-dried (25° C) and oven-dried (600° C) samples. Mineral assemblage is typical of the Peoria loess, as reported by Swineford and Frye (1951), and identical to that of a sample (Bignell loess?) from NC 12.5 (see Figure 5:20).

loess unit overlying the Brady paleosol. More recently in Phillips County, W.C. Johnson (Pers. Comm. 1989) reported that the Bignell loess rests on the Brady paleosol in a cut in Peoria loess, a topographic position that is identical to that of the fill at North Cove. Second, the fill at North Cove is similar in color to the Bignell loess (gray silt, darker in color than the Peoria loess) at its type section in southwestern Nebraska (Schultz and Stout 1945; Schultz and Martin 1970). Finally, the Bignell loess has been identified along the Republican River by previous workers in the region (e.g., Condra et al. 1947).

Sample section NC14 is located at the northern end of the North Cove site (see Figure 5:17). It contains a paleosol and gray clayey silt unit that, based on similarity in color, appears to be an extension of the paleosol and clayey silt unit present in NC 12.5. The exposure also features a basal yellowish-brown (10YR 5/4), fine sand-to-coarse silt fill. The color and texture of this yellowish-brown fill (Figure 5:22) are similar to those of the yellowish-brown fine sand unit observed in the lower part of NC3 at the south end of the exposure, and to those of the yellowish-brown fine sand unit that contains the paleosol dated at 26,260 yr

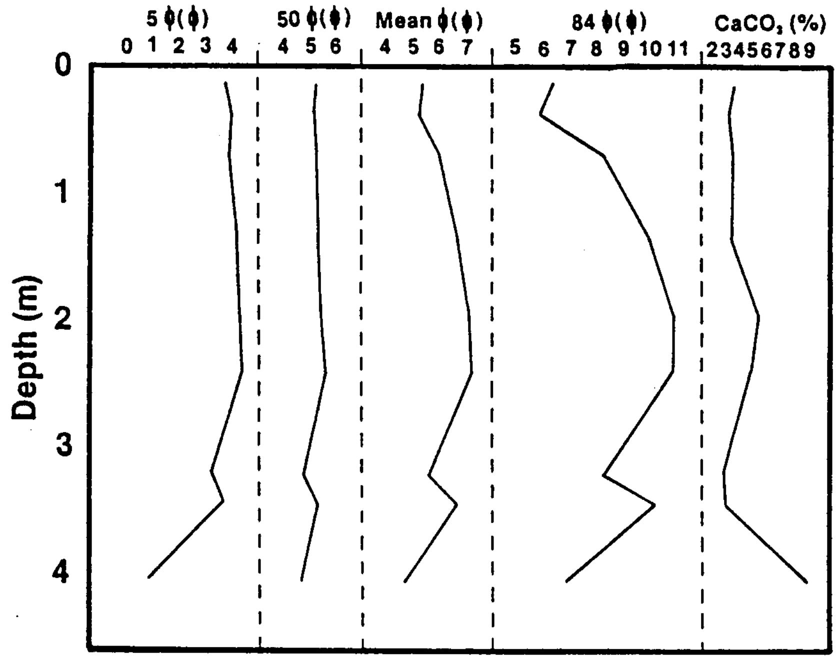

Figure 5:22 Particle size and calcium carbonate trends for NC14 sample section. Depth is measured from banktop. Banktop height is 601 m above mean sea level.

B.P. at the Bone Cove site (see Figure 5:3). On the basis of a radiocarbon age, the unit at the Bone Cove site has been tentatively correlated to the Gilman Canyon Formation.

To summarize, the North Cove site features a complex sequence of deposits ranging in age from the late Pleistocene (the Gilman Canyon Formation equivalent?) to the early and middle Holocene. It has also yielded strong evidence that a boreal environment existed in lowland topographic settings in the vicinity of the site during the late Pleistocene. Cross-bedded sands that extend along the base of the middle and northern parts of the exposure are the oldest deposits found at the site; they may correlate with pre-30,000 yr B.P. sands that were observed at the base of the Bone Cove exposure. In the middle and northern parts of the exposure, these sands are overlain by yellowish-brown fine sand; at the southern end of the exposure, the yellowish-brown fine sand lies directly on the Pierre shale. Unfortunately, extensive slumping in the middle of the exposure has precluded the tracing of these lower units across the length of the North Cove face. Similarities in appearance and texture suggest, however, that the yellowish-brown fine sand at the northern and southern ends of the exposure is the same fill (see Figure 5:17). At the southern end of the exposure, a paleosol developed in the upper portion of the yellowish-brown fine sand was radiocarbon dated (uncorrected for C13 ratio) at 16,130 ± 270 yr B.P. This date should be regarded as a minimum age for the soil, as it likely dates to about 20,000 yr B.P.

The Peoria loess was deposited at the North Cove site beginning sometime after 20,000 yr B.P., ending surface stability and pedogenesis. At one time, the loess likely extended across the length of the North Cove exposure, but was removed from the northern part by channel cutting prior to 13,000 yr B.P. I noted this same pre-13,000 yr B.P. erosion event at Prairie Dog Bay. Following the incision, fill accumulated in a spring, streamside, or swale topographic setting under climatic conditions that were much cooler than those of the present. The range of radiocarbon ages (12,650 to 13,100 yr B.P.) on the Picea wood recovered from these deposits indicates that deposition likely began around 13,000 yr B.P. It apparently continued until sometime before 11,400 yr B.P. (basal date on paleosol), at which time the slow aggradation ceased and soil formation was initiated. Pedogenesis lasted until approximately 10,500 yr B.P., when the paleosol was buried by fill that is likely the Bignell loess. The duration of Bignell loess deposition at the site is unknown, although work elsewhere in Nebraska suggests that Bignell loess deposition took place during the early Holocene (Lutenegger 1985; May 1989a).

Late Pleistocene Sedimentation

The Bone Cove, Prairie Dog Bay, and North Cove sites contain a valuable record of landform evolution in the Republican River valley for the period 30,000 to 10,000 yr B.P. This record is preserved in eolian deposits of the uplands; if there are fluvial deposits that date to this period, they have been eroded from the valley or deeply buried under younger fill.

A calcareous, yellowish-brown (10YR 6/4 to 10YR 5/4) fine sand deposited sometime prior to 26,000 yr B.P. is the oldest depositional unit dated in this study. At Bone Cove, a paleosol developed in the upper portion of this unit was radiocarbon dated at 26,260 yr B.P., whereas at North Cove a paleosol developed in it was radiocarbon dated at 16,130 yr B.P.; the age at North Cove is believed to be several thousand years young. The age of the unit and its stratigraphic position directly beneath the Peoria loess suggest that it correlates with the Gilman Canyon Formation, a unit that has been dated between 34,800 and 21,290 yr B.P. elsewhere in Nebraska (Dreeszen 1970; May and Souders 1988; May 1989a; D.W. May, Pers. Comm. 1989).

Deposition of the Peoria loess, a massive, homogeneous, coarse silt, is thought to have begun in the central Great Plains after 21,000 yr B.P (Thompson and Bettis 1980; Ruhe 1983). Charcoal and three incipient Ab horizons observed in loess in the vicinity of Coyote Canyon, located 6.5 km west of Bone Cove on the south shore of Harlan Lake (Figure 5:23), provide evidence that deposition was episodic (Plate 5:8). At the Prairie Dog Bay and North Cove sites, deposition was apparently interrupted by an episode of erosion prior to 13,000 yr B.P., followed by renewed loess deposition. This erosional episode has also been noted by May (1989a) in the middle Platte River valley of central Nebraska.

Deposition of the Peoria loess is believed to have ended between 14,000 and 12,000 yr B.P. (Thompson and Bettis 1980; Ruhe 1983), an age that matches the end of the loess deposition reported here. Soil development

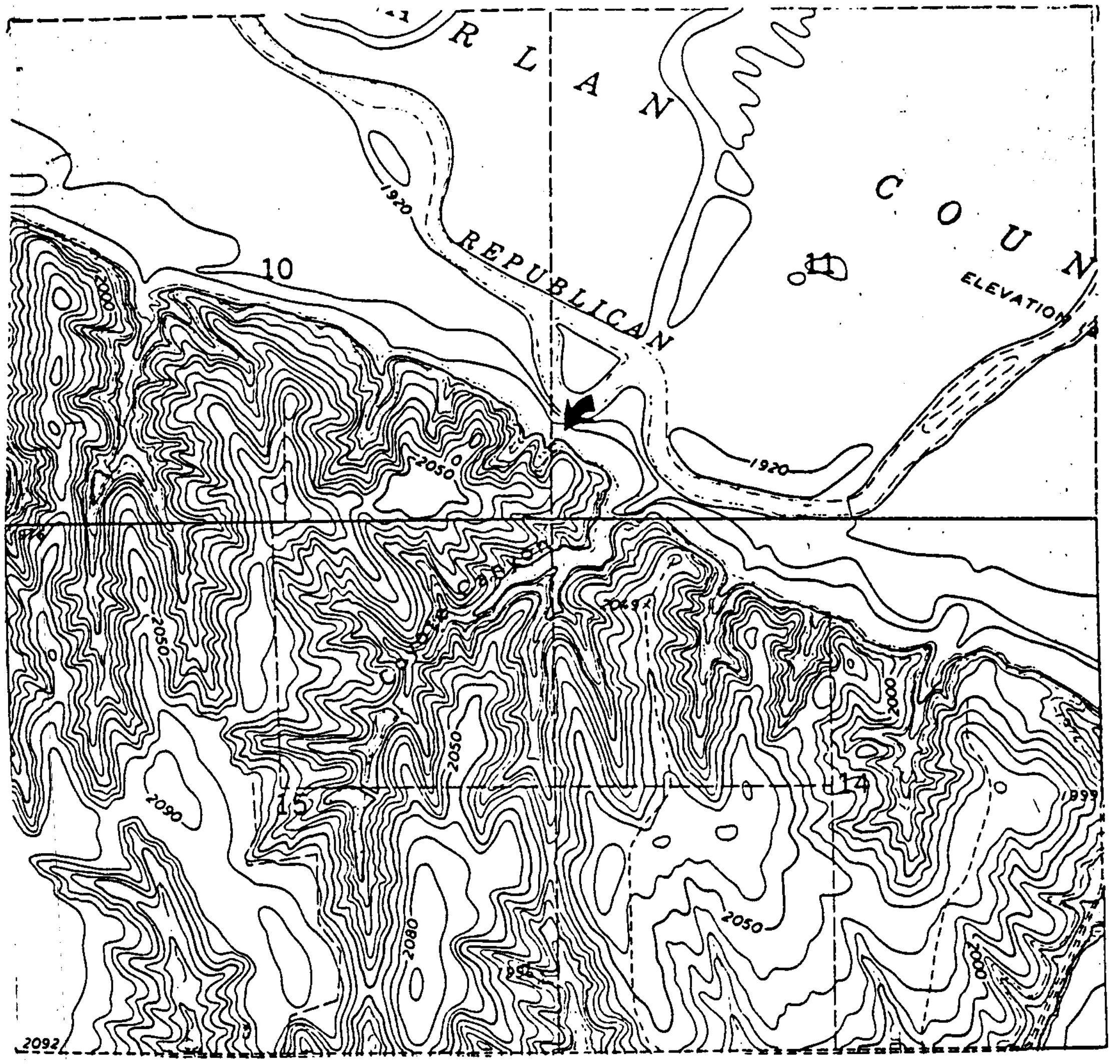

Figure 5:23 Section of topographic map showing location of Coyote Canyon. Incipient paleosols are developed in Peoria loess here, an indication that loess deposition was episodic. Charcoal identified as Picea sp. was extracted from Peoria loess (depth of 2.55 - 2.65 below banktop) at the site. Site located in: SW 1/4, SW 1/4, Sec. 11, T1N, R18W, Alma, NE/KS 7.5 minute quadrangle. Top of the map is north, contour interval is 10 feet.

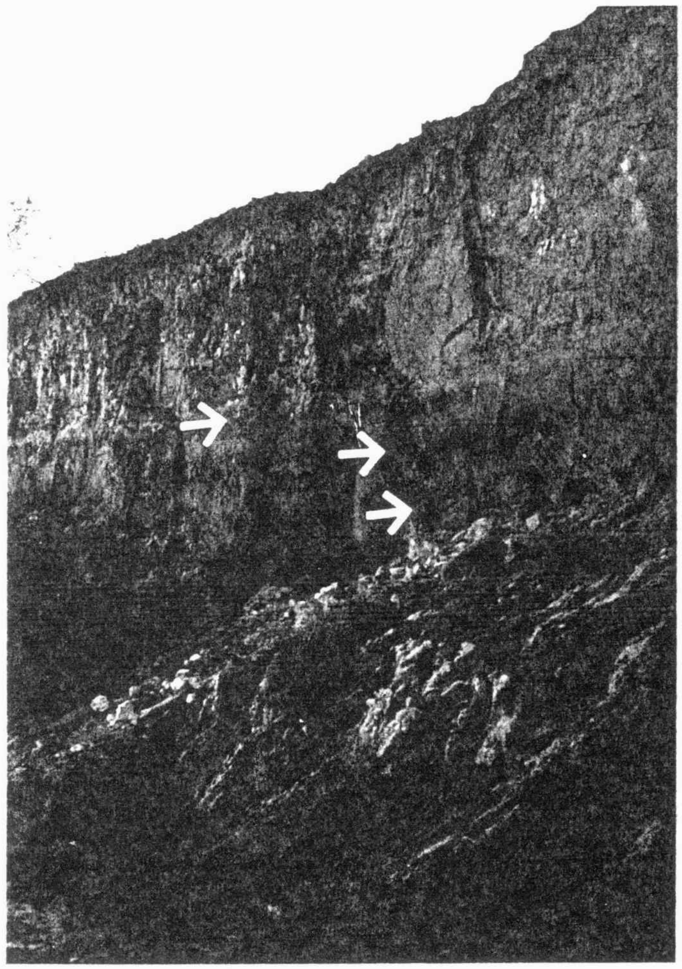

Plate 5:8 Coyote Canyon exposure. Photo shows three paleosols developed in Peoria loess (arrow), an indication that loess deposition was episodic. Exposure located in: SW 1/4, SW 1/4, Sec. 11, T1N, R18W, Alma NE/KS 7.5 minute quadrangle. From banktop to top of slump, exposure is approximately 5.30 m high.

then began in the upper portion of the Peoria loess, forming the Brady paleosol. A well-developed paleosol featuring Ab and Btb horizons, and dating to the late Pleistocene, was observed at several locations in the study area: at North Cove, it developed in the upper portion of late Pleistocene spring or streamside deposits between 11,500 and 10,500 yr B.P.; at Prairie Dog Bay, it formed in the upper portion of the Peoria loess until sometime after 10,360 yr B.P.; and at Bone Cove, it may be present as the surface soil in the upper portion of the Peoria loess. I propose that this paleosol at the three locations is the temporal equivalent of the Brady paleosol.

Late Pleistocene Paleoenvironmental Conditions

Although reconstruction of paleoenvironmental conditions is not the focus of this study, new information concerning the late Pleistocene environment of the study area has recently emerged. Such information is clearly pertinent to a discussion of landform activity. In addition to the floral and faunal remains uncovered at North Cove, remains implying that a parkland vegetational community comprising <u>Populus</u> and Poaceae existed on uplands under climatic conditions cooler than those of the present, I uncovered thin bands of charcoal in Peoria loess deposits at several locations on the south shore of Harlan Lake. Near the mouth of Coyote Canyon (see Figure 5:23), four charcoal samples extracted from Peoria loess (Plate 5:9) were identified as _Picea_ _sp._ (J. Thomas Quirk, Pers. Comm. 1988). To what extent this charcoal is representative of the upland vegetation cover during the late Pleistocene is unknown. On the basis of this evidence alone, it would be imprundent to conclude that uplands supported a mixed _Picea_/deciduous forest, the type of cover that apparently existed in the Republican River valley during the late Pleistocene. It seems more plausible, as is indicated by the limited pollen and faunal data from the North Cove site, that uplands were covered by a parkland vegetational community during that time. Increasingly, evidence from the central Great Plains (e.g., Fredlund and Jaumann 1987; Fredlund 1989; Stewart 1989) suggests that loess was deposited on a vegetated, rather than on a barren, surface.

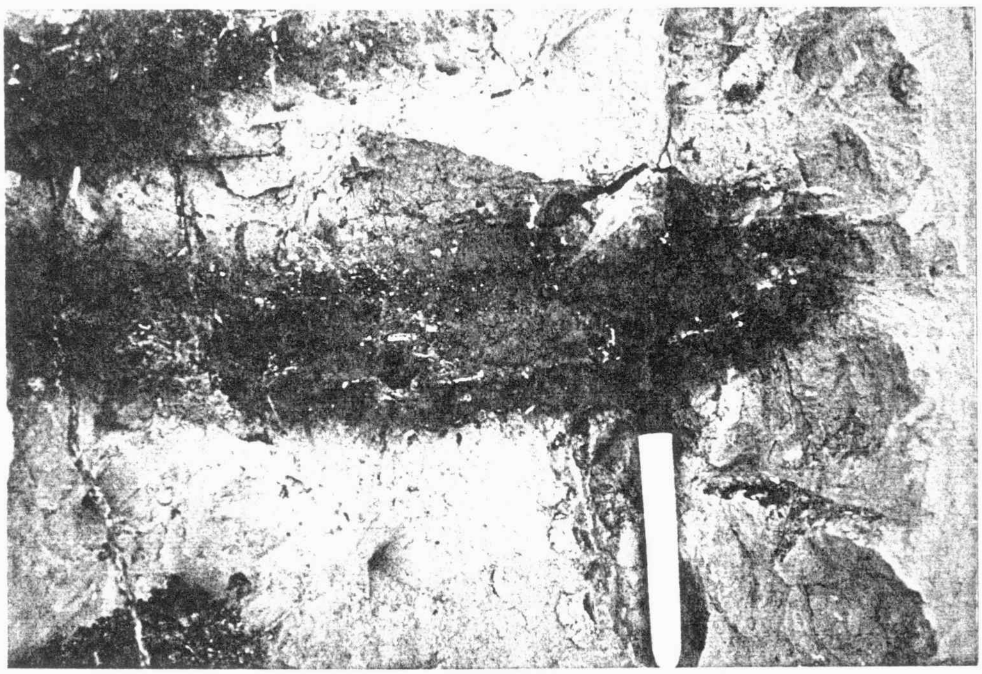

Plate 5:9 Charcoal sample in Peoria loess at Coyote Canyon exposure. Charcoal was identified as Picea sp. (J. Thomas Quirk, Pers. Comm. 1988), and suggests that Picea was present in the study area during deposition of the Peoria loess, ca. 25,000 to 11,500 yr B.P. Length of pen is 13.5 cm.

Holocene Study Sites

Alma Vista

The Alma Vista exposure is located on the south shore of Harlan Lake, about 2.5 km south of Alma and just east of the Alma Vista picnic area (Figure 5:24). Two paleosols are conspicuous in the upper half of the exposure (Plate 5:10), and a thick paleosol that was covered at the time of sampling has recently been unearthed near the base of the exposure. The uppermost and middle paleosols feature thick, structureless Ab horizons, whereas the more well-developed lowermost paleosol features Ab and Bwb horizons. Although lacking structure, the organic-rich uppermost and middle paleosols are deficient in calcium carbonate relative to other levels of the fill (Figure 5:25). Humates extracted from the upper 0.12 m (1.55 - 1.62 m) of the middle paleosol produced a C13-corrected radiocarbon age on humates of 4550 $\pm$ 80 yr B.P. (Tx-5979); neither the upper nor the lowermost paleosol has been radiocarbon dated, however.

Moving from the base to the top of the exposure, the sediment texture changes from a buff (10YR 5/3), fine sand to a buff (10YR 5/3), coarse silt. This change is illustrated by the fifth percentile value,

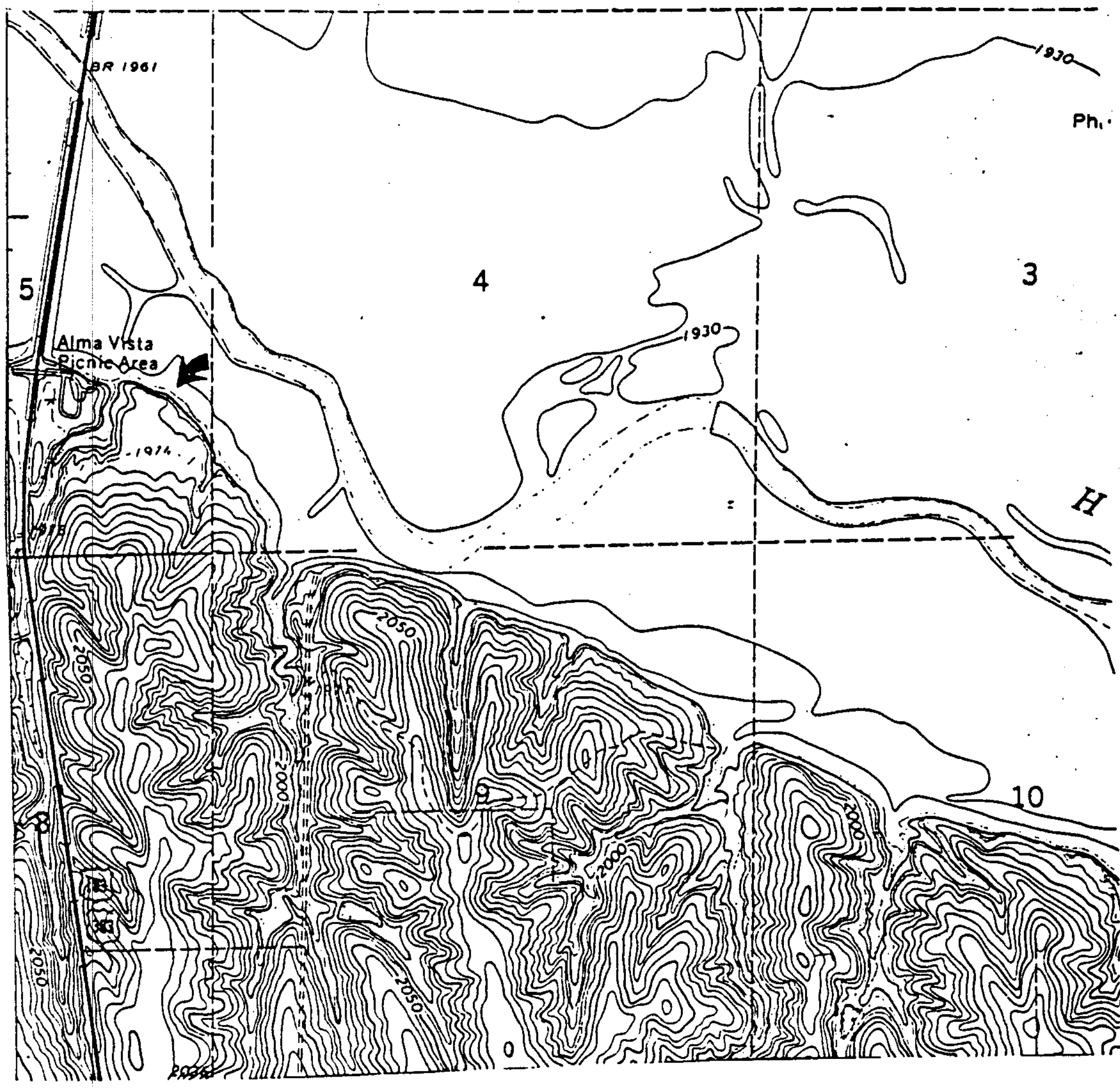

Figure 5:24 Section of topographic map showing location of Alma Vista sample site (arrow). Site located in: NE 1/4, SE 1/4, Sec. 5, T1N, R18W, Alma NE/KS 7.5 minute quadrangle. Top of the map is north, contour interval is 10 feet. Pre-dam channel of the Republican River is visible in the NE 1/4 of section 5.

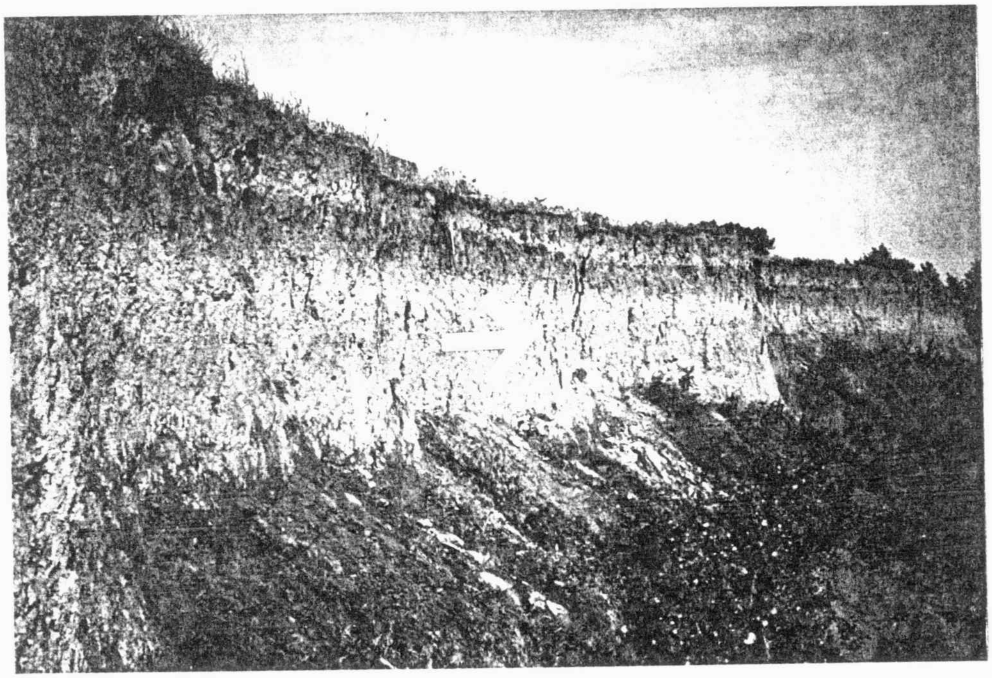

Plate 5:10 Alma Vista site. Lower paleosol visible in photo (arrow) was radiocarbon dated at 4550 ± 80 yr B.P. (Tx-5979). At a later date, an additional paleosol (lower in profile than dated soil) developed in what appears to be Peoria loess was found buried in slump (see Plate 5:11). Height of exposure is approximately 3 m from banktop to top of slump.

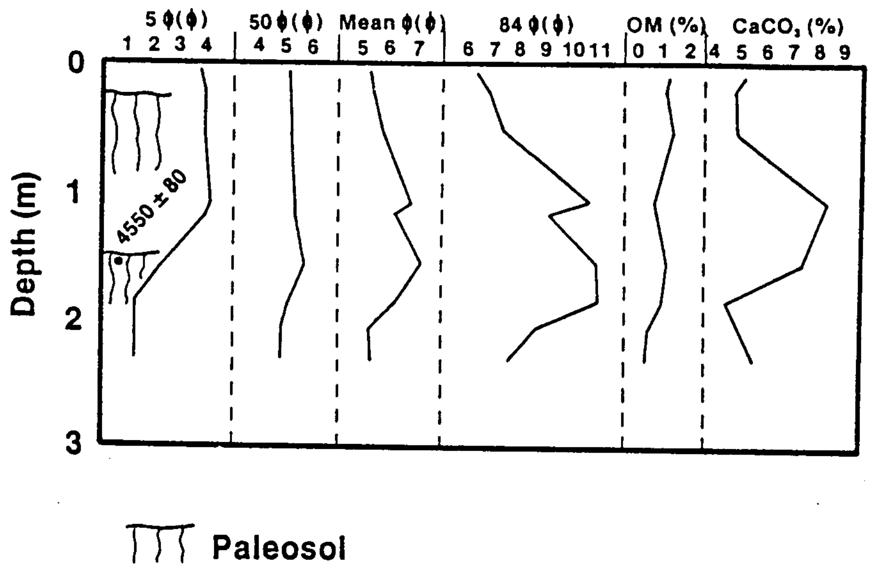

Figure 5:25 Particle size, organic matter, and calcium carbonate trends for the Alma Vista sample site. Depth is measured from banktop. Banktop height is 600 m above mean sea level.

which ranges from 1.05 at the base of the exposure to 3.95 at the banktop (see Figure 5:25). A textural break from fine sand to silt occurs at a depth of 1.50 m, as illustrated by the fifth percentile phi values (see Figure 5:25). The dark brown (10YR 3/3) middle paleosol is developed in fine sand, and is buried by buff (10YR 5/3), highly calcareous coarse silt. The dark brown (10YR 3/1) uppermost paleosol is developed in this silt and buried by a veneer of organic-rich, buff (10YR 5/3) coarse silt.

Because it was covered by slump at the time of sampling, I lack quantitative data for the lowermost paleosol and the material in which it is formed (Plate 5:11). The paleosol is developed in calcareous, coarse silt that is similar in texture, appearance, and color to the Peoria loess at the Prairie Dog Bay, Bone Cove, and North Cove sites. If the material is the Peoria loess, or its alluvial equivalent, then this lowermost paleosol could correlate to the 10,500 yr B.P. soil radiocarbon dated at the North Cove and Prairie Dog Bay sites. Well-sorted sands, apparently fluvial in origin, overlie the lowermost paleosol. Moving upsection, they grade into the buff fine sand in which the middle paleosol is developed (Plate 5:12).

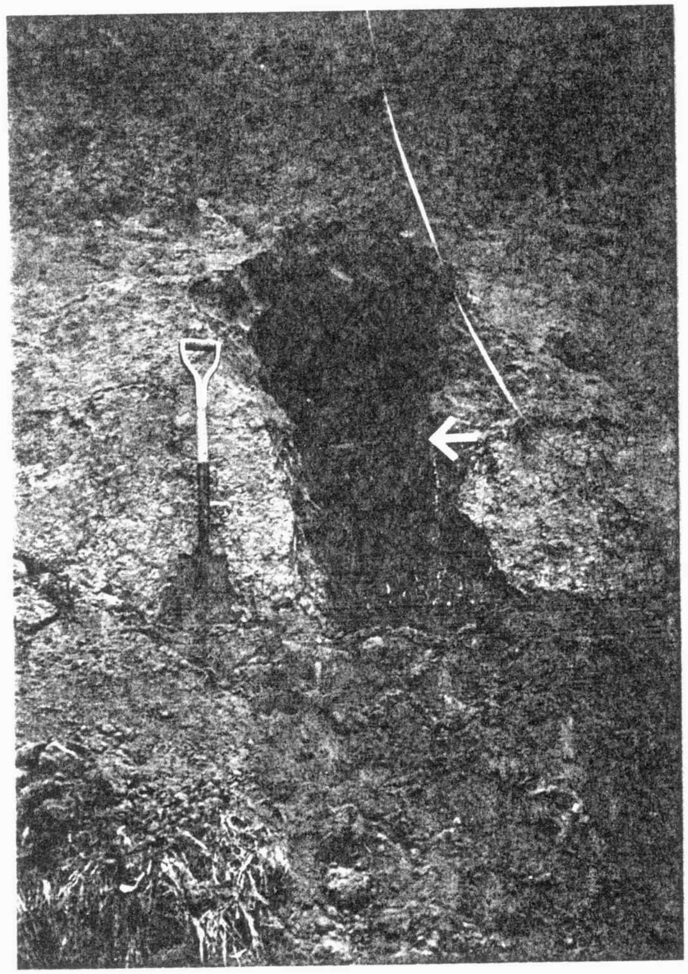

Plate 5:11 Lowermost paleosol at Alma Vista site (top marked with arrow). This soil, which was uncovered after sampling of the exposure, appears to be developed in Peoria loess, and is buried by fluvial sands (see Plate 5:12). Length of shovel is 1.08 m.

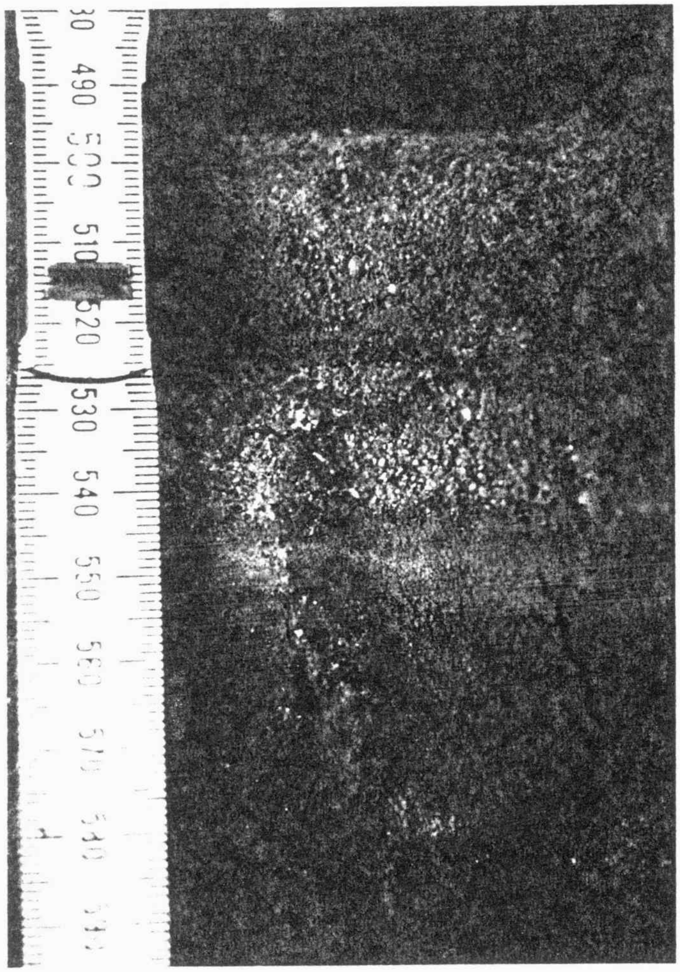

Plate 5:12 Fluvial sands overlying lowermost paleosol (see Plate 5:11) at Alma Vista site. Top of paleosol is at "552" on ruler. Ruler scale is in cm.

The surface of the Alma Vista exposure grades smoothly to upland loess deposits. Hand-augering across the Alma Vista surface to the uplands revealed two paleosols in the fill that could be traced a distance of 200 m back from the exposure (Figure 5:26). Upon reaching upland loess fill, the two paleosols disappeared, evidence that a younger fill has been inset on the valley side against older upland loess deposits.

It appears that two fills constitute the exposure. The lower fill, in which the middle paleosol is developed, comprises fluvial sand that was deposited prior to 4500 yr B.P. The upper fill, which directly overlies the middle paleosol, consists of coarse silt. Similarities between the x-ray diffraction traces (Figure 5:27) and phi values (see Figure 5:25) of the upper fill and those of *in situ* Peoria loess at the Bone Cove site (see Figures 5:5 and 5:3) indicate that the upper fill was derived from adjacent deposits of Peoria loess. The absence from the upper fill of any fluvial structures suggests that it is an eolian deposit, possibly reworked locally from loess deposits shortly after 4500 yr B.P.

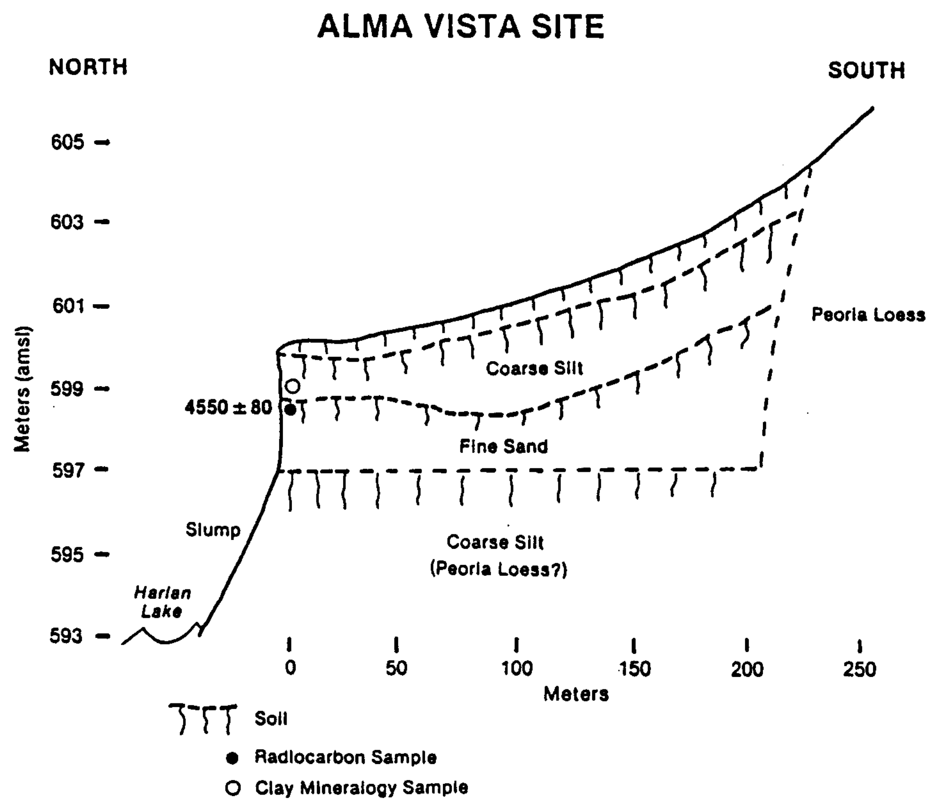

Figure 5:26 Cross-section of Alma Vista site. View is looking to the east (downstream). Height is given in meters above mean sea level (amsl). Note location of radiocarbon and clay mineral samples.

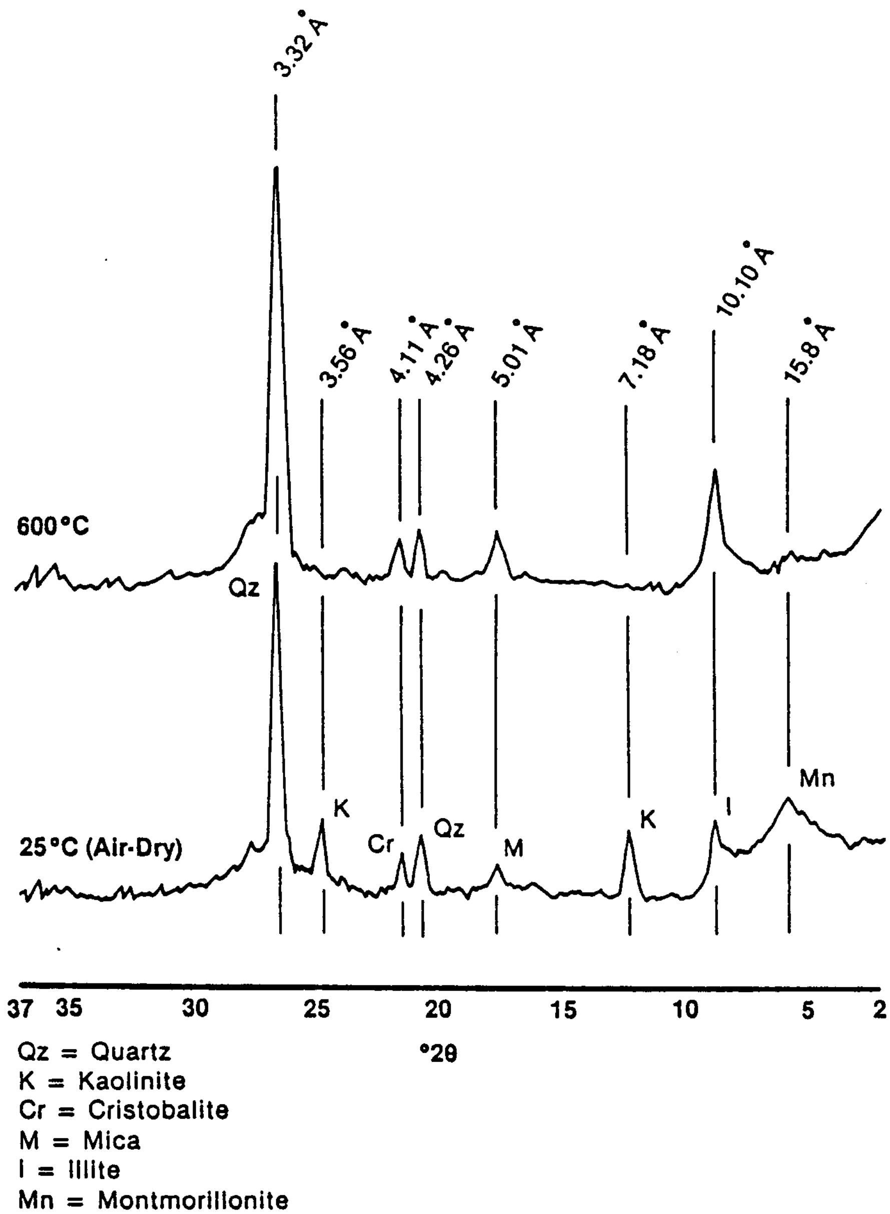

Figure 5:27 X-ray diffraction traces for clays (<2 microns) from sample collected at Alma Vista section, coarse silt overlying middle (4550 yr B.P.) paleosol (depth of 1.18 - 1.22 m). See Figure 5:26 for sample location. Material sampled was deposited shortly after 4550 yr B.P.. X-ray traces are for air-dired (25 $^{\circ}$ C) and oven-dried (600 $^{\circ}$ C) samples. Mineral assemblage, which is typical of the assemblage for the Peoria loess (Swineford and Frye 1951), suggests that the material is derived from upland Peoria loess deposits.

To summarize, the Alma Vista site contains deposits dating from at least the middle Holocene to the present, and possibly from the late Pleistocene to the present if silt recently uncovered at the base of the exposure is *in situ* Peoria loess or its alluvial equivalent. Assuming the basal sediments date to the late Pleistocene, then the lowermost paleosol may have formed contemporaneously with the paleosols dated at 10,300 yr B.P. at the North Cove and Prairie Dog Bay sites. Fluvial deposition of the overlying fine sand unit ended surface stability and soil formation, and formed the fill terrace in which the middle paleosol (4500 yr B.P.) developed. After 4500 yr B.P., local eolian reworking of Peoria loess resulted in burial of the middle paleosol beneath coarse silt. Following this deposition, surface stability returned, and the upper paleosol developed. It was then buried by a veneer of coarse silt (eolian?) in which a thin, weakly-developed surface soil has formed; the weak development of this soil suggests surface stability of, at most, only several hundred years (see Hallberg et al. 1978).

Freer

The Freer site, named after landowner Jack Freer, is located on the south side of the Republican River, about 2.5 km south of Oxford (Figure 5:28). As such, it is the westernmost site sampled in this research. Unlike the sites around Harlan Lake, the Freer site does not feature any natural exposures. Consequently, a transect was positioned across the floodplain to the uplands and five cores were extracted using a trailer-mounted Giddings drill rig (see Figure 5:28). Coring sites were located to ensure that at least one core was recovered from each of the three surfaces identified at the site. Two cores were extracted from the floodplain fill, two from the terrace fill, and one from the uplands. Detailed core descriptions are provided in Appendix A. Although coring was carried out on the south side of the Republican River, three surfaces having approximately the same heights above the river were also identified on the north side.

Core #3, which was collected from sediments in the middle of the floodplain, revealed fine sand and coarse silt overlying fine to medium sand (Figure 5:29). At depths below 2.5 m, the coarse nature of the fill prevented extraction of a core, making it necessary to use a continuous flight auger to explore to greater depths. Augering uncovered a thick fill of gray (10YR

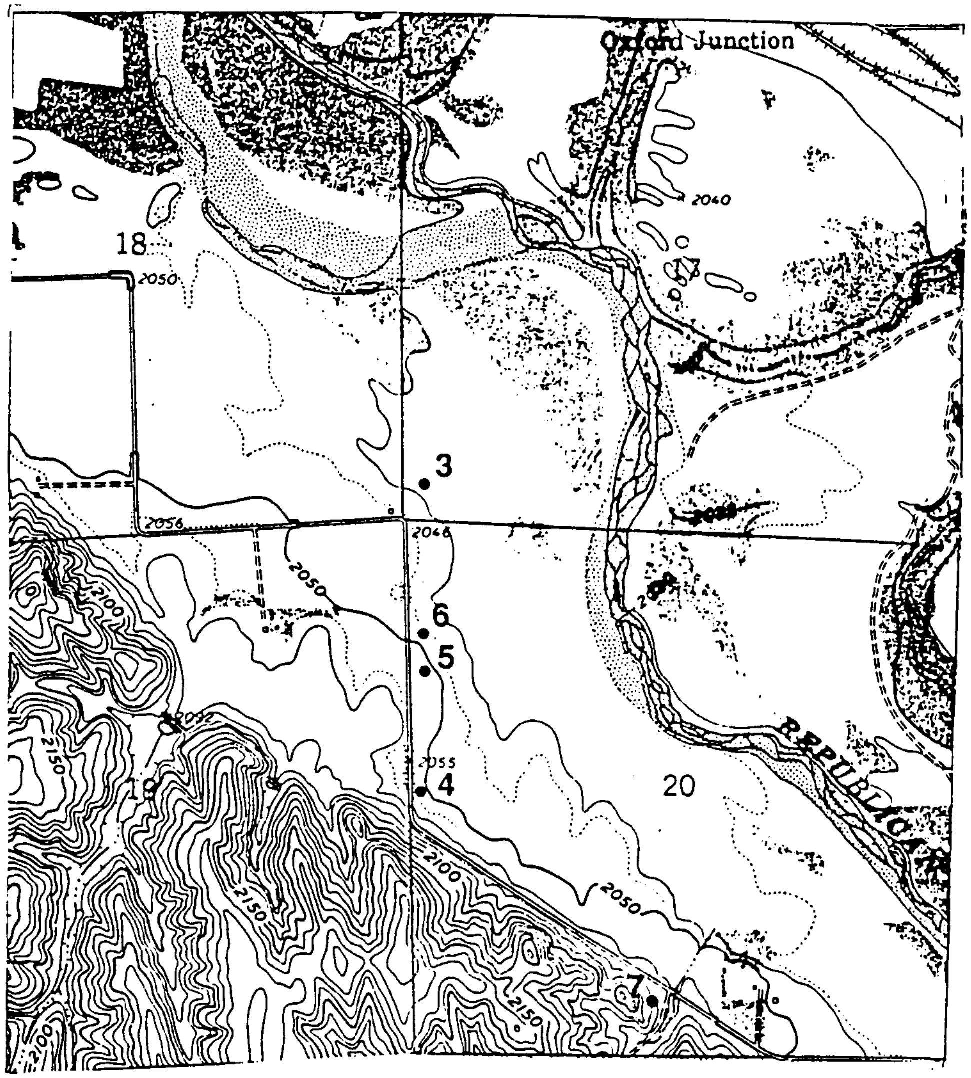

Figure 5:28 Section of topographic map showing location of core holes at Freer site. Freer site located approximately 50 river-km upriver from Harlan Lake. Core locations are as follows: #3 - SW 1/4, SW 1/4, Sec. 17. T3N, R20W; #4 - NW 1/4, SW 1/4, Sec. 20, T3N, R20W; #5 - SW 1/4, NW 1/4, Sec. 20, T3N, R20W; #6 - NW 1/4, NW 1/4, Sec. 20, T3N, R20W; #7 - SE 1/4, SW 1/4, Sec. 20, T3N, R20W Stamford, NE 7.5 minute quadrangle. Cores #3 and #6 were collected from the floodplain, Cores #4 and #5 from the terrace, and Core #7 from upland Peoria loess. Position of the present channel of the Republican River is visible in the NE 1/4 of Sec. 20, SW 1/4 of Sec. 17, and NE 1/4 of Sec. 18. Top of the map is north, contour interval is 10 feet. Direction of river flow is from the northwest to the southeast

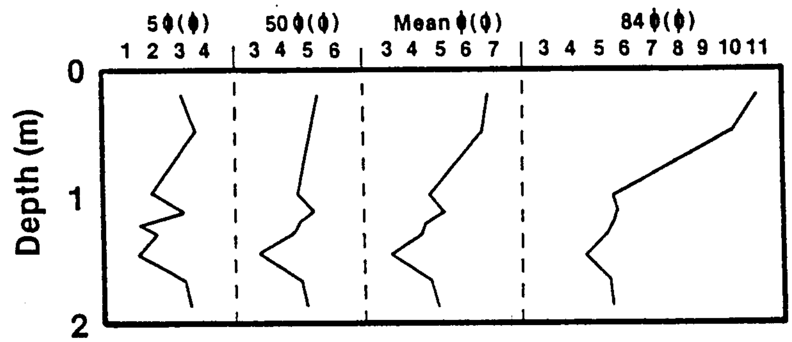

Figure 5:29 Particle size trends for Core #3, Freer site. Core was extracted from the floodplain (see Figure 5:28). Depth is measured from the top of the core. Height of floodplain surface is 623 m above mean sea level.

6/1) coarse sand between 2.5 and 7.1 m, and, at approximately 6.0 m, a layer (ca. 0.50 m) of grayish-brown (10YR 4/1), silty clay. No paleosols or organic-rich layers were noted in the core, an indication that deposition was continuous.

Core #6 was extracted from the floodplain at the base of the terrace scarp (Plate 5:13). Below a depth of approximately 1 m, the texture of sediment in Core #6 (Figure 5:30) is finer than that in Core #3 (see Figure 5:29); in Core #6, the phi mean ranges from 4.0 to 9.5, whereas in Core #3 the value is 3.0 to 4.75. The finer texture of sediment in Core #6 can likely be attributed to a more distal position on the floodplain.

The upper part of Core #6 is composed of fine sand and silty fine sand layers interbedded with continuous and discontinuous organic-rich clay bands; these clay-rich bands are illustrated in the trend of the eighty-fourth percentile line, particularly between 2.5 and 5.0 m (see Figure 5:30). This upper 5.0 m of fill probably represents vertically accreted floodplain deposits that accumulated during repeated overbank flooding. Between depths of 5.03 and 5.97 m, interbedded black (10YR 2/1), organic-rich silt and silty sand was encountered. This material overlies a fill of gleyed coarse sand and

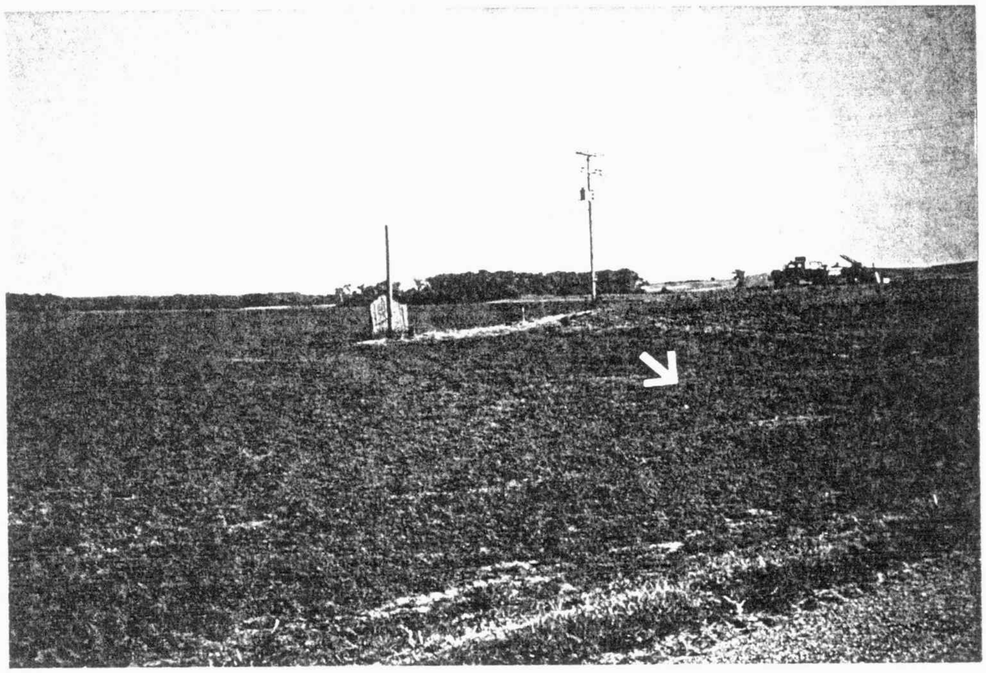

Plate 5:13 Freer site, location of Cores #6 and #5. Position of Core #6 is marked with arrow, and truck is parked at location of Core #5. Lower surface is modern floodplain, upper surface is fluvial fill terrace that is thought to have been deposited between ca. 4000 and post-2000 yr B.P., and then incised sometime after 2000 yr B.P. View is to the southeast from gravel road (see Figure 5:28). Height of floodplain is 623 m above mean sea level, height of terrace surface is 626 m above mean sea level.

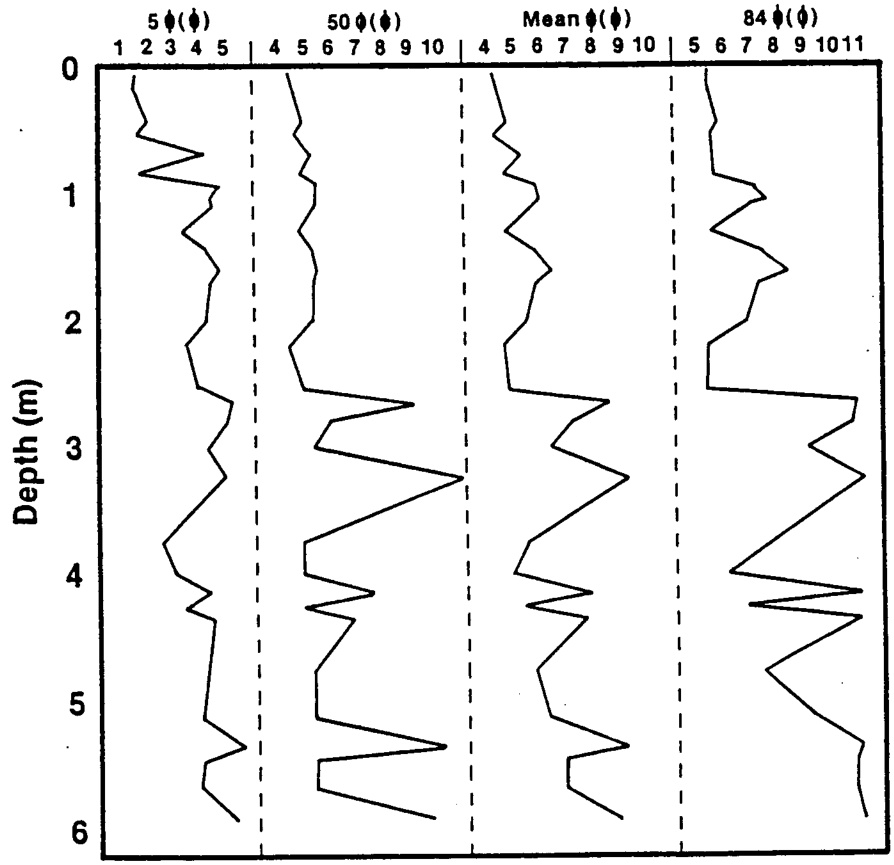

Figure 5:30 Particle size trends for Core #6, Freer site. Core was extracted from the floodplain at the base of the terrace scarp (see Figures 5:28 and 5:31). Depth is measured from the top of the core. Height of floodplain surface is 623 m above mean sea level.

gravel that appears to be an extension of fill noted in Core #3 at depths below 2.5 m. (see Appendix A).

The core data reveal that the floodplain at the Freer site comprises an upper unit of interbedded fine sand and silt that overlies a basal fill of coarse sand and gravel. The thickness of the upper unit increases with distance from the Republican River. At the margin of the floodplain (Core #6), this fill is approximately 5.0 m thick, and includes a series of organic-rich clay bands in the lower 3.5 m; closer to the river (Core #3), the fill is only 2.5 m thick, and the organic-rich clay bands constitute only the lower 0.20 m of the fill (Figure 5:31).

The subsurface data also indicate that the top of the basal coarse sand and gravel unit is at a greater depth at the margin of the floodplain (5.97 m) than it is in the middle of the floodplain (2.5 m). This variation may have been caused by river incision that occurred after deposition of the sand and gravel unit, but before deposition of the organic-rich bands. In the vicinity of Core #6, a topographic low, possibly an oxbow lake, served as a sink in which organics and clay accumulated during overbank flow, forming the organic silt and clayey silt bands in Core #6. These bands are

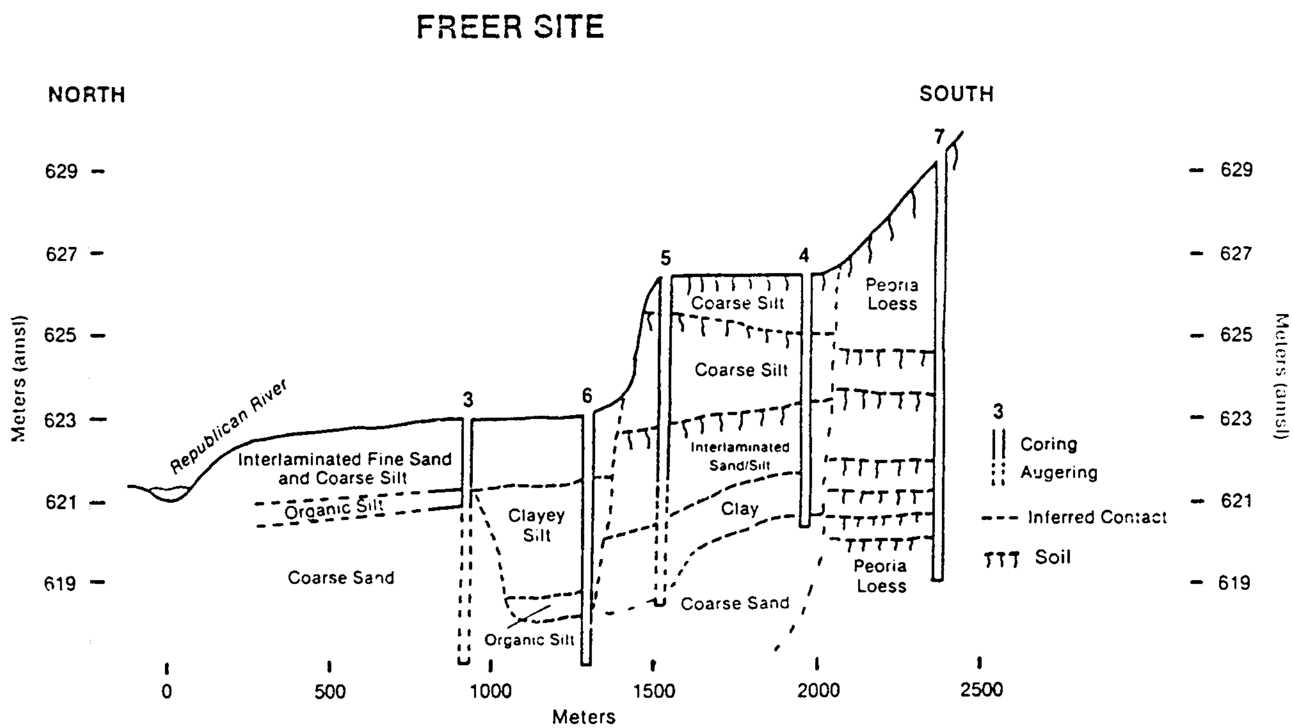

Figure 5:31 . Cross-section of Freer site and location of core holes. View is to the southeast (downstream), with north to the left of the cross-section and south to the right. Terrace fill of late Holocene age is inset against late Pleistocene Peoria loess. Fill of the modern floodplain is, in turn, inset against the Holocene fill. Height is given in meters above mean sea level (amsl). Depth of coring and augering for each hole are shown on cross-section.

overlain by coarser sediments deposited by the active channel of the Republican River as it meandered across the floodplain.

Cores #4 and #5 were extracted from the terrace whose surface stands about 3 m above the floodplain and 5.5 m above the Republican River (Plate 5:14); Core #5 was taken at the edge of the terrace scarp, Core #4 near the contact of the terrace surface and the slope leading to the uplands (see Figure 5:31). The homogeneous texture of the upper 3-to-4 m of terrace fill (Figures 5:32 and 5:33) stands in stark contrast to the heterogeneous character of the floodplain fill (see Figures 5:29 and 5:30). In Core #4, the mean phi value in the upper 3 m is approximately 5.0, while in the upper 4 m of Core #5 it ranges between 4.5 and 6; in contrast, the mean phi values of the floodplain fill are highly variable, ranging from 3 to 7 in Core #3, and from 4 to 9 in Core #6. At depths of 3.5 to 5.5 m in Core #4, and 4.36 to 4.70 in Core #5, the fill comprises interlaminated coarse silt and fine sand; in Core #4, the laminations are clearly visible in the phi size trends (see Figure 5:32). The upper 3-to-4 m of the fill consist of buff-to-brown (10YR 5/3 to 10YR 3/3), massive coarse silt. A lower paleosol featuring black

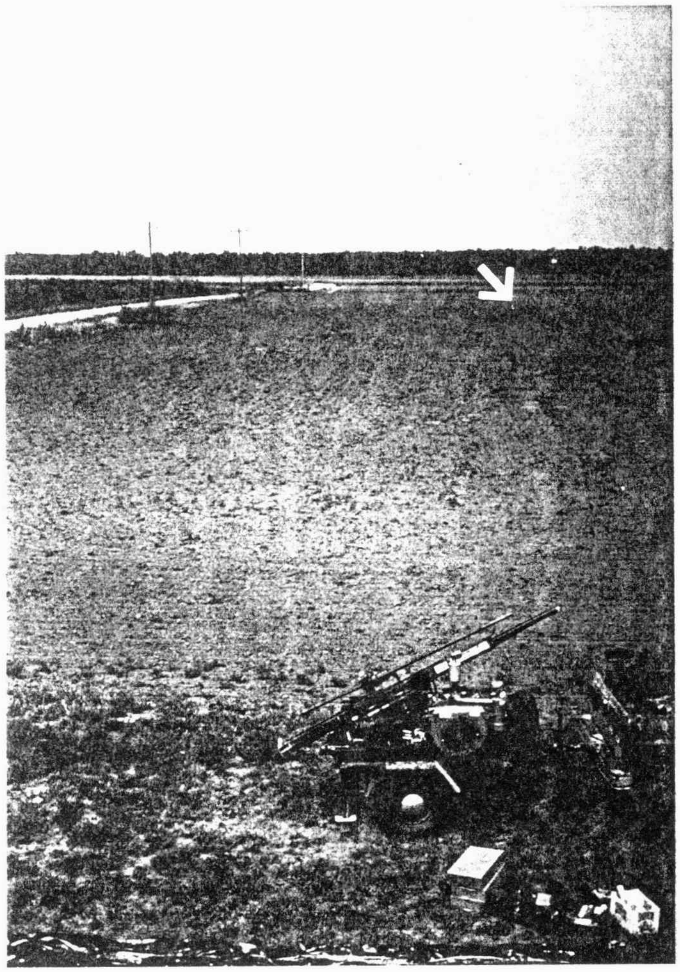

Plate 5:14 Freer site, looking from position of Core #4 across terrace to floodplain and Republican River (marked by trees). View is looking directly north. Drill rig is parked at location of Core #4, and arrow marks position of Core #5. Height of terrace surface is 626 m above mean sea level, and height of floodplain is 623 m above mean sea level (see Figure 5:31).

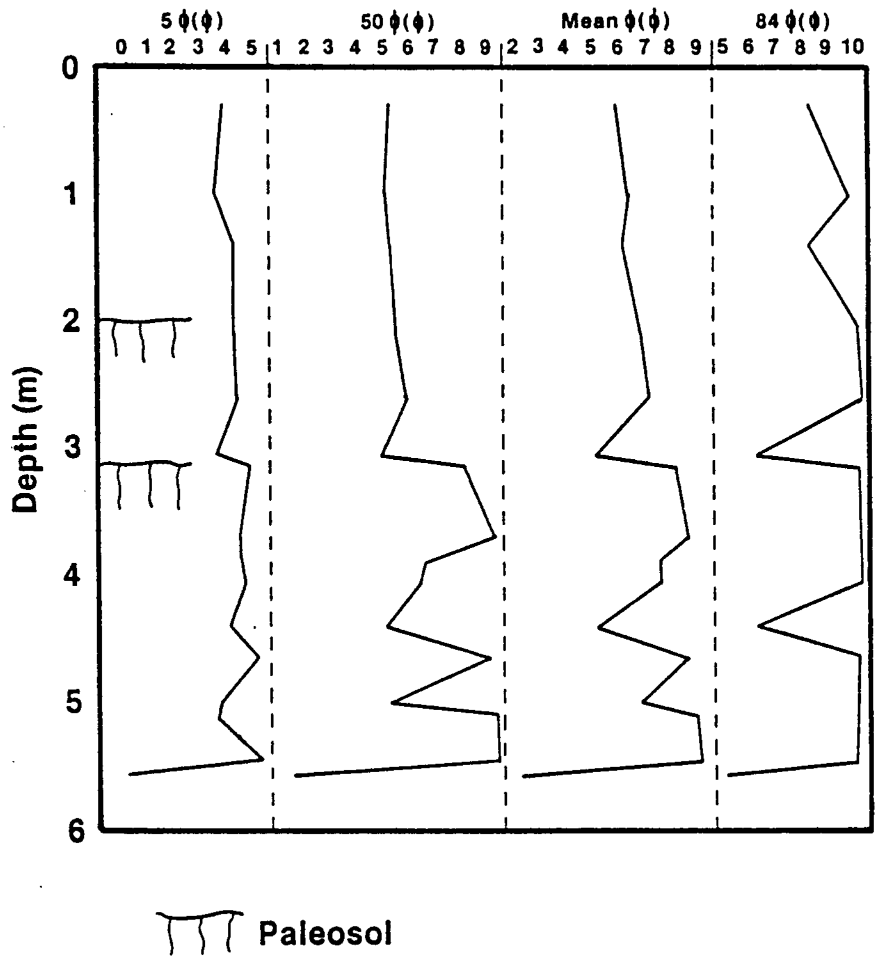

Figure 5:32 Particle size trends for Core #4, Freer site. Core was extracted from terrace where the surface intersects upland deposits (see Figures 5:28 and 5:31). Depth is measured from top of the core. Height of terrace is 626 m above mean sea level.

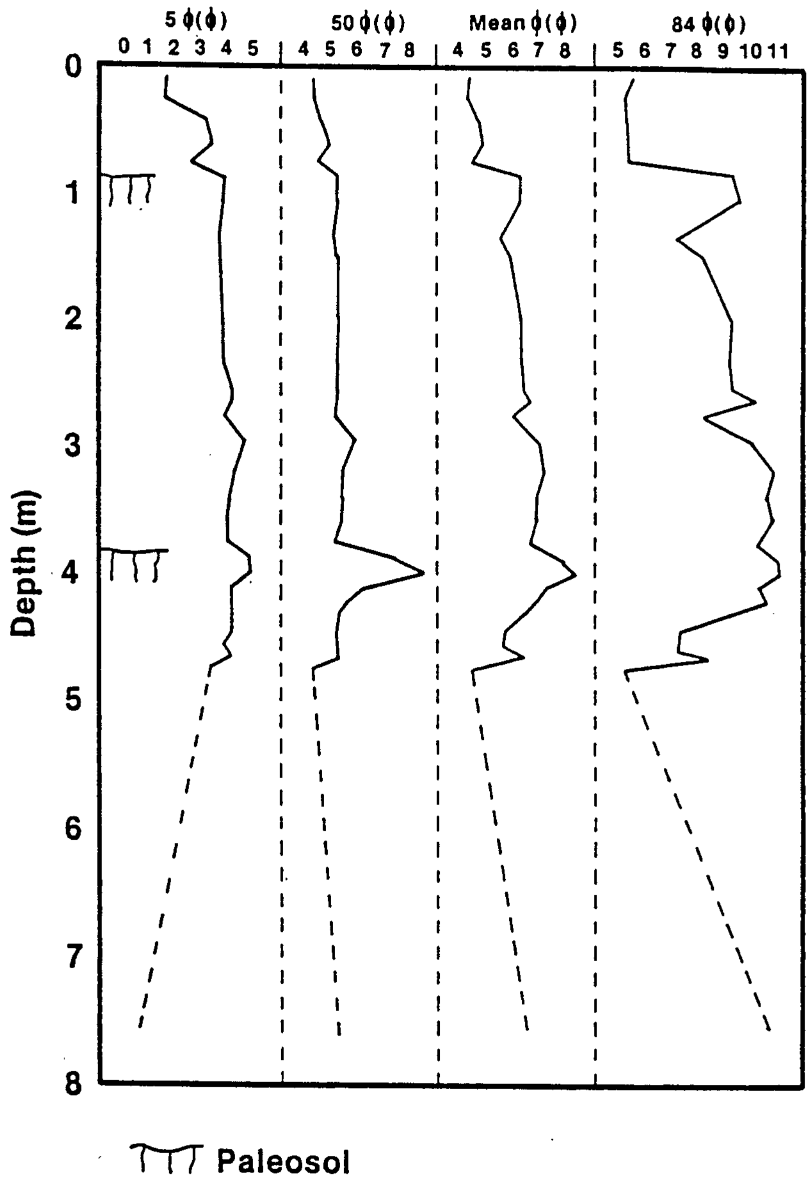

Figure 5:33 Particle size trends for Core #5, Freer site. Core was extracted from terrace surface near the terrace scarp (see Figures 5:28 and 5:31). Depth is measured from top of the core. Height of terrace is 626 m above mean sea level.

(10YR 3/1) Ab and Btb horizons is formed in buff (10YR 5/3) silt at depths of 3.14 to 3.78 m in Core #4 and 3.83 to 4.36 m in Core #5; the paleosol appears as a prominent clay bulge in the particle size distributions for Cores #4 and #5 (see Figures 5:32 and 5:33). An upper soil featuring black (10YR 3/1) Ab and Bwb horizons is present at a depth of 0.88 to 1.43 m in Core #5, and 2.00 to 2.24 m in Core #4. There is a clay bulge present in this soil in Core #5 (see Figure 5:33), but not in Core #4 (see Figure 5:32).

Clayey fill prevented coring below depths of 5.70 in Core #4 and 4.76 in Core #5. Continuous flight augering revealed gray (7.5R 4/0) clay at a depth of approximately 6.10 m in Core #5. This layer may overlie the gray (7.5R 4/0), coarse fluvial sand found in Core #4 at depths greater than 5.70 m. This sand appears to be an extension of the basal sand and gravel unit noted in Cores #3 and #6.

To determine whether fill of the terrace at the Freer site differs from fill beneath the upland surface, an additional core (Core #7) was collected from the adjacent uplands (see Figure 5:31). The upper 8.50 m of this core consists of buff-to-light brown (10YR 6/3-to-10YR 5/3), massive, homogeneous calcareous silt that displays little variation in the mean, fifth, fiftieth, and eighty-fourth phi values (Figure 5:34). Dispersed charcoal flecks were noted between 4.20 and 5.0 m, and brown (10YR 4/3), organic-rich zones (incipient A horizons?) at depths of 5.40 to 5.80, 6.95 to 7.11, 7.35 to 8.16, and 8.34 to 8.44 m. The matrix of charcoal and incipient soils is similar to that noted in in situ Peoria loess near Coyote Canyon on the southern shore of Harlan Lake. In contrast to the long-term stability denoted by paleosols developed in loess at the Bone Cove, Prairie Dog Bay, and North Cove sites, these incipient soils indicate relativity short-term interruptions in loess deposition.

As revealed by the eighty-fourth percentile line (see Figure 5:34), sediment texture is finer near the bottom of the core (between 8.80 and 11.0 m) than it is in the upper portion of the core; the change represents an increase in the fine clay component (clays < 0.5 microns) from values of 2% to 5% above 8.80 m to values of 10% to 18% between 8.80 and 11.0 m. This break is also reflected in the mean sorting values of the material above and below 8.80 m: from the surface to a depth of 8.80 m, the mean sorting value is 1.28

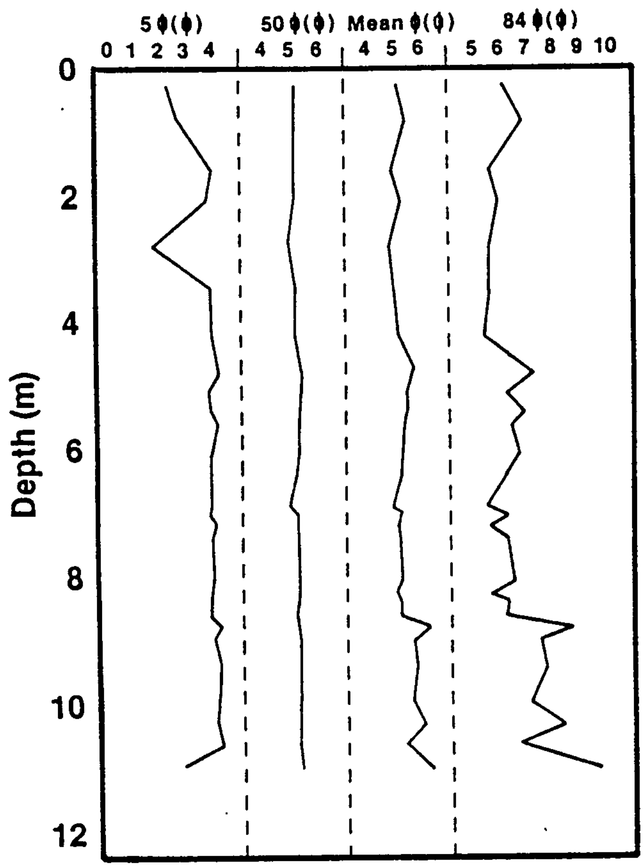

Figure 5:34 Particle size trends for Core #7, Freer site. Core was extracted from upland Peoria loess deposits (see Figures 5:28 and 5:31). Depth is measured from top of the core. Height of surface is 629 m above mean sea level.

(Standard Deviation = 0.24), whereas from depths of 8.80 to 11.0 m, the mean value is 1.99 (Standard Deviation = 0.32) (Figure 5:35). Low values of sorting correspond to a higher degree of sorting (see Appendix C). Note, too, that the degree of sorting for the loess is much greater than that for the terrrace fill. Owing to the small number of samples at depths of 8.80 to 11.0 m, however, it was not possible to determine whether the means are statistically different. The basal sediments in the core may represent a different depositional unit in the Peoria loess, or perhaps a zone of weathering. The absence from the fill of a strongly developed paleosol, such as was noted in the terrace fill, and the homogeneous nature of the deposit confirm that the fill from which Core #7 was extracted is different than that from which Cores #4 and #5 were taken. That is, the terrace is a fill terrace rather than a surface cut into the Peoria loess (i.e., a cut terrace).

To summarize, although radiometric ages on the floodplain and terrace fills have not been obtained, the vertical proximities of the floodplain (2.5 m) and terrace (5.5 m) to the Republican River imply that the surfaces and at least the upper portion of the fills are relatively young, probably dating to the middle and late

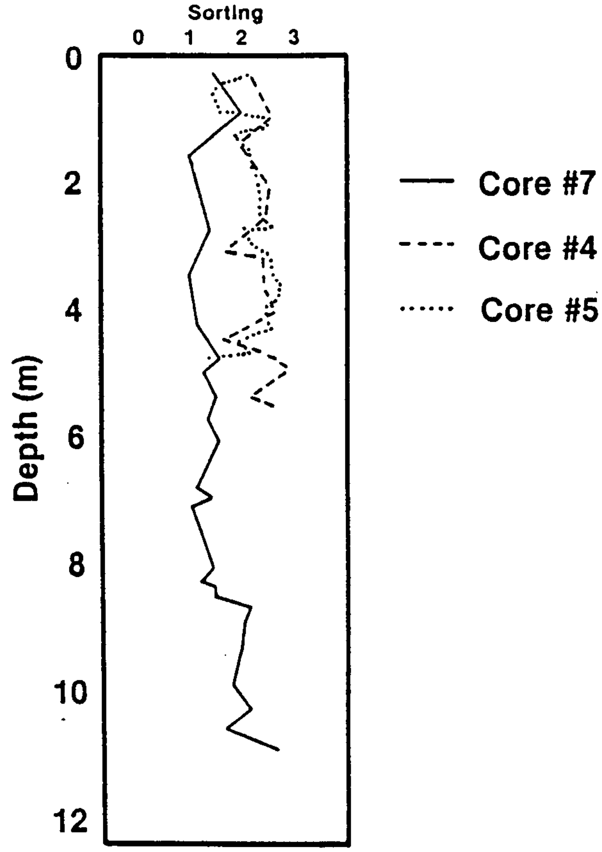

Figure 5:35 Sediment sorting values for Cores #4, #5, and #7, Freer site. Lower values of sorting correspond to a higher degree of sorting, i.e., better sorting (see Folk and Ward 1957; Appendix C). Graph reveals that sorting of loess (Core #7) is better than sorting of fluvially-deposited terrace fill (Cores #4 and #5). Core #7 was extracted from upland Peoria loess, Cores #4 and #5 from terrace fill (see Figures 5:28 and 5:31). Depth is measured from top of the core. Surface height for Core #7 is 629 m above mean sea level, and for Cores #4 and #5 is 626 m above mean sea level.

Holocene. As such, the Freer site affords a unique view of the continuum of deposits from the valley to the upland, and provides critical information about the stratigraphic relationship between Pleistocene and Holocene deposits in the study area. Further, it contains a valuable record of landform evolution at the late Pleistocene/early Holocene boundary.

Sometime during the late Pleistocene, perhaps contemporaneous with erosion that occurred at the Prairie Dog Bay and North Cove sites around 13,000 yr B.P., loess was removed from valley margins and upland loess deposits were truncated. Early and middle Holocene deposits were probably removed from the valley, although it is possible that the coarse sand and gravel underlying the floodplain and terrace date to this period. During the late Holocene, the clay layer at the base of the terrace fill was deposited in a low-energy environment, possibly in an oxbow lake at the margin of the floodplain. Following this, overbank deposits of interlaminated fine sand and coarse silt accumulated on the floodplain. Deposition then ceased and a lengthy period of stability was initiated, permitting formation of the Ab and Btb horizons of the lower paleosol. Following this period of floodplain stability, the paleosol was buried by fluvial deposits of massive coarse silt. Next, the upper paleosol formed during a relatively short interval of surface stability and pedogenesis. It was buried, in turn, by a veneer of colluvium or alluvium near the valley margin.

During the last 1000 years, this fill was truncated by river incision to create the terrace, and 2.5 m of laminated fine sand and coarse silt were deposited in the valley. Aside from scattered organic silts at the base, no organic accumulations were found in the most recently deposited fill, implying that floodplain accretion has been rapid. A weak cumulic A horizon has developed on the floodplain.

Schoenenburg

The Schoenenburg site, named after landowner Glenn Schoenenburg, is a cutbank exposure located 3 km south of Orleans (Figure 5:36). Nearly 7 m high, it features three paleosols that can be traced the length of the exposure, a distance of approximately 185 m (Plate 5:15). The paleosols are developed in and separated by deposits of silt, silty clay, and medium-to-fine sands. For a distance of 270 m back from the banktop, the surface of the fill is nearly level, but it rises

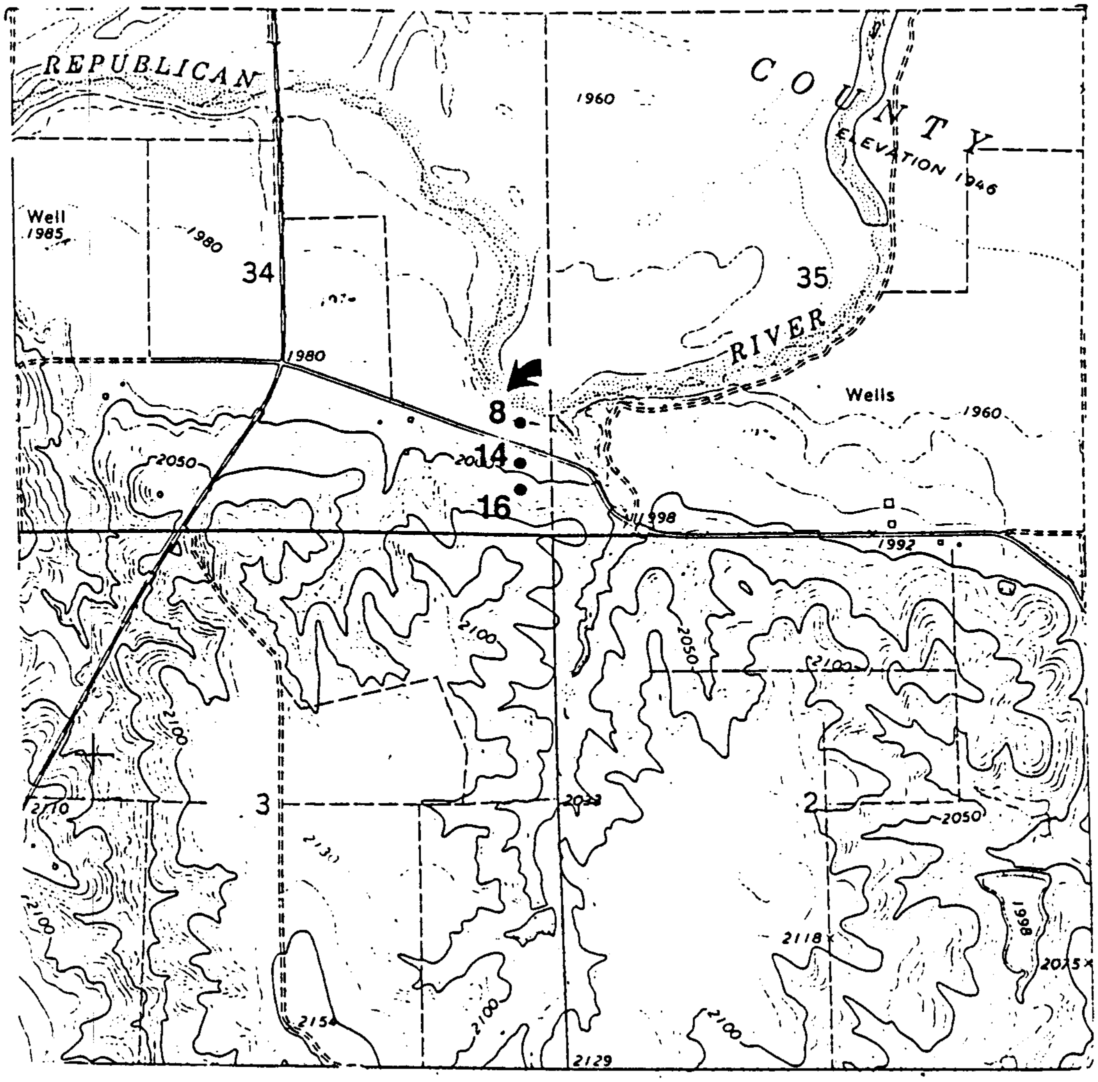

Figure 5:36 Section of topographic map showing location of Schoenenburg site cutbank and core holes. Schoenenburg site located approximately 15 river-km upriver from Harlan Lake. Cutbank position is marked with an arrow, and core holes with a dot. Exposure and core holes located in: SE 1/4, SE 1/4, Sec. 34, T2N, R19W Alma SW, NE 7.5 minute quadrangle. Top of the map is north, contour interval is 10 feet. Channel of Republican River is visible in sections 35 and 34. Direction of river flow is from west to east.

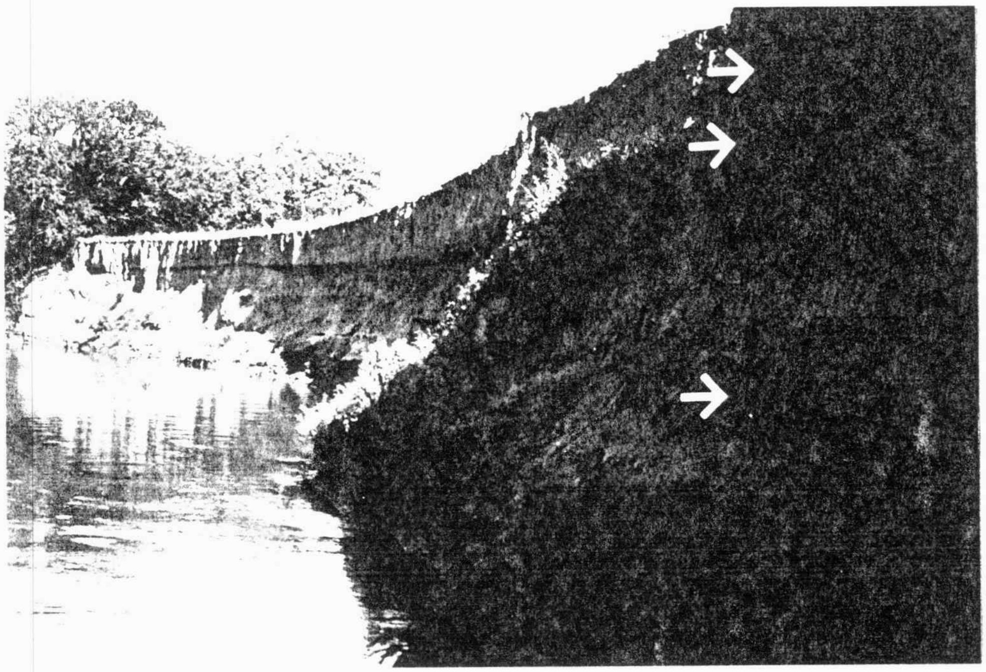

Plate 5:15 Cutbank at the Schoenenburg on the Republican River. Three paleosols developed in late Holocene fluvial deposits are marked with arrows. Radiocarbon ages on paleosols are as follows: Lowermost paleosol - (basal Ab) 3720 ± 90 yr B.P. (Tx-5977), (upper Ab) 3050 ± 60 yr B.P. (Tx-5912); Middle paleosol - (upper Ab) 2780 ± 80 yr B.P. (Tx-5978); Uppermost paleosol - (upper Ab) 2020 ± 60 yr B.P. (Tx-5911). View is to the southwest. Banktop height is 603 m above mean sea level. Banktop is approximately 7 m above level of the Republican River.

sharply where valley fill intersects upland loess.

Along the base of the exposure, there is a layer of cross-bedded, well-sorted medium fluvial sands (Figure 5:37). These coarse sands are capped by a 0.20 m-thick bed of fine sand that, in turn, is overlain by coarse sands. At a depth of 6.00 m, there is an abrupt change to silty fill, and at 5.80 m the deposit has a silty clay texture; these changes are reflected by the high eighty-fourth percentile phi values at a depth of 6.00 m (see Figure 5:37). The silty clay unit is gray (10YR 6/1) and highly calcareous (Figure 5:38). A well-developed paleosol featuring a black (10YR 3/1), organic-rich (OM = 1.9%), cumulic Ab horizon and dark brown (10YR 4/2) Bwb or Btb horizon is formed in the fill at depths of 4.87 to 5.57 m (Plate 5:16). The Ab horzon is clearly marked by high organic matter content and leaching of carbonates between 4.87 and 5.30 m (see Figure 5:38).

To establish a minimum age for the start of pedogenesis, a bulk sample was collected from the lower 0.16 m (5.41 - 5.47 m) of this soil. It yielded a C13-corrected radiocarbon age on humates of 3720 ± 90 yr B.P. (Tx-5977). A second sample, from the upper 0.15 m (4.96 - 5.02 m) of the soil, yielded a C13-corrected

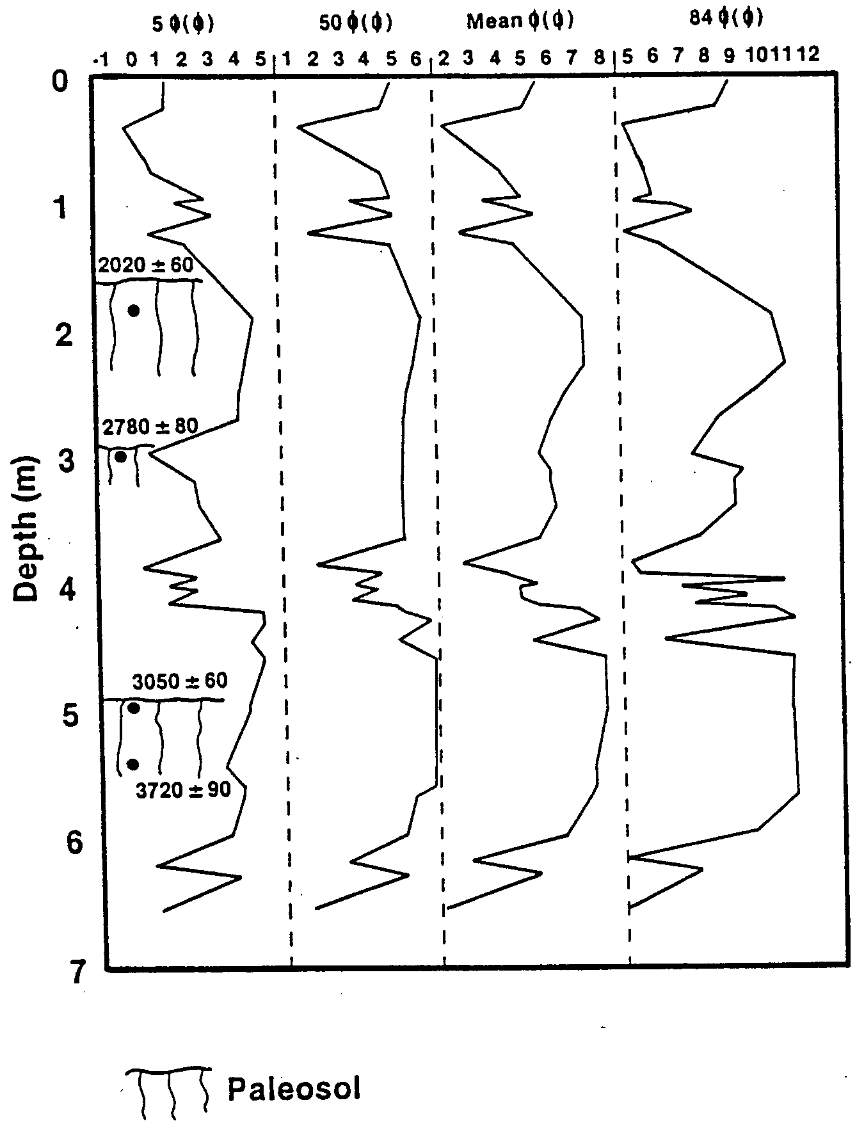

Figure 5:37 Particle size trends for Schoenenburg cutbank. Depth is measured from the banktop. Banktop height is 603 m above mean sea level (amsl).

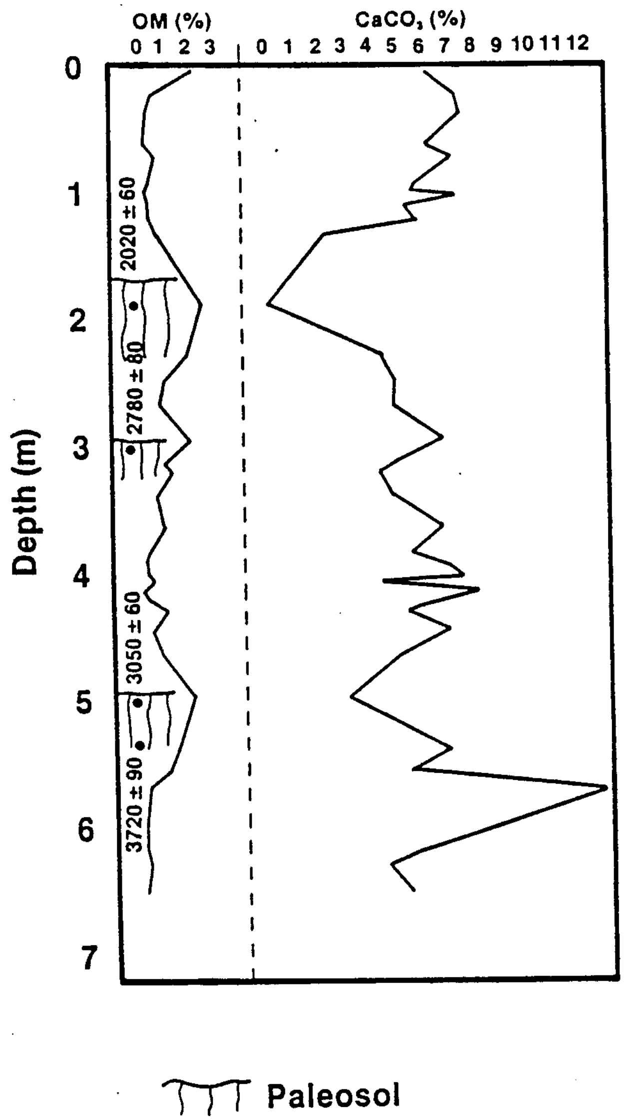

Figure 5:38 Organic matter and calcium carbonate trends for Schoenenburg cutbank. Note high organic matter and low calcium carbonate contents in the three paleosols. Depth is measured from the banktop. Banktop height is 603 m above mean sea level (amsl).

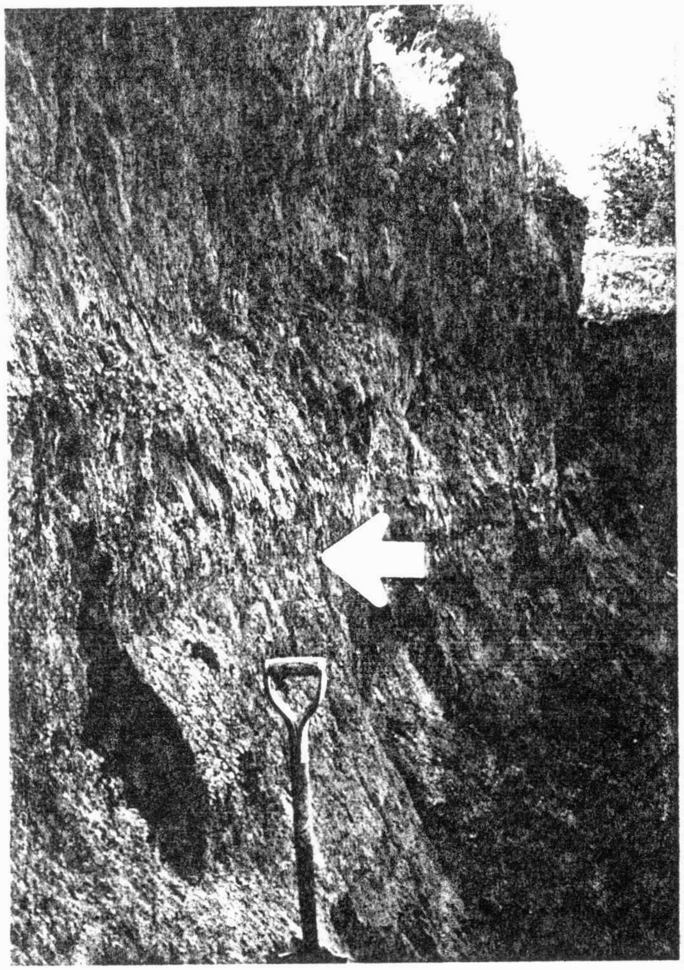

Plate 5:16 Lowermost paleosol at the Schoenenburg cutbank along the Republican River. Arrow marks the top of the Ab. Lower Ab of soil radiocarbon dated at 3720 $\pm$ 90 yr B.P. (Tx-5977), and upper Ab at 3050 $\pm$ 60 yr B.P. (Tx-5912). Soil is developed in fluvially-deposited silty clay. Length of shovel is 1.08 m. Middle and uppermost paleosols are visible in upper part of photo.

radiocarbon age on humates of 3050 ± 60 yr B.P. (Tx-5912). Assuming that the basal sample represents a minimum date for the start of pedogenesis (see Haas et al. 1986), then soil development had commenced by at least 3720 yr B.P; it continued until burial of the soil around 3050 yr B.P. Regardless of the exact beginning of pedogenesis, the two ages demonstrate that B horizon development in a floodplain environment can occur in as little as 1000 years; studies in Iowa have indicated that development of a Bt horizon can occur within 2500 years (e.g., Parsons et al. 1962). The ages also demonstrate that the Schoenenburg exposure comprises middle Holocene and younger fills.

The lowermost paleosol is buried by 2 m of overbank deposits. Grayish-brown (2.5Y 5/2) clay overlies the paleosol, and is in turn overlain by laminated clayey silt that is interbedded with clay drapes (see Figure 5:37). At a depth of 4.20 m, there is an abrupt change from clayey silt to fining upwards sequences of interbedded fine and medium sand, a change that may represent an increase in streamflow energy. It is in these overbank deposits that the middle paleosol is developed at a depth of 2.94 to 3.20 m (Plate 5:17). A sample from the upper 0.09 m (2.96 - 3.03 m) of this

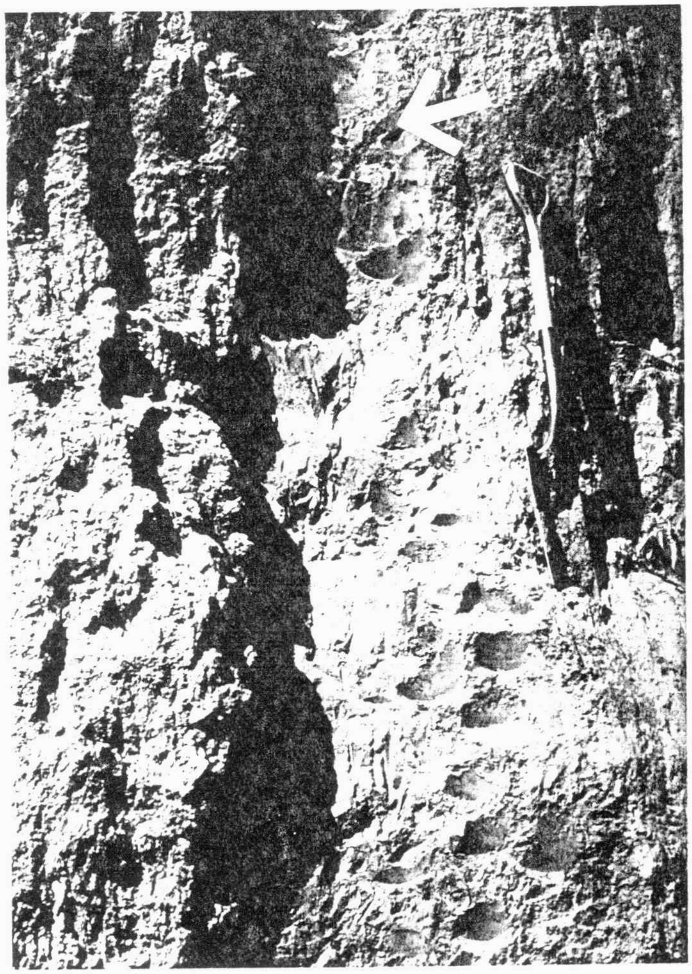

Plate 5:17 Middle paleosol at the Schoenenburg cutbank along the Republican River. Arrow marks the top of the Ab. Upper Ab of soil radiocarbon dated at 2780 $\pm$ 70 yr B.P. (Tx-5978). The radiocarbon age of 3050 yr B.P. on upper Ab of lowermost paleosol demonstrates that the fine and medium sand fill in which the middle paleosol is developed was deposited and the A horizon of the paleosol was formed in ca. 300 years. Length of shovel is 1.08 m. Lowermost paleosol is visible at the bottom of photo.

soil produced a C13-corrected radiocarbon age of 2780 ± 80 yr B.P. (Tx-5978). The paleosol consists of only a dark brown (10YR 3/3), cumulic Ab horizon that, in contrast to the lowermost soil, is not enriched in clays (see Figure 5:37); additionally, the organic matter content of the middle paleosol is lower (1.7%) than that of the lowermost soil (1.9%). Based on these two measures, it appears that the episode of surface stability during which the middle paleosol formed was shorter than the one during which the lowermost paleosol developed. The radiocarbon ages of 3050 yr B.P. for the burial of the lowermost paleosol and 2780 yr B.P. for the burial of the middle paleosol establish that floodplain aggradation and development of the A horizon (middle paleosol) took place in only 300 years.

Burial of the middle paleosol by massive and homogeneous coarse silt took place shortly after 2780 yr B.P. The overlying coarse silt unit displays little of the bedding and few of the laminations that characterize most other deposits at the Schoenenburg site (see Figure 5:37), suggesting that it may have accumulated in an environment of lower energy streamflow than that in which the other units were deposited. At depths of 1.63 to 2.50 m, the uppermost paleosol, which consists of a black (10YR 2/1) Ab and dark brown (10YR 3/3) clayey Bwb horizons, is developed in the coarse silt (Plate 5:18). The uppermost paleosol has the highest organic matter content (2.3%) of the three paleosols in the exposure, a condition that is likely due, in part, to the relative youth of the soil. A bulk sample collected from the upper 0.33 m (1.90 - 1.96) of the paleosol (the most organic-rich interval) yielded a C13-corrected radiocarbon age on humates of 2020 ± 60 yr B.P. (Tx-5911). The Ab horizon is leached of carbonates (see Figure 5:38), while the Bwb is relatively rich in carbonates and clays that were probably translocated from above (see Figure 5:37). Profile development of the uppermost paleosol, as measured by degree of horizonation and clay content, is greater than that of the middle paleosol, and roughly equal to that of the lowermost soil. I suggest that the period of surface stability during which the uppermost paleosol formed was longer in duration than the period during which the middle soil formed, and approximately equal in length to the period during which the lowermost paleosol developed.

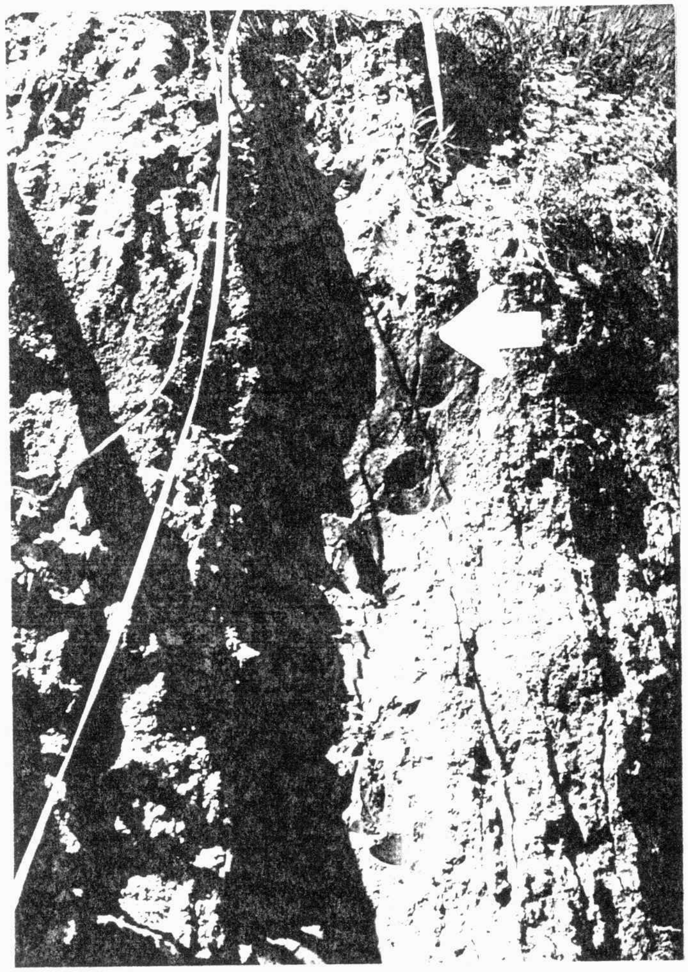

Plate 5:18 Uppermost paleosol at the Schoenenburg cutbank along the Republican River. Arrow marks the top of the Ab. Upper Ab of soil radiocarbon dated at 2020 ± 60 yr B.P. (Tx-5911). Uppermost paleosol is developed in fluvially-deposited coarse silt, and is buried by 1.6 m of laminated medium and fine sand (overbank deposits). Length of shovel is 1.08 m. Middle paleosol is visible in the lower part of photo.

This third episode of floodplain stability and pedogenesis ended after 2020 yr B.P. with truncation and deformation of the Ab to produce a wavy upper soil boundary. This deformation was followed by deposition of fining upwards sequences of medium and fine sand. Thirteen discrete sandy laminations, some only 0.02 m thick, were noted in the upper 1.60 m of the cutbank (see Figure 5:37). These laminations are postulated to have resulted from individual overbank flows. Following deposition of these fining upwards sequences, floodplain stability returned and a surface soil formed. Subsequent incision of 7 m has exposed the late Holocene deposits.

To trace the lateral extent of fills and paleosols to the south of the exposure, I extracted two cores from the terrace fill (Cores #8 and #14) and one from the valley side (Core #16) (Figure 5:39). Core #8, collected 28.5 m from the edge of the cutbank, contains three paleosols at depths closely approximating the depths of the three radiocarbon dated paleosols in the cutbank exposure. Core #14, which was extracted about 92 m from the banktop, contains two paleosols. In light of the variation in sediment texture that commonly exists across a floodplain, grain size parameters are of

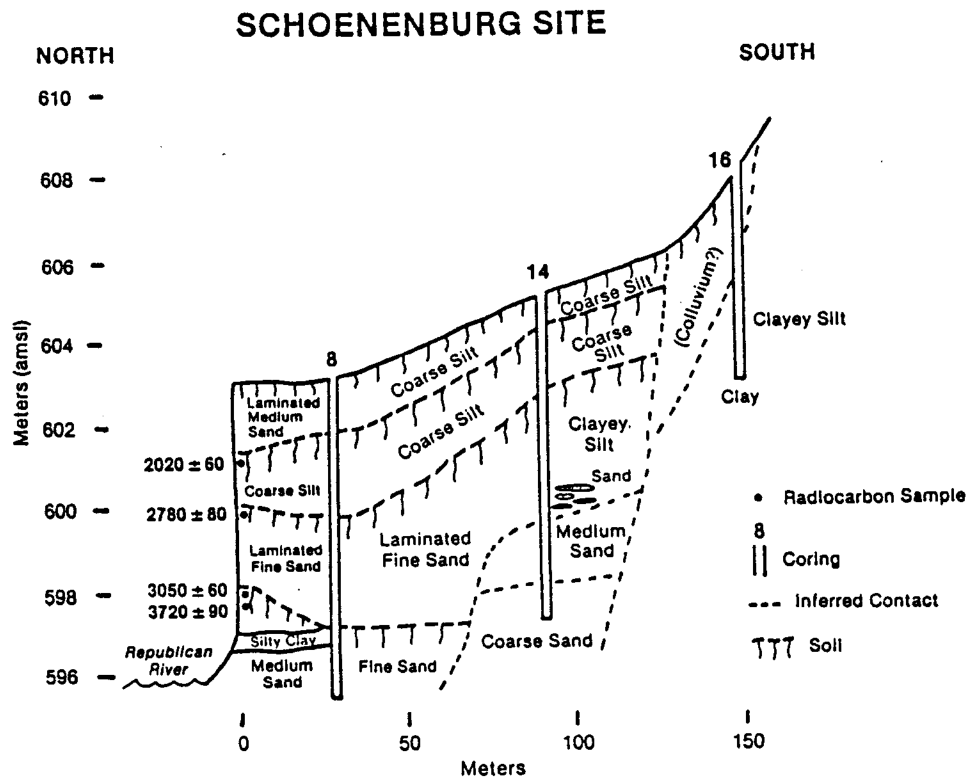

Figure 5:39 Cross-section of Schoenenburg site and location of core holes along Republican River. View is to the southeast (downstream), with north to the left and south to the right. Height is given in meters above mean sea level (amsl). Note that three paleosols are present in Core #8, but not in Core #14. Core #14 penetrated what may be late Pleistocene or early Holocene fluvial deposits that were removed from most of the Republican River valley, probably between 4500 and 3780 yr B.P. Late Holocene fluvial deposits are inset against older upland deposits (Peoria loess?). Depth of coring and augering for each hole are depicted on cross-section

little use in correlating soils across floodplain fills. Consequently, it was not possible to determine whether the soils that are present in the cores are the same soils that are present in the Schoenenburg exposure. To underscore this uncertainty, soils are depicted with dashed lines in the valley cross-section (see Figure 5:39).

Additional differences between the stratigraphy of Cores #8 and #14 and that of the exposure were also observed. First, the character of the sediments at a distance from the exposure is different than the character of the sediments in the exposure. Specifically, the cores lack the fining upward sequences that are so prevalent in the exposure between the lowermost and middle paleosols; this is illustrated in the phi value trends for Cores #8 and #14 (Figures 5:40 and 5:41), which display fewer oscillations than do lines for samples taken from approximately the same depths on the exposure (see Figure 5:37). Further, the standard deviation around the mean phi values decreases and the coefficient of variation (ratio of standard deviation to mean) increases with distance from the cutbank, reflecting a trend to more homogeneous sediment (Table 5:1). This trend can likely be attributed to the

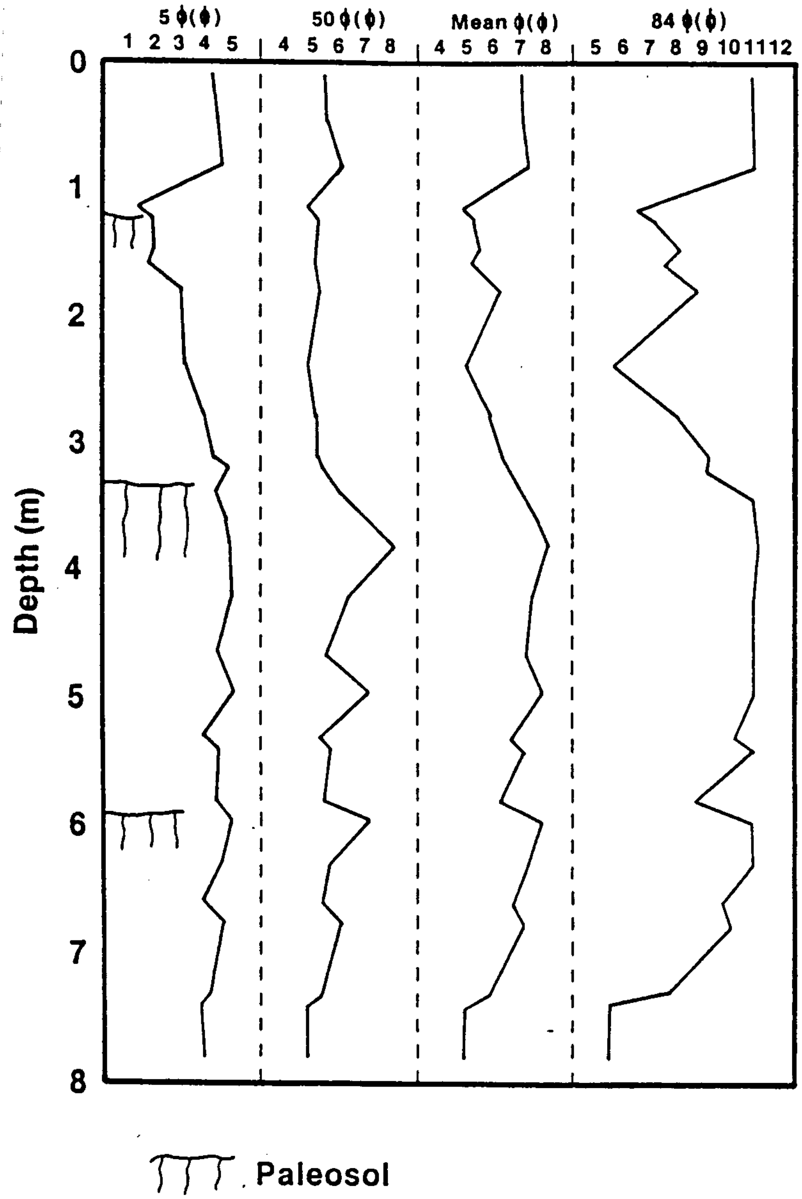

Figure 5:40 Particle size trends for Core #8, Schoenenburg site. Core was extracted from terrace fill a distance of 28.5 m back from the Schoenenburg cutbank (see Figures 5:36 and 5:39). Depth is measured from top of the core. Height of terrace is 603 m above mean sea level.

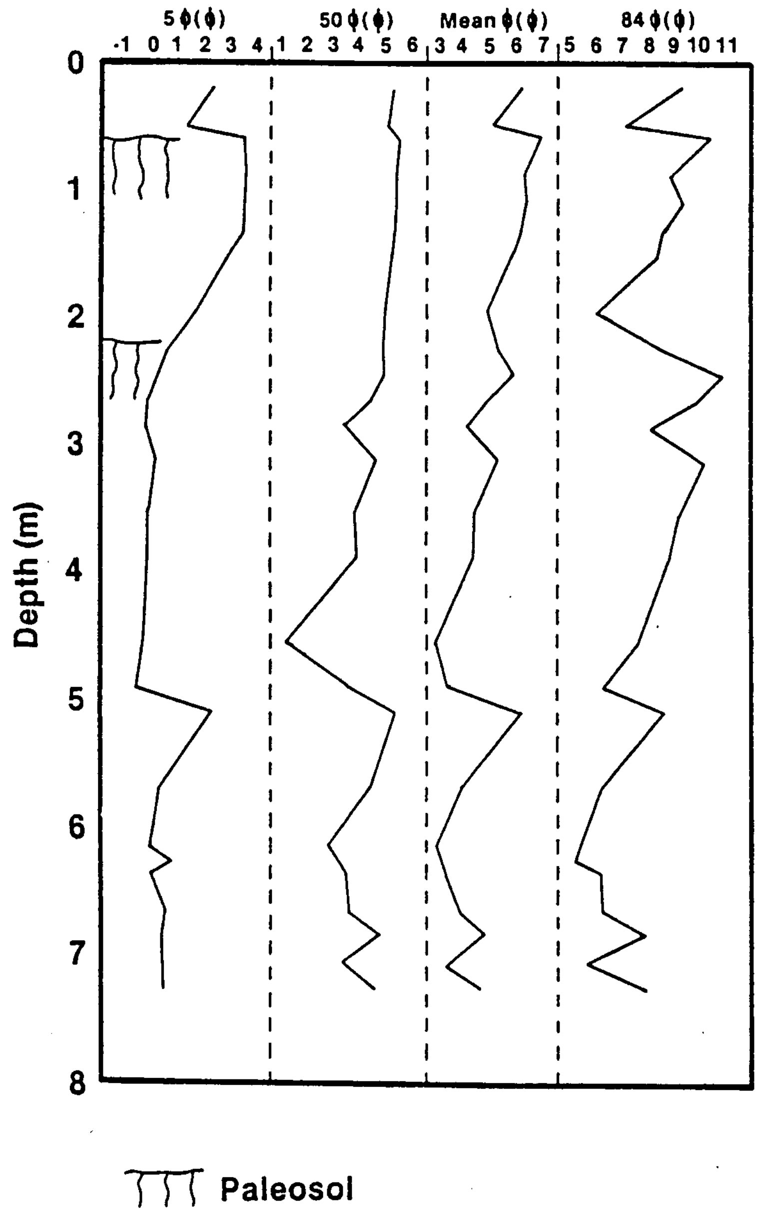

Figure 5:41 Particle size trends for Core #14, Schoenenburg site. Core was extracted from terrace fill a distance of 92 m back from the Schoenenburg cutbank (see Figures 5:36 and 5:39). Depth is measured from top of the core. Height of terrace is 604 m above mean sea level.

TABLE 5:1 : Schoenenburg Site - Sorting*

Site	N	Mean ($\overline{x}$)	St. Dev.	CRV
Cutbank	38	2.35	0.58	0.25
Core #8	28	2.25	0.36	0.16
Core #14	26	2.95	0.62	0.21
Core #16	22	2.95	0.30	0.10

CRV = Coefficient of Relative Variation
(Standard Deviation / Mean)

* The degree of sorting measures the dispersion of sediment sizes around the mean phi value (see Folk and Ward 1957). The higher the value of sorting, the worse the sorting of the sediment. For qualitative descriptions of the degree of sorting, see Appendix C.

mechanics of overland flow on the floodplain: coarser sediments settled out of suspension closer to the active channel, leaving only homogeneous sediment suspended in the overbank flow that reached the more distal portions of the floodplain (Allen 1970). In addition, the degree of sediment sorting decreases with distance from the channel. Although the mean sorting value of sediments in the exposure ($\overline{x}$ = 2.35) is similar to the value of sediments in Core #8 ($\overline{x}$ = 2.25), the value of mean sorting is substantially higher (which corresponds to poorer sorting) on the distal part of the terrace at Core #14 ($\overline{x}$ = 2.95) (Figure 5:42). This decrease in sorting may reflect inputs of poorly sorted colluvium to the alluvial fill on the distal part of the terrace.

A second difference between stratigraphy of the cores and that of the exposure is that the depth to the paleosols and the thickness of the fills decreases with distance from the exposure (see Figure 5:39). This observation is also consistent with theories of overbank flow and sediment deposition: highest rates of deposition, and therefore the thickest fills, occur near the active channel where sediment concentration in overbank flows is at a maximum (Allen 1970).

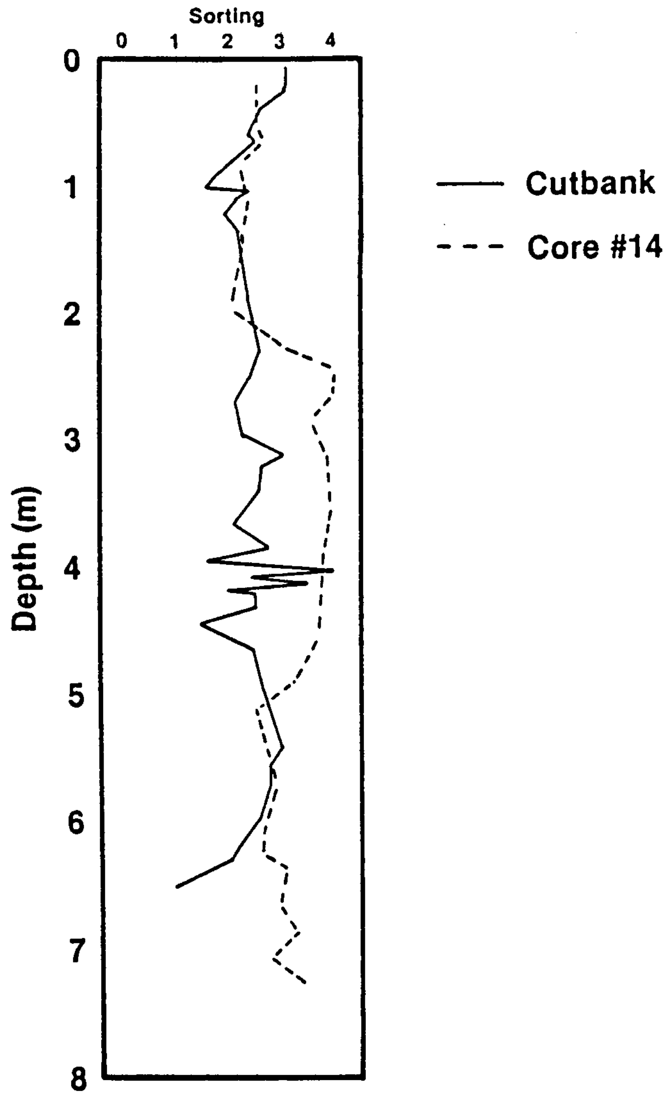

Figure 5:42 Sediment sorting values for Core #14 and cutbank, Schoenenburg site. Lower values of sorting correspond to a higher degree of sorting, i.e., better sorting (see Folk and Ward 1957; Appendix C). Core #14 was extracted 92 m back from the cutbank (see Figures 5:36 and 5:39). Graph reveals that the sorting of Core #14 is worse than the sorting of samples collected from the cutbank, possibly because there were inputs of colluvium to fill on the distal part of the terrace. Depth is measured from top of exposure and core. Height of banktop at cutbank is 603 m above mean sea level, and height of surface at Core #14 is 604 m above mean sea level.

Core #16, extracted from fill at the edge of the uplands about 160 m from the cutbank, contains no paleosols and features poorly sorted fill (see Figure 5:39) comprising interbedded sandy clay and sandy silt (Figure 5:43). Pebbles are also scattered throughout the material. The absence from the core of a paleosol demonstrates that this valley-side fill is distinct from the fill exposed in the cutbank, and the poor sorting suggests that it may not be in situ Peoria loess; indeed, an unpaired T-test revealed that the mean sorting of sediment in Core #16 ($\bar{x}$ = 2.95) is significantly different (0.05 level) than the mean sorting of in situ Peoria loess identified in Core #7 ($\bar{x}$ = 1.28) at the Freer site (see Figure 5:34). Because the fill is so poorly sorted, and because it rests along the margin of the valley, I propose that it is colluvium derived from the uplands. Because the fill is truncated by valley deposits, it must have accumulated prior to deposition of the basal fill of the Schoenenburg exposure, that is sometime before 3700 yr B.P. The colluvium overlies an oxidized and weathered yellowish-brown (2.5Y 6/4) clay layer. This clayey and oxidized unit may represent an eroded Bt horizon dating to at least the late Pleistocene.

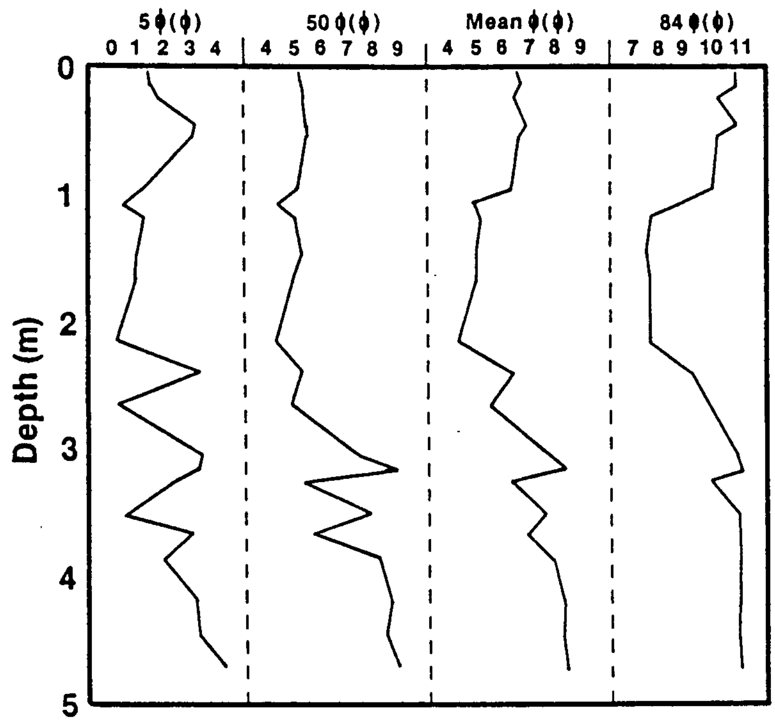

Figure 5:43 Particle size trends for Core #16, Schoenenburg site. Core was extracted from early or middle Holocene colluvium a distance of 160 m back from the Schoenenburg cutbank (see Figures 5:36 and 5:39). Depth is measured from top of the core. Height of terrace is 607 m above mean sea level.

To summarize, the Schoenenburg site, in addition to providing a radiocarbon-dated record of late Holocene fluvial activity in the study area, also affords a view of the stratigraphic relationship between valley and upland deposits (see Figure 5:39). River entrenchment occurred sometime before 3780 yr B.P., truncating colluvium that had extended down valley sides (Core #16 sediments). Then, the fill of coarse fluvial sands at the base of the exposure was deposited; these sands may correlate with basal coarse sediment noted below the floodplain and lower terrace fills at the Freer site. Floodplain stability and soil formation were initiated prior to 3780 yr B.P., and persisted until 3050 yr B.P. Between 3050 and 2780 yr B.P., deposition of 2 m of fill took place, followed by development of an A horizon. The radiocarbon ages on the lowermost paleosol demonstrate that 2 m of fill was depositied and an organic-rich A horizon formed in about 300 years.

Floodplain aggradation resumed around 2780 yr B.P., and terminated before 2020 yr B.P. Floodplain stability lasted until approximately 2020 yr B.P., at which time renewed overbank deposition first truncated part of the A horizon of the uppermost paleosol, and then buried the soil beneath fining upwards sequences of medium and fine sand. Contemporaneous with this floodplain accretion, colluvium derived from upland deposits of in situ Peoria loess accumulated at the margin of the floodplain adjacent to the valley wall. Following the episode of alluviation and colluviation, the Republican River entrenched 7 m to its pre-3780 yr B.P. level, exposing the sequence of late Holocene fills and paleosols preserved in the Schoenenburg exposure. Although the date of this incision in the study area is unknown, terminal radiocarbon ages on paleosols developed in fill of the lowermost terrace in other drainage basins of the central Great Plains indicate that it occurred shortly after 1200 yr B.P. (see Johnson and Martin 1987).

Although adjustment of the fluvial system is often complex, I maintain that variation in streamflow energy and frequency account for the abrupt changes in the character of sediment observed in the Schoenenburg exposure. Changes in sediment source area, while altering sediment size, would not accelerate the rate of floodplain construction, and therefore cannot account for the episodes of rapid deposition (e.g., between 3000 and 2780 yr B.P.) noted in the exposure. So too, movement of the active channel closer to the position of the cutbank would increase the rate of deposition, but would also result in deposition of sediment that is much coarser (medium or coarse sand) than that preserved in the Schoenenburg exposure. Finally, the episodes of floodplain aggradation identified here are generally contemporaneous with episodes of alluviation noted by May (1986,1989) in the Loup River system of central Nebraska. This regional synchrony argues against basin-specific controls (e.g., change in sediment source area, position of the active channel) as the cause of changes documented in the Schoenenburg stratigraphy, and in favor of regional climatic fluctutions. A similar hypothesis has been advocated by Knox (1983), but refuted by Trimble (1983, 1989); the latter has emphasized basin-specific intrinsic variables (e.g., amount of sediment in storage, conveyance capacity of streams) rather than climatic variables in explaining change in the fluvial system.

Unfortunately, there are no additional cutbank exposures of Schoenenburg terrace fill in the portion of the Republican River valley under study here. At the Freer site and in the valley between Orleans and Oxford, isolated remnants of a terrace stand about 5.5 m above the Republican River. Elsewhere in the valley, migration of the Republican River during the late-Holocene has removed most of this terrace. Assuming these remnants correlate with the Schoenenburg terrace, as I believe they do, then they were also deposited between 3780 and 2000 yr B.P.

Holocene Sedimentation

Based on analysis of deposits in the study area, the early and middle Holocene were dominated by deposition both on uplands (Bignell loess) and in the valley. On uplands, this deposition buried the equivalent of the Brady paleosol at the North Cove, Prairie Dog Bay, and Alma Vista sites. It ended sometime prior to 4500 yr B.P., at which time surface stability and soil development were initiated; the paleosol dated at 4500 yr B.P. at the Alma Vista site developed during this interval. This period of stability ended shortly after 4500 yr B.P. with deposition (eolian?) of coarse silt, followed by soil formation. Sometime between 4500 and 3780 yr B.P., an episode of incision occurred.

In the Republican River valley, the late Holocene has been characterized by alternating floodplain aggradation and surface stability. Following the episode of incision that took place between 4500 and 3780 yr B.P., a brief period of floodplain accretion ensued, and then development of the lowermost paleosol was initiated. Pedogenesis had begun by at least 3780 yr B.P., and ended shortly after 3050 yr B.P. Deposition of 2 m of fill followed by formation of the middle paleosol took place between 3050 and 2780 yr B.P. Floodplain aggradation resumed after 2780 yr B.P., burying the middle paleosol.

Development of the uppermost paleosol followed this period of floodplain alluviation. Soil formation persisted until shortly after 2020 yr B.P., at which time the soil was buried. Sometime during the last 2000 years, the Republican River incised about 7 m to its pre-3780 yr B.P. level. This incision formed the Schoenenburg terrace and the terrace at the Freer site. Since the incision, the river has deposited about 2.5 m of sediment that constitutes the modern floodplain.

CHAPTER 6: CHRONOLOGY OF LANDFORM EVOLUTION AND CONCLUSIONS

Chronology of Late Quaternary Landform Evolution

Integration of data collected from the six sample localities has yielded a chronology of landform evolution over the last 30,000 years along a reach of the Republican River in south central Nebraska that had been previously unstudied. Chronologies of this type are appearing with greater frequency in the literature (e.g., Knox 1983; Johnson and Martin 1987; Karlstrom 1988; Waters 1988; Blum and Valastro 1989; May 1989), and are helping to facilitate regional correlations among periods of erosion and sedimentation. Because such correlations are contingent upon radiocarbon control, however, it has proven difficult to reconcile older studies, most of which lack radiocarbon ages, with the emerging chronologies. Consequently, substantial revisions have been made to much of the earlier work in the central Great Plains and the classical late Quaternary stratigraphy that developed from it. This study joins the growing list of such revisions.

My model of landform evolution for the reach of the Republican River under study documents change on uplands and in the river valley. With one exception, I organize the chronology of landform change according to the same temporal categories that were used in earlier chapters; the exception is the early and middle Holocene, which are combined because landform change that was initiated in the former continued into the latter. Where possible, I speculate on the paleoclimatic conditions that may have triggered landform change. As a supplement to the text, diagrammatic stratigraphic sections from the six study localities are depicted and probable correlations among fills are suggested (Figure 6:1). All sections are tied to the longitudinal profile of the pre-dam Republican River floodplain as obtained from 7.5 minute, 1:24,000 United States Geological Survey topographic maps. In addition, Table 6:1 provides a summary of the geologic and environmental history of the study area during the last 30,000 years. Finally, to establish the degree of regional synchrony between this study and those completed elsewhere in the central Great Plains, I compare the activity that I document to episodes of erosion and deposition reported in studies from contiguous areas (i.e., eastern Colorado, central and southern Nebraska, and northern and northeastern Kansas).

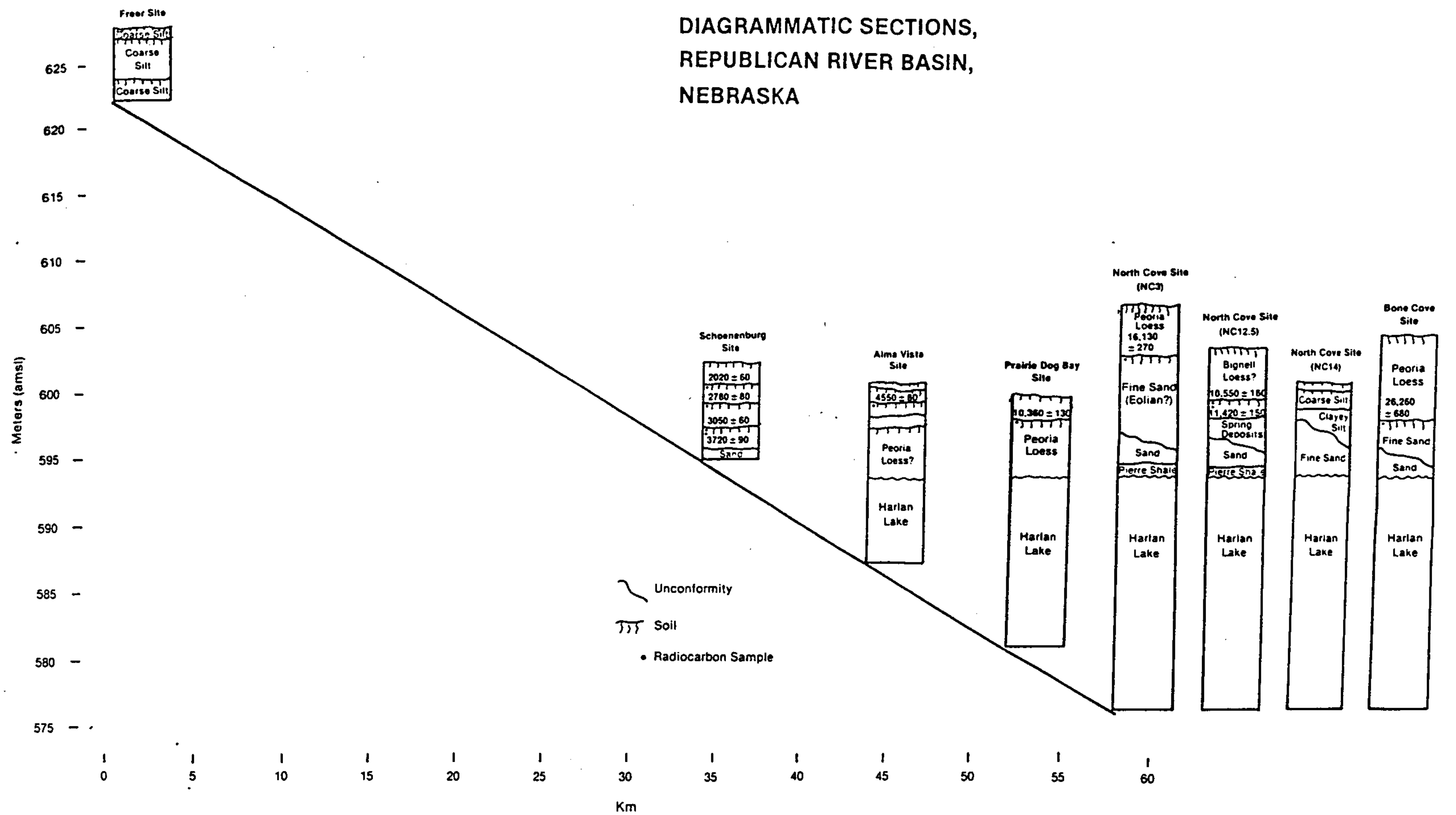

Figure 6:1 Diagrammatic stratigraphic sections for six study sites, Republican River basin, Nebraska. The height of all sections is tied to the longitudinal profile of the pre-dam Republican River. Note location of radiocarbon samples. For location of sites, see Figure 4:3. For a more detailed portrayal of individual sites, see the following figures: Freer site - Figures 5:28 and 5:31; Schoenenburg site - Figures 5:36 and 5:39; Alma Vista site - Figures 5:24 and 5:26; Prairie Dog Bay site - Figures 5:6 and 5:7; North Cove site - Figures 5:16 and 5:17; and Bone Cove site - Figures 5:1 and 5:2.

TABLE 6:1 : Geologic and Environmental Chronology for Study Site, 30,000 yr B.P. to Present

Time (yr B.P.)	Event	Environmental Conditions
Historical	Deposition	--
Post-1200	Incision (SC,FR)	?
1200-2000	Deposition (SC,FR)	Increased Flood Magnitudes
2000	Soil Formation (SC,FR)	Reduced Flood Magnitudes
2000-2700	Deposition (SC,FR)	Increased Flood Magnitudes
2700	Soil Formation (SC)	Reduced Flood Magnitudes
2700-3000	Deposition (SC,FR)	Increased Flood Magnitudes
3000-3700	Soil Formation (SC,FR)	Wet, Reduced Flood Magnitudes
Post-3700	Incision (AV,SC,FR)	More Effective Moisture
Pre-4500	Deposition (AV)	Shift to Less Effective Moisture
4500	Soil Formation (AV,PDB(?),NC(?))	Shift to More Effective Moisture
4500-10,200	Deposition (Reworked Loess) (AV,PDB,NC)	Shift to Less Effective Moisture
10,200-11,500	Soil Formation (Brady Equivalent?) (PDB,NC,AV(?))	More Effective Moisture

TABLE 6:1 : Geologic and Environmental Chronology for Study Site, 30,000 yr B.P. to Present

Time (yr B.P.)	Event	Environmental Conditions
11,500-13,000	Slow Deposition (PDB,NC)	More Effective Moisture
13,000	Incision (PDB,NC)	?
13,000-26,000	Deposition (Peoria Loess) (BC,NC,PDB,AV(?))	More Effective Moisture (Parkland(?))
26,000	Soil Formation (BC)	Less Effective Moisture
26,000-30,000	Deposition (Gilman Canyon Equivalent)	?

Site Locality Key (see Figure 4:3)

AV = Alma Vista Site
BC = Bone Cove Site
FR = Freer Site
NC = North Cove Site
PDB = Prairie Dog Bay Site
SC = Schoenenburg Site

Late Pleistocene (30,000 - 10,000 yr B.P.)

A yellowish-brown, fine sand unit was deposited in the study area and a soil developed in its upper portion prior to 26,260 yr B.P. This unit, which is exposed at the Bone Cove and North Cove sites, overlies medium-to-coarse, cross-bedded fluvial sands that were likely deposited during the Pleistocene by the ancestral Republican River (Johnson 1989). Although lighter in color than the Gilman Canyon Formation as described at its type section in Lincoln County, Nebraska (Reed and Dreeszen 1965), the age of pre-26,260 yr B.P. on this yellowish-brown fine sand unit falls within the 34,800 to 21,200 yr B.P. age range reported for the Gilman Canyon Formation elsewhere in Nebraska (Dreeszen 1970; May and Souders 1988; May 1989a). Furthermore, the position of the yellowish-brown unit directly beneath the Peoria loess matches the position of the Gilman Canyon Formation at its type section. Based on the radiocarbon age and the position of the fill, I postulate that the yellowish-brown fine sand unit identified at the Bone Cove and North Cove sites is the temporal equivalent of the Gilman Canyon Formation (see Figure 6:1).

As observed at the Bone Cove site, the episode of soil formation that was initiated in the upper portion of what I postulate to be the temporal equivalent of the Gilman Canyon Formation ceased around 26,260 yr B.P., and deposition of the Peoria loess began. This age for the start of the Peoria loess deposition is several thousand years older than the generally accepted age on basal Peoria loess reported elsewhere in the central Great Plains (e.g., Thompson and Bettis 1980; Ruhe 1983; L.D. Martin, Pers. Comm. 1989). At least some of this discrepancy may result from the level in the Bone Cove paleosol from which the aforementioned radiocarbon sample was collected. As shown in Plates 5:1 and 5:2, the paleosol has a diffuse upper Ab horizon characteristic of a cumulic soil. To determine the terminal age of an episode of pedogenesis, it is standard practice to sample the upper few cm of a paleosol. In this instance, however, to ensure that there was sufficient organic carbon for dating, the radiocarbon sample was collected from the most organic-rich level, which lies just below the diffuse upper part of the paleosol. Consequently, the radiocarbon age likely pre-dates the actual termination of pedogenesis. That is, deposition of the Peoria loess appears to have begun shortly after 26,260 yr B.P., but at a rate slow enough that pedogenesis could keep pace with it; this period of deposition is preserved as the diffuse upper portion of the Ab horizon. At some later time, deposition of loess accelerated to a rate faster than the rate of pedogenesis, a change that is marked by the top of the diffuse Ab horizon.

Three incipient Ab horizons noted in Peoria loess near Coyote Canyon indicate that deposition of the Peoria loess in the study area was episodic (see Figure 5:23 and Plate 5:8). A similar conclusion has been reached in other studies of loess deposits in the central Great Plains and adjacent areas (e.g., Daniels et al. 1960; Ruhe et al. 1971; McKay 1986). The presence of Picea charcoal in Peoria loess near Coyote Canyon, together with the identification of floral and faunal remains in loess deposits elsewhere in the vicinity of south-central Nebraska (e.g., Leonard 1952; Fredlund et al. 1985; Wells and Stewart 1987, 1987a), ha or parkland vegetation existed on uplands of the central Great Plains during deposition of the Peoria loess. This hypothesis is supported by evidence from modern loess depositional environments showing that loess accumulates only on well-vegetated surfaces (Yaalon and Dan 1974).

The climate conditions during deposition of the Peoria loess are unknown. Oxygen isotopes for opal phytoliths that were extracted from Peoria loess at the Eustis Pit, southwestern Nebraska, have implied that loess deposition occurred during conditions of more effective moisture, whereas soil formation took place under conditions of less effective moisture (Fredlund et al. 1985). The variation in effective moisture may reflect a change in temperature (and hence evaporation rates) rather than a change in precipitation amounts. Conversely, several computer models have postulated that cold, dry conditions persisted over the interior of North America during the late Pleistocene (e.g., Kutzbach and Guetter 1986). Whatever the climatic conditions, there appears to have been a tenuous balance between the rate of Peoria loess deposition and the growth of vegetation on the aggrading loess surface. Generally, deposition was slow enough to permit the establishment of vegetation, yet too rapid to allow the formation of thick mollic epipedons. High moisture conditions in topographic lows may have enhanced the preservation of organic matter and encouraged the formation of the incipient Ab horizons that have been observed in exposures of Peoria loess.

Deposition of the Peoria loess in the study area continued until just prior to 13,000 yr B.P., a time that falls within the generally accepted age range of 14,000 to 12,000 yr B.P. for the termination of Peoria loess deposition in the Great Plains (Thompson and Bettis 1980; Ruhe 1983). As observed at the Prairie Dog Bay and North Cove sites, deposition was interrupted by an episode of entrenchment just prior to 13,000 yr B.P. This incision was apparently restricted to topographical settings in close lateral and vertical proximity to the Republican River and its tributaries, as evidence for it is lacking at upland localities. Colluviation may have occurred during this time along the walls of the recently-cut valleys, but away from valleys upland deposits appear to have been little affected. The incision formed the valleys at Prairie Dog Bay in which the PDB1 and PDB3 fills were deposited, as well as the channel cut at the North Cove site in which the spring or streamside deposits accumulated. Elsewhere in Nebraska, entrenchment has been noted by Brice (1964) before 10,500 yr B.P. on the Loup River, Nebraska, and by May (1989a) around 13,000 yr B.P. on the middle Platte River, Nebraska.

The episode of incision around 13,000 yr B.P. may reflect a transition in the central Great Plains from the climatic conditions of the glacial maximum (ca. 18,000 yr B.P.) to those of the late Pleistocene. Recent climatic models suggest that with the accelerated wasting of the Laurentide ice sheet after 12,000 yr B.P., the glacial anticyclone over the western United States weakened. This, in turn, mitigated the strength of the winter polar jet (COHMAP Members 1988), and may have been sufficient to increase winter precipitation and amplify the magnitude of flooding. Additional evidence for wetter conditions during the late Pleistocene comes from the North Cove site, where spring activity, which indicates increased precipitation and a higher groundwater table, occurred between 13,000 and 11,420 yr B.P. The flora and fauna that lived in the vicinity of the spring are indicative of cooler and wetter conditions than those that exist in the region today (Stewart 1989).

The recently-incised valleys were filled sometime after 13,000 yr B.P. with late Peoria loess and reworked loess (i.e., the PDB1 and PDB3 fills). May (1989a) noted a similar episode of valley filling in the middle Platte River valley prior to 10,500 yr B.P. Loess deposition and valley filling ceased sometime between 13,000 yr B.P. and 11,500 yr B.P., and formation of the Brady paleosol equivalent was initiated. Depending on topographic position, the Brady is preserved either as a paleosol or as the surface soil. It is exposed as a paleosol at the Prairie Dog Bay (PDB1 and PDB3 fills), Alma Vista, and North Cove sites, and possibly as the surface soil at the Bone Cove site and in the upper portion of the PDB4 and PDB5 fills at the Prairie Dog Bay site. Radiocarbon ages obtained at the Prairie Dog Bay and North Cove sites demonstrate that development of the Brady equivalent had begun by at least 11,420 yr B.P., and continued until 10,550 to 10,270 yr B.P. This range of ages falls within the generally accepted dates of 12,700 to 8000 yr B.P. reported for the Brady paleosol in the central Great Plains by several researchers (e.g., Reed and Dreeszen 1965; Frye et al. 1968; Luttenegger 1985), but is older than an age of 9160 yr B.P. obtained on humates from the upper Ab of the Brady soil at its type section in southwestern Nebraska (Reed and Dreeszen 1965).

Early and Middle Holocene (10,000 - 4000 yr B.P.)

Deposition on uplands, valley sides, and floodplains occurred during the early Holocene from shortly after 10,300 yr B.P. to sometime before 4500 yr B.P., burying the Brady paleosol equivalent. I suggest that on uplands and valley sides, deposition was in the form of the Bignell loess, whereas on floodplains fluvial sand and fluvially-reworked loess were deposited. For example, at the Alma Vista site, the Brady paleosol equivalent is overlain by fluvial sand; conversely, at the North Cove and Prairie Dog Bay sites, which are stratigraphically higher than the Alma Vista site (see Figure 6:1), the paleosol is buried by Bignell loess and eolian coarse silt reworked locally from Peoria loess, respectively. Although deposition of the Bignell loess may have been widespread along the Republican River valley (Condra et al. 1947), this gray loess is present today only in topographic depressions on sheltered valley sides (e.g., North Cove). I surmise that it was blown-off of the exposed uplands (e.g., Bone Cove) and eroded from the valley floor at a later date. Deposits of Bignell loess ranging in thickness from 1-to-3 m have also been identified in contiguous north-central Kansas (Frye and Leonard 1949; Frye and Leonard 1952; W.C. Johnson Pers. Comm. 1989). In central Nebraska, May (1989) noted widespread alluviation between 10,200 and 4800 yr B.P. in the South Loup River valley, while elsewhere in the central Great Plains several studies documented alluviation during the early Holocene (e.g., Davis 1962; Brice 1964, 1966; Brakenridge 1980; Knox 1983; Johnson and Martin 1987).

Such strong synchrony among activity in the region implies a climatic control. Although the paleoenvironmental record of the early Holocene is limited, models of climatic conditions during the period indicate a trend towards warmer and drier conditions, apparently as a result of increased summer insolation values (COHMAP Members 1988) and stronger zonal atmospheric flow at the surface (Kutzbach and Guetter 1986; Kutzbach 1987). Assuming these reconstructions are accurate, then the sedimentation that occurred in the Republican River valley between 10,300 and pre-4500 yr B.P. appears to have resulted from a shift to dry conditions. Such a linkage between dry conditions and periods of aggradation has been noted by other workers (e.g., Huntington 1914; Love 1979; Brakenridge 1981, 1984; Knox 1983).

Prior to 4500 yr B.P., valley and upland sedimentation ceased and an episode of landform stability was initiated. It was during this period that the middle paleosol at the Alma Vista site formed; development of the surface soils at the Prairie Dog Bay and North Cove sites may also date to this time. Episodes of pedogenesis contemporaneous with this have been noted at several other locations in the central Great Plains (see Johnson and Martin 1987; May 1989), which suggests a climate control for soil formation.

Because the middle Holocene paleoclimatic record for the central Great Plains is limited (Fredlund and Jaumann 1987), little is known about the specific climatic fluctuations that prompted landform change during the period. In perhaps the most detailed study of paleoenvironmental conditions in the region during that time, Gruger (1973) cited the presence of prairie vegetation as evidence that dry conditions dominated the region until about 5000 yr B.P. After 5000 yr B.P., a shift to more mesic conditions led to the establishment of the modern-day mixed deciduous/prairie mosaic vegetation cover. In a study from the Nebraska Sand Hills, Sears (1961) documented an increase in arboreal pollen indicative of wetter conditions around 5000 yr B.P. More distant from the central Great Plains, Holliday (1989) reported that eolian sedimentation ended in northern Texas around 4500 yr B.P. In central Kansas, terminal ages of 6500 to 5000 yr B.P. on paleosols developed in dune sand also indicate an interval of mesic conditions around 5000 yr B.P. (W.C. Johnson Pers. Comm. 1989). It appears, then, that the interval of landscape stability and soil formation that persisted in the study area until 4500 yr B.P. can be linked to an increase in effective moisture. The onset of mesic conditions may reflect a shift from a zonal to a meridional pattern of upper atmospheric flow (COHMAP Members 1988).

Shortly after 4500 yr B.P., soil formation was terminated by renewed deposition. This deposition may have resulted from a relatively short-term drought. At the Alma Vista site, the middle paleosol was buried by a veneer of coarse silt that may have been the product of eolian reworking of Peoria loess deposits. Deposition ended with another episode of surface stability and soil formation, the date of which is unknown.

Late Holocene (4000 yr B.P. - Present)

The record of late Holocene fluvial activity in the study area features three paleosols that denote periods of floodplain stability (i.e., periods of non-aggradation) and several fills indicative of fluvial aggradation. When compared to the middle and early Holocene record, which depicts generally continuous aggradation in an increasingly arid environment, the record of the late Holocene suggests frequently changing climatic conditions. For example, the pre-4500 yr B.P. record at the Alma Vista site implies relatively slow, uninterrupted floodplain construction; conversely, the late Holocene record at the Freer and Schoenenburg sites documents up to four periods of aggradation interrupted by intervals of soil formation. Some researchers have attributed the variability in the late Holocene record to changes in flood magnitude and frequency (e.g., Brakenridge 1981; Knox et al. 1981; Haynes 1985; Knox 1985). Although reflected in the alluvial record of the central Great Plains, such changes are imperceptible in the floral and faunal records of the region (Fredlund and Jaumann 1987

As illustrated in Figure 6:1, the upper portion of the fluvial deposits in which the middle paleosol (4500 yr B.P.) at the Alma Vista site developed is slightly more than 10 m above the fluvial sand unit at the base of the Schoenenburg site. Therefore, at least 10 m of incision occurred between 4500 and 3700 yr B.P., probably accounting for the absence from the valley of most middle and early Holocene deposits. Valley fill inset against Peoria loess and located beneath floodplain sediment at the Freer site and at the base of the Schoenenburg exposure may be remnants of these deposits (see Figures 5:31 and 5:39). The entrenchment correlates with an episode of incision noted by May and Holen (1985) and May (1989) in the South Loup River valley, central Nebraska, between 4780 and 3030 yr B.P.

Following entrenchment and a brief episode of floodplain accretion, floodplain stability and soil formation occurred between 3720 and 3050 yr B.P. Floodplain aggradation resumed after 3050 yr B.P. with deposition of massive fine sand and coarse silt at the Freer site and laminated medium and fine sand at the Schoenenburg site. This episode of deposition ended prior to 2780 yr B.P. with renewed floodplain stability. About 2 m of fill accumulated between 3050 and 2780 yr B.P. at the Schoenenburg site, and nearly 2.5 m at the

Freer site.

The duration of floodplain stability around 2780 yr B.P. was relatively short. This conclusion is based on the following observations at and results of analyses from the Schoenenburg site: 1) Radiocarbon ages limit soil formation to the period 3050 to 2780 yr B.P., 2) Carbonates are present in the middle paleosol, suggesting that leaching did not occur, 3) The middle paleosol lacks a Btb, or even a Bwb, horizon, 4) The organic matter content of the middle paleosol (1.7%) is less than that of the lowermost paleosol (1.9%), and 5) The Ab horizon of the middle paleosol is thinner (0.26 m) than the Ab horizons of the uppermost (0.87 m) and lowermost (0.70 m) paleosols.

Floodplain accretion resumed after 2780 yr B.P., burying the middle paleosol. Similar to deposits lower in the section, the unit that overlies the middle paleosol has a coarse silt texture, but it lacks the laminations that characterize the units in which the middle and lowermost paleosols developed. The absence of laminations suggests that the upper fill was deposited in an environment of reduced streamflow energy. Deposition ended prior to 2020 yr B.P., and the uppermost paleosols at the Schoenenburg and Freer sites developed. Profile development of the uppermost paleosol, as measured by the thickness of the Ab horizon, clay content, and organic matter content, is greater than that of the middle paleosol, and roughly equal to that of the lowermost paleosol (see Figures 5:37 and 5:38). Shortly after 2020 yr B.P., renewed alluviation on the floodplain and colluviation at the margin of the floodplain ended soil development. The medium-to-fine sand texture of the valley fill that buries the uppermost paleosol at the Schoenenburg site may reflect an increase in flood frequency or magnitude.

Sometime after 2020 yr B.P., and following the period of alluviation and colluviation, there was a shift in fluvial activity from a regime characterized by alternating aggradation and non-aggradation to a period of incision. The Republican River entrenched its floodplain approximately 7 m at the Schoenenburg and Freer sites, reaching the level it had occupied prior to 3780 yr B.P. Research elsewhere in the central Great Plains has dated this entrenchment to shortly after 1200 yr B.P. (see Johnson and Martin 1987). As noted at the Freer site, valley filling of 2.5 m to form the modern floodplain has occurred since this entrenchment.

The record of late Holocene fluvial activity in the study area is roughly synchronous with observations from the Loup River system in central Nebraska. There, May and Holen (1985) and May (1989) documented incision between 4780 and 3030 yr B.P., and then a period of alluviation shortly thereafter. Before 2080 yr B.P., there was an episode of soil formation, and then entrenchment; this episode of entrenchment apparently did not occur in the Republican River valley as there is no evidence of downcutting at the Schoenenburg and Freer sites until sometime after 2020 yr B.P. Alluviation took place in the Loup River system from before 2080 until sometime after 900 yr B.P., at which time a short interval of incision was initiated; although this alluvation is contemporaneous with sedimentation that occurred in the Republican River valley after 2020 yr B.P., the episode of incision has not been noted in the study area. Deposition of sandy fill in the Loup River system after 900 yr B.P. was followed by incision of approximately 8 m. The latter may correlate to the entrenchment by the Republican River that occurred sometime during the last 2020 years.

Although regional synchrony among fluvial activity does not prove climatic forcing of the system, it does imply some climatic influence. Unfortunately, widely-used surrogate indicators of paleoclimatic change (e.g., micro- and macrobotanical fossils, landsnails, opal phytoliths) have been unable to detect Holocene climatic fluctuations in the central Great Plains, probably because such fluctuations resulted only in a change in grass species (Fredlund and Jaumann 1987). Consequently, only generalized reconstructions of Holocene paleoclimatic conditions, as gleaned from the alluvial record, have been presented here.

It is also possible that changes in fluvial activity during the late Holocene resulted from minor rather than major climatic change. That is, climatic change may have been of sufficient magnitude to affect fluvial activity, and yet too small to alter vegetational patterns. For example, studies of modern fluvial systems have revealed that rivers are sensitive to change in precipitation and flood frequency and magnitude (e.g., Schumm and Lichty 1963; Burkham 1972; Hereford 1984, 1986; Martin and Johnson 1987). But at the same time, such change had no effect on vegetational communities.

Contributions of This Study

This study has added to the emerging chronology of late Quaternary landform evolution in the central Great Plains. It also represents one of the few radiocarbon-controlled, late Quaternary chronologies in the region. Among the important findings are:

1) The basal Peoria loess may be older (ca. 26,200 yr B.P.) than the generally accepted age of ca. 21,000 yr B.P. (Thompson and Bettis 1980; Ruhe 1983).
2) Deposition of Peoria loess was interrupted by an episode of entrenchment around 13,000 yr B.P. Following this, Peoria loess deposition resumed, and continued until sometime just prior to 11,400 yr B.P.
3) The episode of soil development around 10,500 yr B.P. that has been noted elsewhere in the central Great Plains also occurred in the study area. This soil is tentatively designated the Brady paleosol equivalent.
4) The early Holocene was characterized by uninterrupted aggradation that persisted until sometime shortly before 4500 yr B.P. This alluviation is linked to a warm and dry climatic regime.
5) The late Holocene (post-4500 yr B.P.) has been characterized by alternating fluvial aggradation and floodplain stability, activity that stands in marked contrast to the relatively continuous aggradation of the early Holocene. Late Holocene activity appears to have resulted from frequent shifts in flood magnitude and frequency under relatively mesic climatic conditions.

6) Dark, organic-rich A horizons can develop on floodplains in 300 years or less, a finding that had been noted on the relatively mesic eastern periphery of the Great Plains in Iowa (Hallberg et al. 1978), but not in the more arid central Great Plains.

The record of late Holocene fluvial activity is especially intriguing in that it reveals the sensitivity of river systems to what were apparently short-term climatic fluctuations.

To a limited degree, this research also documents the linkage between upland and lowland geomorphic activity. Unfortunately, lowland activity during the late Pleistocene is unknown; deposits of that age appear to have been removed from or deeply buried in the valley. Consequently, the following discussion is limited to the Holocene.

My findings suggest that uplands have been relatively stable since the start of the Holocene about 10,000 years ago. As indicated by the stratigraphic position of early Holocene fluvial sand at the base of the Alma Vista site (see Figure 6:1), the Republican River aggraded to an elevation of up to 10 m above its present level during the early and middle Holocene. At the same time that this aggradation occurred, the

Bignell loess was deposited on valley sides and uplands. This loess, which may have been derived from the aggrading floodplain (Condra et al. 1947), is apparently restricted to sheltered topographic settings such as the channel cut at the North Cove site. At exposed upland sites such as Bone Cove, it was subsequently removed by erosion, or perhaps was never deposited because such sites were too far removed from the source area of the loess (i.e., the aggrading floodplain). Karlstrom (1988) has recently postulated a similar coupling between periods of valley-floor aggradation and eolian deposition on uplands in Arizona.

Landform stability returned to both valley floors and valley sides prior to 4500 yr B.P. Development of the middle paleosol at the Alma Vista site, and perhaps the present-day surface soils at the North Cove and Prairie Dog Bay sites, as well, began during this interval of stability. Sometime between 4500 yr B.P. and 3780 yr B.P., the Republican River entrenched at least 10 m, and then aggraded about 2 m. During this same period, coarse silt was reworked by wind from Peoria loess and deposited on valley sides at the Alma Vista site. Since the entrenchment, landform evolution in the valley has alternated between periods of non-aggradation (i.e., floodplain stability), which are denoted by paleosols, and periods of floodplain accretion, which are represented by alluvial fills; valley sides, however, appear to have been stable during the late Holocene. Flood magnitude, as interpreted from the texture of fills (see Starkel 1983; Knox 1985), was at a maximum after 2020 yr B.P., and at a minimum between 2780 and 2020 yr B.P. Following the deposition of approximately 2 m of fill after 2020 yr B.P., the Republican River entrenched at least 7 m, forming the terrace remnants at the Freer and Schoenenburg sites and at other localities in the valley between Oxford and Orleans. This entrenchment may have resulted from a shift to even more mesic climatic conditions. As noted previously, this episode of incision has been reported at drainage basins elsewhere in the central Great Plains (see Johnson and Martin 1987). The approximately 2.5 m of fill deposited since the entrenchment indicates that the fluvial system has returned to a regime characterized by floodplain aggradation.

Future Research

Much remains to be done in the central Great Plains in both valley and upland settings before a regional model of late Quaternary landform evolution can emerge. Some possible avenues of research include: 1) Examine the spatial extent of periods of landform stability by radiocarbon dating paleosols developed in colluvial deposits, and then comparing the ages to those obtained on paleosols in valley fills; 2) Establish the minimum age for the initiation of landform stability at upland localities (e.g., Bone Cove, Prairie Dog Bay, North Cove) by radiocarbon dating the lower A horizon of surface soils in upland deposits; 3) Facilitate the calculation of rates of Holocene floodplain aggradation by obtaining basal and close-interval radiocarbon ages from paleosols contained in Holocene alluvium; 4) Examine the problems inherent in the radiocarbon dating of soils (e.g., contamination, lab variability, variability among humate fractions), a critical need since paleosols constitute the most widespread source of organic material for radiocarbon assays in the Great Plains; and 5) Seek to clarify the relationship between climatic fluctuations and geomorphic processes by further documenting historical (post-European settlement) landform change, ideally in landscapes where historical human disturbance of biological and physical systems has been minimal and well-documented (e.g., Konza Prairie preserve, east central Kansas).

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APPENDIX A: FIELD AND LABORATORY DESCRIPTIONS OF SAMPLED EXPOSURES AND CORES

This appendix contains field and laboratory descriptions of exposures and cores, and the location from which samples were collected. All measurements are given as depths (in meters) below the banktop. The sample sites are as follows:

BC = Bone Cove Site (see Figures 5:1 and 5:2, Plates 5:1 and 5:2)
SW 1/4, NE 1/4, Sec. 20, T1N, R17W
Alma, NE/KS Quadrangle (1:24,000)
Banktop Elevation = 603 m (amsl)

PDB = Prairie Dog Bay Site (see Figures 5:6 and 5:7, Plate 5:3)
PDB1 = SE 1/4, SE 1/4, Sec. 23, T1N, R18W
PDB4 = SW 1/4, SE 1/4, Sec. 23, T1N, R18W
PDB3 = SW 1/4, SE 1/4, Sec. 23, T1N, R18W
PDB5 = SE 1/4, SW 1/4, Sec. 23, T1N, R18W
Alma, NE/KS Quadrangle (1:24,000)
Banktop Elevation = 598 m (amsl)

NC = North Cove Site (see Figures 5:16 and 5:17, Plate 5:6)
N 1/2, NE 1/4, Sec. 7, T1N, R17W
Alma, NE/KS Quadrangle (1:24,000)
Banktop Elevation = 605 m (amsl)

AV = Alma Vista Site (see Figures 5:24 and 5:26, Plate 5:10)
NE 1/4, SE 1/4, Sec. 5, T1N, R18W
Alma, NE/KS Quadrangle (1:24,000)
Banktop Elevation = 600 m (amsl)

Cores #3, #4, #5, #6, #7 = Freer site (see Figures 5:28 and 5:31, Plates 5:13 and 5:14)
Core #3 = SW 1/4, SW 1/4, Sec. 17, T3N, R20W.
Core #4 = NW 1/4, SW 1/4, Sec. 20, T3N, R20W.
Core #5 = SW 1/4, NW 1/4, Sec. 20, T3N, R20W.
Core #6 = NW 1/4, NW 1/4, Sec. 20, T3N, R20W.

Core #7 = SE 1/4, SW 1/4, Sec. 20, T3N, R20W.
Stamford, NE Quadrangle (1:24,000)
Surface Elevation (Cores #3, #6) = 623 m (amsl)
Surface Elevation (Cores #4, #5) = 626 m (amsl)
Surface Elevation (Core #7) = 629 m (amsl)

Schoenenburg = Schoenenburg Site Exposure (see Figures 5:36 and 5:39, Plate 5:15)
SE 1/4, SE 1/4, Sec. 34, T2N, R19W
Alma SW, NE Quadrangle (1:24,000)
Banktop Elevation = 603 m (amsl)

Cores #8, #14, #16 = Schoenenburg Site Cores (see Figures 5:36 and 5:39)
SE 1/4, SE 1/4, Sec. 34, T2N, R19W
Alma SW, NE Quadrangle (1:24,000)
Surface Elevation (Core #8) = 603 m (amsl)
Surface Elevation (Core #14) = 604 m (amsl)
Surface Elevation (Core #16) = 607 m (amsl)

DESCRIPTIONS

Bone Cove

Stratigraphy (m)		Samples (m)
BC5		
0 - 0.40	Surface soil, dark brown (10YR 3/3), coarse silt	
0.40 - 3.75	Buff (10YR 5/3), non-calcareous coarse silt	#0 (3.30 - 3.38)
3.75 - 4.70	Dark buff (10YR 4/4), non-calcareous coarse silt	#1 (3.85 - 4.00)
4.70 - 5.10	Brown (10YR 4/3), calcareous coarse silt	#2 (4.74 - 4.82) #3 (4.95 - 5.03)

Stratigraphy (m)	Samples (m)
5.10 - 6.15 Paleosol, dark gray (10YR 3/1) highly	#4 (5.20 - 5.27)
calcareous to 5.60	#5 (5.90 - 5.96)
6.15 - 6.94 Massive, buff (10YR 5/3) fine sand-to-coarse silt	#6 (6.57 - 6.63)
6.94 - 8.60 Yellowish-brown (10YR 6/4) fine sand	#7 (7.50 - 7.56)
8.60 - 9.15 Coarse, laminated sand	#8 (8.80 - 8.96)
9.15 + Slump	

Prairie Dog Bay

PDB1

0 - 0.75 Surface soil, dark brown (10YR 3/1), coarse silt	#7 (0.25 - 0.40)
0.75 - 1.40 Massive, buff (10YR 5/3) coarse silt	#6 (0.90 - 1.05)
1.40 - 2.10 Massive, buff (10YR 5/3) coarse silt, some CaCO3	#5 (1.55 - 1.70)
2.10 - 2.55 Paleosol, crumb structure, calcareous,	#4 (2.10 - 2.28)
dark brown (10YR 3/2)	#3 (2.43 - 2.55)
2.55 - 4.95 Massive, buff (10YR 5/3)	#2 (3.02 - 3.15)
silt, non-calcareous	#1 (3.85 - 4.15)
4.95 + Slump	

PDB4

0 - 0.73 Surface soil, dark brown (10YR 4/2) silt	#7 (0.25 - 0.32)
0.73 - 3.04 Buff (10YR 5/3) coarse	#6 (1.20 - 1.27)
silt, fossiliferous (snails)	#5 (1.53 - 1.60)
from 1.50 - 3.04	#4 (2.47 - 2.54)
3.04 - 5.90 Dark buff (10YR 4/4)	#3 (3.05 - 3.11)
coarse silt, fossiliferous	#2 (4.09 - 4.16)
(snails)	#1 (5.10 - 5.20)
5.90+ Slump	

Stratigraphy (m)		Samples (m)
PDB3		
0 - 1.30	Surface soil, dark brown (10YR 3/2)	#1 (0.70 - 0.79)
1.30 - 1.85	Buff (10YR 5/3) silt, slightly calcareous	#2 (1.56 - 1.63)
1.85 - 2.63	Brown (10YR 3/1) silt calcareous	#3 (2.23 - 2.34)
2.63 - 3.20	Paleosol, grayish-black (10YR 2/2), blocky structure	#4 (2.65 - 2.75) #5 (3.05 - 3.14)
3.20 - 5.05	Buff (10YR 5/4) silt non-calcareous	#6 (3.90 - 3.96)
PDB5		
0 - 0.70	Surface soil, black (10YR 3/1) silt	#7 (0.55 - 0.60)
0.70 - 2.10	Buff (10YR 5/2) coarse silt, fossiliferous	#6 (0.81 - 0.90) #5 (1.25 - 1.32) #4 (1.79 - 1.86)
2.10 - 5.20	Dark buff (10YR 4/4), coarse silt	#3 (2.20 - 2.28) #2 (2.90 - 3.00) #1 (4.00 - 4.10)
5.20+	Slump	
North Cove		
NC3		
0 - 0.70	Surface soil, dark brown (10YR 3/3)	
0.70 - 3.78	Buff (10YR 5/3), non-calcareous, soft coarse silt	#1 (2.75 - 2.80) #2 (3.40 - 3.47)
3.78 - 4.28	Paleosol, brown (10YR 4/4), calcareous	#3 (3.81 - 3.88) #4 (4.03 - 4.10)
4.28 - 5.34	Yellowish-brown (10YR 5/4), hard, calcareous fine sand	#5 (4.86 - 4.94)
5.34 +	Slump	

Stratigraphy (m)	Samples (m)
NC12.5	
0 - 0.75 Surface soil, black (10YR 3/1)	
0.75 - 3.20 Brown (10YR 4/3), calcareous coarse silt	#1 (2.05 - 2.12)
3.20 - 3.63 Dark brown (10YR 4/2), calcareous coarse silt	#2 (3.30 - 3.35) #3 (3.54 - 3.59)
3.63 - 4.50 Paleosol, dark brown (10YR 3/1), non-calcareous	#4 (3.80 - 3.84) #5 (4.25 - 4.30)
4.50 - 5.08 Gray (5Y 5/1), clayey silt	#6 (4.62 - 4.69)
5.08 - 5.22 Laminated, organic-rich clayey silt	#7 (5.08 - 5.13)
5.22 - 6.50 Gray (5Y 5/1) clayey-silt appears reduced	#8 (5.30 - 5.34)
6.50 - 6.70 Coarse-to-medium, cross-bedded fluvial sands	
6.70 + Slump	
NC14	
0 - 0.28 Dark brown (10YR 3/2), fine sand - surface soil	#1 (0.14 - 0.21)
0.28 - 0.53 Buff (10YR 5/4) fine sand	#2 (0.40 - 0.46)
0.53 - 1.12 Paleosol, dark brown (10YR 3/1)	#3 (0.70 - 0.80)
1.12 - 1.74 Brown (10YR 4/2), calcareous coarse silt	#4 (1.35 - 1.42)
1.74 - 2.04 Buff (10YR 5/3), calcareous coarse silt	#5 (1.96 - 2.02)
2.04 - 2.96 Gray (5Y 5/1), clayey-silt, appears reduced	#6 (2.41 - 2.48)
2.96 - 4.20 Yellowish-brown (10YR 5/4) fine sand-to-coarse silt	#7 (3.21 - 3.29) #8 (3.42 - 3.44) #9 (4.06 - 4.12)
4.20 + Slump	

Stratigraphy (m)		Samples (m)
Alma Vista		
AV1		
0 - 0.15	Surface soil, structureless dark brown (10 YR 3/3)	#9 (0.05 - 0.10)
0.15 - 0.25	Massive buff (10YR 5/3) silt	#8 (0.20 - 0.30)
0.25 - 1.18	Upper paleosol, non-calcareous (0.90 - 1.20), dark brown (10YR 3/1)	#7 (0.50 - 0.58) #6 (1.08 - 1.13)
1.18 - 1.50	Massive, buff (10YR 5/3) silt, calcareous	#5 (1.18 - 1.22)
1.50 - 2.00	Middle paleosol, non-calcareous, crumb and blocky structure, dark brown (10YR 3/3)	#4 (1.55 - 1.62) #3 (1.85 - 1.98)
2.00 - 2.80	Buff (10YR 5/3) fine sand	#2 (2.07 - 2.18) #1 (2.30 - 2.45)
2.80 +	Slump	

Freer Site		
Core #3		
0 - 0.55	Surface soil, cumulic A, dark brown (10YR 3/2), fine sand to sandy silt	#1 (0.25 - 0.30)
0.55 - 0.86	Buff (10YR 5/3) silt with organic flecks	#2 (0.63 - 0.68)
0.86 - 1.03	Buff (10YR 5/3) fine sand	#3 (0.93 - 0.98)
1.03 - 1.42	Buff (10YR 5/3) interlaminated coarse silt and fine sand	#4 (1.09 - 1.12) #5 (1.27 - 1.30) #6 (1.33 - 1.37)
1.42 - 1.53	Buff (10YR 5/3) fine sand	#7 (1.48 - 1.52)
1.53 - 1.80	Buff (10YR 5/2) sandy silt	#8 (1.69 - 1.74)
1.80 - 2.00	Laminated buff (10YR 5/2) sandy silt, with organic-rich laminae	#9 (1.82 - 1.87)
(Augering)		
ca. 2.5	Medium-to-coarse sand (10YR 6/1)	

Stratigraphy (m)		Samples (m)
ca. 3.0	Coarse sand with pebbles	
ca. 6.0	Gray (10YR 4/1) silty-clay, pebbles in clay	
ca. 7.1	Coarse, pebbly sand	

Core #6

0 - 0.14	Brown (10YR 6/4), fine sand - surface soil	#1 (0.04 - 0.08)
0.14 - 0.52	Buff (10YR 5/3), calcareous fine sand	#2 (0.17 - 0.22) #3 (0.45 - 0.50)
0.52 - 0.65	Dark brown (10YR 4/2), calcareous fine sand	#4 (0.56 - 0.60)
0.65 - 0.78	Buff (10YR 6/2), non-calcareous fine sand	#5 (0.71 - 0.75)
0.78 - 1.40	Laminated fine and medium sand, organic-rich laminae (1.02 - 1.05, 1.07 - 1.08, 1.31 - 1.33), oxidized (5YR 4/2) (0.94 - 1.01)	#6 (0.86 - 0.90) #7 (0.95 - 1.00) #8 (1.02 - 1.05) #9 (1.05 - 1.07) #10 (1.30 - 1.35)
1.40 - 1.60	Buff (10YR 4/3) fine sand	#11 (1.44 - 1.48)
1.60 - 1.67	Gray (10YR 3/3), oxidized clayey-silt	#12 (1.61 - 1.65)
1.67 - 2.40	Brown (10YR 4/2), interlaminted coarse silt and fine sand	#13 (1.69 - 1.73) #14 (2.00 - 2.03) #15 (2.20 - 2.24)
2.40 - 2.68	Buff (10YR 5/3), laminated, wavy fine sand with scattered flecks of organic debris	#16 (2.56 - 2.60)
2.68 - 2.80	Dark gray (10YR 3/1) clay, appears organic-rich	#17 (2.69 - 2.73)
2.80 - 2.88	Buff (10YR 5/3), laminated fine sand	#18 (2.80 - 2.83)
2.88 - 2.96	Clay laminae	
2.96 - 3.94	Grayish-black (10YR 3/1) silty and sandy clay, snails, more sand at 3.73	#19 (3.04 - 3.07) #20 (3.25 - 3.29) #21 (3.78 - 3.83)
3.94 - 4.07	Grayish-brown (10YR 5/2) interlaminated fine sand and organic-rich clay drapes	#22 (4.01 - 4.06)
4.07 - 4.20	Gray (10YR 3/1) clay snails	#23 (4.15 - 4.19)

Depth	Stratigraphy (m)	Samples (m)
4.20 - 5.03	Grayish-brown (10YR 5/2), interlaminated fine sand and organic-rich clay	#24 (4.23 - 4.27) #25 (4.36 - 4.39) #26 (4.74 - 4.77)
5.03 - 5.17	Black (10YR 2/1), organic-rich silty and clayey fine sand	#27 (5.11 - 5.14)
5.17 - 5.39	Grayish-brown (10YR 5/2) interlaminated fine sand and organic-rich clay	#28 (5.34 - 5.38)
5.39 - 5.56	Black (10YR 2/1), organic-rich silty and clayey fine sand	#29 (5.43 - 5.46)
5.56 - 5.60	Laminated medium sand (10YR 6/4), grading into gray clay at 5.60	
5.60 - 5.97	Gray (7.5Y 4/0) clay, organic flecks and lens of fine sand 5.73 - 5.75	#30 (5.65 - 5.68) #31 (5.90 - 5.93)
5.97 +	Coarse fluvial sand and pebbles	

Core #4

0 - 0.16	Surface soil, plow zone, brown (10YR 3/1) fine sand	
0.16 - 0.34	Surface soil, dark brown (10YR 2/1) fine sand	#1 (0.30 - 0.35)
0.34 - 0.90	Surface soil (Bw), dark brown (10YR 3/3) silt	
0.90 - 1.21	Buff (10YR 5/3), calcareous coarse silt	#2 (1.00 - 1.10)
1.21 - 2.00	Massive, non-calcareous buff (10YR 5/3) coarse silt	#3 (1.40 - 1.45)
2.00 - 2.24	Dark buff (10YR 4/3) calcareous coarse silt, likely weakly developed paleosol (eroded Bw?)	#4 (2.10 - 2.16)
2.24 - 2.94	Buff (10YR 5/3) coarse silt	#5 (2.60 - 2.65)
2.94 - 3.14	Buff (10YR 5/3) coarse silt-to-fine sand	#6 (3.05 - 3.10)

Stratigraphy (m)		Samples (m)
3.14 - 3.40	Ab, black (10YR 3/1) clayey-silt, columnar structure	#7 (3.16 - 3.20)
3.40 - 3.78	Bwb, dark brown (10YR 5/2) clayey-silt, carbonate nodules	#8 (3.72 - 3.76)
3.78 - 4.00	Brown (10YR 5/4) calcareous, clayey-silt, oxidized at 3.94 - 4.00	#9 (3.88 - 3.93)
4.00 - 4.65	Buff (10YR 5/3) coarse silt, charcoal bands at 4.09, 4.54, 4.56	#10 (4.05 - 4.10) #11 (4.41 - 4.46)
4.65 - 5.05	Gray (10YR 4/1) clayey-silt, sand inclusions at 5.00 - 5.05	#12 (4.65 - 4.69) #13 (5.00 - 5.04)
5.05 - 5.51	Dark gray (5Y4/1) clay, bands of sand at 5.24 - 5.26 and 5.39 - 5.41	#14 (5.11 - 5.15) #15 (5.44 - 5.48)
5.51 - 5.70	Fine sand (5.51 - 5.54) grading into medium sand	#16 (5.57 - 5.61)
5.70+	Gray (7.5R 4/0), coarse sand	

Core #5

0 - 0.20	Surface soil, cumulic A, dark brown (10YR 3/3) fine sand	#1 (0.10 - 0.14)
0.20 - 0.40	Bw, brown (10YR 4/2) fine sand	#2 (0.24 - 0.27)
0.40 - 0.88	Buff (10YR 4/3) fine sand	#3 (0.42 - 0.45) #4 (0.62 - 0.66) #5 (0.75 - 0.79)
0.88 - 1.08	Ab (paleosol), black (10YR 3/1) fine sand	#6 (0.89 - 0.93)
1.08 - 1.43	Bwb, black (10YR 3/1) fine sand, blocky structure	#7 (1.08 - 1.12) #8 (1.36 - 1.40)
1.43 - 1.66	Brown (10YR 4/3) calcareous coarse silt	#9 (1.50 - 1.54)
1.66 - 2.60	Buff (10YR 5/2) calcareous coarse silt	#10 (2.00 - 2.05) #11 (2.35 - 2.40) #12 (2.56 - 2.60)

Stratigraphy (m)		Samples (m)
2.60 - 3.11	Buff (10YR 5/3) coarse silt, calcareous	#13 (2.63 - 2.66) #14 (2.78 - 2.81) #15 (2.97 - 3.01)
3.11 - 3.41	Buff (10YR 5/3), non-calcareous fine sand	#16 (3.20 - 3.24)
3.41 - 3.63	Ab (paleosol), dark brown (10YR 4/2) coarse silt	#17 (3.43 - 3.47) #18 (3.57 - 3.61)
3.63 - 3.83	Buff (10YR 5/3) calcareous coarse silt	#19 (3.76 - 3.80)
3.83 - 4.05	Ab (paleosol), black (10YR 3/1), calcareous clayey silt	#20 (3.88 - 3.91) #21 (4.00 - 4.03)
4.05 - 4.36	Btb, dark brown (10YR 3/2), silty clay, calcareous	#22 (4.10 - 4.13) #23 (4.22 - 4.25) #24 (4.32 - 4.35)
4.36 - 4.54	Buff (10YR 5/3), non-calcareous coarse silt	#25 (4.47 - 4.50)
4.54 - 4.76	Interlaminated coarse silt and fine sand	#26 (4.54 - 4.58) #27 (4.68 - 4.71)
4.76 +	Fine sand	#28 (4.76 - 4.78)
(Augering)		
ca. 6.10	Interbedded gray (7.5R 4/0) clay and clayey sand	#29 (ca. 7.60)

Core #7

0 - 0.40	Surface soil (cumulic A), dark brown (10YR 3/2) coarse silt	
0.40 - 0.63	Bw, calcareous, brown (10YR 4/3), coarse silt	#1 (0.60 - 0.65)
0.63 - 4.20	Buff (10YR 6/3), calcareous coarse silt, with lenses of fine sand (2.56 - 2.63, 2.78 - 2.89)	#2 (0.87 - 0.92) #3 (1.62 - 1.66) #4 (2.15 - 2.18) #5 (2.82 - 2.86) #6 (3.52 - 3.56)
4.20 - 5.00	Brown (10YR 3/3) coarse silt, charcoal flecks (incipient A?)	#7 (4.24 - 4.27) #8 (4.83 - 4.87)
5.00 - 5.40	Buff (10YR 4/2), coarse silt	#9 (5.07 - 5.11)

Stratigraphy (m)		Samples (m)
5.40 - 5.80	Brown (10YR 4/3), coarse silt, cumulic Ab, charcoal (5.40 - 5.48)	#10 (5.40 - 5.43) #11 (5.77 - 5.80)
5.80 - 6.95	Buff (10YR 5/3) coarse silt	#12 (6.04 - 6.07) #13 (6.40 - 6.43) #14 (6.86 - 6.91)
6.95 - 7.11	Ab, brown (10YR 4/3) coarse silt	#15 (6.99 - 7.01)
7.11 - 7.35	Buff (10YR 5/4), calcareous coarse silt	#16 (7.17 - 7.20)
7.35 - 8.16	Brown (10YR 4/3) coarse silt, calcareous (incipient Ab?)	#17 (7.40 - 7.43) #18 (8.04 - 8.07)
8.16 - 8.34	Buff (10YR 5/3), coarse silt	#19 (8.26 - 8.29)
8.34 - 8.44	Brown (10YR 4/2), coarse silt, charcoal (incipient Ab?)	#20 (8.37 - 8.39)
8.44 - 8.78	Buff (10YR 5/3), calcareous coarse silt	#21 (8.54 - 8.57)
8.78 - 9.00	Brown (10YR 4/2), calcareous coarse silt, charcoal flecks (incipient Ab?)	#22 (8.79 - 8.81) #23 (8.97 - 9.00)
9.00 - 10.84	Buff (10YR 5/3), calcareous coarse silt, carbonate nodules	#24 (9.40 - 9.43) #25 (9.90 - 9.93) #26 (10.30 - 10.33) #27 (10.63 - 10.66)
10.84 - 11.0	Buff (10YR 4.3), calcareous clayey silt	#28 (10.92 - 10.95)

Schoenenburg Site

Schoenenburg Cutbank

0 - 0.20	Black (10YR 3/1) fine sand	#38 (0.08 - 0.15)
0.20 - 0.60	Fine sand	#37 (0.26 - 0.29) #36 (0.40 - 0.43)
0.60 - 0.62	Medium sand	
0.62 - 0.69	Fine sand	#35 (0.65 - 0.69)
0.69 - 0.71	Medium sand	
0.71 - 0.90	Fine sand	#34 (0.76 - 0.82)
0.90 - 0.93	Medium sand	
0.93 - 1.01	Fine sand	#33 (0.96 - 1.01)

Stratigraphy (m)		Samples (m)
1.01 - 1.02	Clay	
1.01 - 1.02	Medium sand	
1.02 - 1.05	Fine sand	#32 (1.03 - 1.05)
1.04 - 1.07	Medium sand	#31 (1.05 - 1.07)
1.07 - 1.20	Fine sand	#30 (1.12 - 1.14)
1.20 - 1.28	Medium sand	#29 (1.23 - 1.26)
1.28 - 1.63	Laminated fine sand and coarse silt	#28 (1.34 - 1.38)
1.63 - 2.00	Black (10YR 2/1) clayey-silt, upper paleosol Ab, top appears churned	#27 (1.90- 1.96)
2.00 - 2.50	Bwb, dark brown (10YR 3/3) clayey-silt	#26 (2.30 - 2.40)
2.50 - 2.94	Massive coarse silt	#25 (2.50 - 2.54) #24 (2.70 - 2.76)
2.94 - 3.20	Brown (10YR 3/3) fine sand, middle paleosol Ab, top appears churned	#23 (2.96 - 3.03) #22 (3.16 - 3.20)
3.20 - 3.78	Fine sand	#21 (3.22 - 3.27) #20 (3.40 - 3.43) #19 (3.66 - 3.71)
3.78 - 3.92	Medium sand	#18 (3.87 - 3.91)
3.92 - 4.02	Fine sand	#17 (3.94 - 3.98)
4.02 - 4.07	Medium sand	#16 (4.03 - 4.05)
4.07 - 4.12	Fine sand	#15 (4.09 - 4.12)
4.12 - 4.14	Medium sand	
4.14 - 4.15	Fine sand	
4.15 - 4.19	Medium sand	#14 (4.15 - 4.18)
4.19 - 4.22	Fine sand	#13 (4.21 - 4.22)
4.22 - 4.35	Laminated clayey-silt, 5 clay, 4 silt laminae	#12 (4.28 - 4.30) #11 (4.30 - 4.33)
4.35 - 4.60	Fine sand	#10 (4.45 - 4.50)
4.60 - 4.87	Grayish-brown (2.5Y 5/2) clay	#9 (4.65 - 4.71)
4.87 - 5.30	Black (10YR 3/1) clayey-silt, lowermost paleosol Ab, wavy upper boundary, granular structure, non-calcareous	#8 (4.96 - 5.02)
5.30 - 5.57	Bwb, dark brown (10YR 4/2), clayey, calcareous	#7 (5.41 - 5.47)
5.57 - 5.80	Gray (2.5Y 5/2), calcareous clay	#6 (5.58 - 5.63) #5 (5.70 - 5.77)

Stratigraphy (m)		Samples (m)
5.80 - 6.13	Mottled, gray (10YR 6/1) silty clay	#4 (5.98 - 6.04)
6.13 - 6.30	Gray (10YR 5/3), oxidized massive coarse sand	#3 (6.19 - 6.27)
6.30 - 6.50	Gray (10YR 6/1), oxidized massive fine sand	#2 (6.31 - 6.39)
6.50 - 6.60	Plane bedded, coarse sand	#1 (6.52 - 6.60)
6.60+	Slump	

Core #8

0.0 - 0.45	Dark brown (10YR 2/2), fine sand, cumulic A	#1 (0.11 - 0.15)
0.45 - 0.81	Brown (10YR 4/3), fine sand, Bw	#2 (0.45 - 0.48)
0.81 - 1.23	Buff (10YR 4/3), fine sand, carbonate nodules	#3 (0.83 - 0.87) #4 (1.15 - 1.20)
1.23 - 1.57	Ab (uppermost paleosol), black (10YR 2/2), coarse silt	#5 (1.26 - 1.30) #6 (1.49 - 1.54)
1.57 - 3.36	Buff (10YR 6/4), inter-laminated coarse silt and fine sand, calcareous (1.80 -2.24)	#7 (1.59 - 1.63) #8 (1.80 - 1.84) #9 (2.38 - 2.41) #10 (2.79 - 2.83) #11 (3.11 - 3.13) #12 (3.16 - 3.19)
3.36 - 4.00	Ab (middle paleosol), black (10YR 2/2), fine sand, load features at upper boundary	#13 (3.42 - 3.48) #14 (3.62 - 3.65) #15 (3.83 - 3.88)
4.00 - 4.96	Buff (10YR 6/4), interlaminated coarse silt and fine sand	#16 (4.18 - 4.22) #17 (4.64 - 4.67)
4.96 - 5.20	Incipient Ab, brown (10YR 6/3), clayey silt	#18 (4.96 - 5.00)
5.20 - 5.38	Buff (10YR 5/3), coarse silt	#19 (5.30 - 5.34)
5.38 - 5.76	Brown (10YR 4/3), coarse silt (incipient Ab?)	#20 (5.40 - 5.44)
5.76 - 5.92	Buff (10YR 5/4), coarse silt	#21 (5.82 - 5.86)

Stratigraphy (m)	Samples (m)
5.92 - 6.15 Ab (lowermost paleosol), black (10YR 3/3) clayey silt	#22 (5.93 - 5.97)
6.15 - 6.60 Bwb, dark brown (10YR 4/4), clayey silt, carbonate nodules	#23 (6.32 - 6.35)
6.60 - 6.64 Gray (7.5Y 4/1) clay	#24 (6.60 - 6.62)
6.64 - 7.38 Brown (10YR 6/4), interlaminated fine sand and coarse silt	#25 (6.75 - 6.78) #26 (7.31 - 7.35)
7.38 - 8.00 Buff (10YR 5/3), fine sand	#27 (7.40 - 7.43) #28 (7.80 - 7.83)

Core #14

0 - 0.43 Cumulic Ab, black (10YR 1/1) fine sand	#1 (0.20 - 0.24)
0.43 - 0.63 Buff (10YR 4/3), coarse silt	#2 (0.51 - 0.54)
0.63 - 1.1 Ab (uppermost paleosol), black (10YR 1.1) coarse silt, calcareous (0.89 - 1.10)	#3 (0.66 - 0.70) #4 (0.86 - 0.90)
1.10 - 1.84 Bwb, dark brown (10YR 3/1) coarse silt	#5 (1.11 - 1.15) #6 (1.35 - 1.39) #7 (1.50 - 1.56)
1.84 - 2.24 Buff (10YR 5/3) coarse silt	#8 (1.96 - 2.00)
2.24 - 2.70 Ab (lowermost paleosol), black (10YR 3/3) coarse silt, calcareous (2.48 - 2.70)	#9 (2.28 - 2.32) #10 (2.47 - 2.50) #11 (2.65 - 2.69)
2.70 - 5.08 Buff (10YR 5/3) clayey silt, carbonate nodules, pebbles, lenses of medium sand and coarse sand (4.75 - 4.90)	#12 (2.83 - 2.87) #13 (3.13 - 3.16) #14 (3.53 - 3.56) #15 (3.90 - 3.94) #16 (4.56 - 4.59) #17 (4.90 - 4.93)
5.08 - 6.00 Buff (10YR 5/3), coarse silt	#18 (5.11 - 5.14) #19 (5.70 - 5.74)
6.00 - 6.20 Coarse sand	#20 (6.14 - 6.19)
6.20 - 6.27 Fine sand	#21 (6.24 - 6.27)

Stratigraphy (m)	Samples (m)
6.27 - 6.63 Buff (10YR 5/3), coarse silt with pebbles	#22 (6.38 - 6.41)
6.63 - 6.81 Medium sand	#23 (6.65 - 6.70)
6.81 - 6.92 Buff (10YR 5/3), silty clay	#24 (6.84 - 6.87)
6.92 - 7.23 Medium sand	#25 (7.08 - 7.12)
7.23 - 7.80 Interlaminated buff (10YR 5/3) fine sand and coarse silt	#26 (7.23 - 7.26) #27 (7.26 - 7.29) #28 (7.63 - 7.66) #29 (7.70 - 7.73)

Core #16

0 - 0.19 Black (10YR 2/1), coarse silt, cumulic A	#1 (0.02 - 0.05) #2 (0.15 - 0.19)
0.19 - 0.55 Bw, dark brown (10YR 5/4), coarse silt	#3 (0.24 - 0.29) #4 (0.48 - 0.52)
0.55 - 1.05 Buff (10YR 5/3), calcareous coarse silt	#5 (0.55 - 0.60) #6 (0.93 - 0.97)
1.05 - 1.14 Buff (10YR 6/3), calcareous sandy silt	#7 (1.09 - 1.13)
1.14 - 3.03 Buff (10YR 5/4), sandy silt, lenses of fine sand, carbonate nodules	#8 (1.19 - 1.23) #9 (1.46 - 1.50) #10 (1.68 - 1.72) #11 (2.16 - 2.20) #12 (2.40 - 2.44) #13 (2.64 - 2.69)
3.03 - 3.12 Buff (10YR 6/3), calcareous clayey silt	#14 (3.04 - 3.07)
3.12 - 3.16 Oxidized (10YR 5/6), silty clay	#15 (3.12 - 3.16)
3.16 - 3.48 Buff (10YR 6/3), clayey coarse silt and fine sand, carbonate nodules	#16 (3.22 - 3.26)
3.48 - 3.58 Buff (10YR 6/4), clayey sand, pebbles	#17 (3.51 - 3.54)
3.58 - 3.81 Buff (10YR 6/4), coarse silt	#18 (3.65 - 3.69)
3.81 - 4.80 Yellowish-brown (2.5Y 6/4) clay, oxidized	#19 (3.85 - 3.89) #20 (4.19 - 4.23) #21 (4.19 - 4.49) #22 (4.70 - 4.75)

APPENDIX B: RADIOCARBON DATING OF PALEOSOLS AND RADIOCARBON AGES USED IN THIS STUDY

Since its development in the 1950s, radiocarbon dating has been used extensively by researchers attempting to reconstruct late Quaternary history. Radiocarbon dating of material in river and wind deposits has proven to be a major tool in investigating the environmental changes that have occurred during that period. Because many of these deposits lack datable materials such as wood, charcoal, or bone, investigators have turned to radiocarbon dating of organic material from paleosols to determine the timing of depositional and erosional events. This is especially true in the central Great Plains, where the scarcity of trees during the late Quaternary has reduced the chances of finding wood or charcoal preserved in deposits.

Radiocarbon dating of paleosols measures the decay of organic carbon present in soil humates. As a soil forms at the surface, fresh humus from decaying plants and animals is continually added to it and mixed with existing humus. The total humus content of a soil, then, is a mixture of young and old humus. Because the ages of the humates vary, humate dates on soils measure the mean residence time of the humates in the soil. The soils that were dated in this research are cumulic soils which formed when pedogenesis proceeded in conjunction with slow sedimentation (see Riecken and Poetsch 1960).

Soil formation proceeds until the land surface is disturbed by either erosion and removal of the soil or by an increase in the rate of sediment deposition and burial of the soil. Once sufficiently buried, and thereby isolated from the surface, the soil no longer receives inputs of young humus (or does so at a reduced rate), and the mixing of young and old humus ceases (or slows) (see Schaetzl and Sorenson 1987).

Although the process of radiocarbon dating of soil humates appears fairly straightforward, there are several factors that affect the ages obtained. First, the total humate content of a soil comprises three fractions: fulvic acid, humic acid, and fine residuals (humin) (Matthews 1980). The three fractions differ in solubility, with humin being completely insoluble, and humic acid and fulvic acid being alkali and acid soluble, respectively (Matthews 1980; Haas et al. 1986). In theory, fulvic and humic acids, because they are soluble, should possess the youngest mean residence time, and therefore produce the youngest ages. Conversely, humin, being the least mobile of the fractions, should result in the oldest ages (Matthews 1980). The humic acid fraction was dated in this study. Second, fresh humus is continually mixed with older humus in a soil until that soil is sufficiently buried, at which time the addition of young material ceases (Schaetzl and Sorenson 1987). Therefore, the total humus content of a soil comprises organic matter of various ages. A radiocarbon age from a buried soil, because it measures the mean residence time of a mixture of young and old humates, provides only an approximate age for the soil. Third, the freshness of the sampled exposure affects the radiocarbon age obtained. Freshly excavated exposures have been found to produce ages that are up to 1000 years older than ages obtained from exposures that have been open for several years (Haas et al. 1986). Finally, the size of the organic fraction being dated has been shown to affect radiocarbon ages. Specifically, smaller size fractions produced younger ages than did larger size fractions (Stuckenrath et al. 1979).

Notwithstanding these limitations, a sample from the top of a cumulic A horizon should produce a younger age than a sample from the base of a cumulic A horizon since the humus higher in the profile was incorporated after that in the lower part of the profile. That is, in a paleosol a date from the base of the A horizon provides a minimum time for the beginning of pedogenesis, whereas a date from the upper portion of the A horizon yields the approximate time for the burial of the soil (Haas et al. 1986).

Radiocarbon Ages Used in this Study (1)

Lab#	Uncorrected Age (2)	C13 Corrected Age (2)	Calibrated Age (3)
Bone Cove			
Tx-5910	26,260±680	--	N/A
North Cove			
DIC-3358	16,130±270	--	N/A
Tx-6321	11,420±150	11,530±150	N/A
Tx-6319	10,400±160	10,550±160	N/A
Tx-6320	10,120±160	10,270±160	N/A
DIC-3357	6860±100	--	7676+93/-97
Prairie Dog Bay			
Tx-5909	10,360±130	--	N/A
Alma Vista			
Tx-5979	--	4550±80	5290+30/-240
Schoenenburg			
Tx-5977	--	3720±90	4087+146/-152
Tx-5912	--	3050±60	3296+65/-87
Tx-5978	--	2780±80	2870+102/-83
Tx-5911	--	2020±60	1985+69/-86

(1): All ages were obtained on soil humates from bulk (ca. 2 kg) samples

(2): For a discussion of the C13 isotope correction procedure, see Stuiver and Polach (1977) and Taylor (1987).

(3): Calibration from a conventional radiocarbon age to calibrated calendar years using a tree ring calibration curve. The calibration accounts for variation in C14 activity in the atmosphere. All calibrations reported here were done using the 20-year atmospheric curve (see Linick et al. 1985, 1986; Kromer et al. 1986; Mook 1986; and Stuiver et al. 1986). Calibration possible for only the last 9200 years. Program used for calibration is contained in Stuiver and Reimer (1986).

APPENDIX C: PHI CALCULATIONS AND STATISTICAL TESTS USED IN THIS STUDY

Phi Calculations

This appendix contains the formulas and statistical tests used in the analysis of sediments in the study area. As noted in Chapter 4, particle size data were converted to the logarithmic phi scale using a computer program developed by M.C. Prante of the Department of Geography, University of Kansas. This program constructs a cumulative frequency distribution for the particle size distribution of a given sample, and then reads the desired phi value from it. In this manner, the fifth, fiftieth, and eighty-fourth phi values were calculated. Two additional parameters, the mean phi value (Mean ϕ) and the degree of sorting ($\sigma\phi$), were also calculated using the following equations (Folk and Ward 1957):

$$\text{Mean } \phi = \frac{16\phi + 50\phi + 84\phi}{3}$$

$$\sigma\phi = \frac{84\phi - 16\phi}{4} + \frac{94\phi - 5\phi}{6.6}$$

The degree of sorting represents the dispersion of sediment sizes around the mean phi value. Folk and Ward (1957, 13) designate the following qualitative descriptions of the degree of sorting based on the calculated value of $\sigma\phi$:

< 0.35 = Very Well Sorted
0.35 - 0.50 = Well Sorted
0.50 - 1.00 = Moderately Sorted
1.00 - 2.00 = Poorly Sorted
2.00 - 4.00 = Very Poorly Sorted
> 4.00 = Extremely Poorly Sorted

Statistical Tests

For the Freer and Schoenenburg sites, several basic statistical tests were used to differentiate among the degree of sorting of fluvial, colluvial, and eolian deposits. Means ($\bar{x}$) were calculated as follows (Hammond and McCullagh 1974, 7):

$$\bar{x} = \frac{\Sigma x}{n}$$

x = variable
n = number of variates

Standard deviations (s) were calcualted as follows (Hammond and McCullagh 1974, 11):

$$s = \sqrt{\frac{\Sigma(x - \bar{x})^2}{n}}$$

x = variable
$\bar{x}$ = mean of x
n = number of variates

The coefficient of relative variation (CRV), which permits the comparison of two means that differ in magnitude, was calculated as follows:

$$CRV = \frac{s}{\bar{x}}$$

s = standard deviation of x
$\bar{x}$ = mean of x

To test whether the difference between two means was the result of chance or of consistent differences in the data, an unpaired t-test was used. Because the t-test is a parametric test, two conditions must be met in order to conduct the analysis:

1) The data must be normally distributed
2) The variances of the data must be equal

In this case, the large sample size ($n > 25$) allows one to assume that the population is normally distributed. The variances were checked using a variance ratio test (F-test) (see Hammond and McCullagh 1974, 169-171). The calculated F value was less than the critical F, meaning that the variances were not statistically different, and a t-test could be used.

The following unpaired t-test was used to determine whether the mean sediment sorting of Core #16 (Schoenenburg site, p. 120) was significantly different than that of Core #7 (Freer site, pp. 108-110). Critical values of F and t were obtained from statistical tables in Taylor (1977,352,346). The hypotheses were:

H_o: There is no significant difference in the mean sediment sorting values of Core #16 and Core #7.

H_1: The mean sediment sorting value of Core #7 is significantly less than the mean sorting value of Core #16.

Level of Significance (one-tailed) = 0.05

F-calculated = 1.43
F-critical = 2.03 (at 0.05 level)
Variances are not significantly different.

t-calculated = 18.9
t-critical = 2.02 (at 0.05 level)
Reject H , accept H . The mean sorting value of Core #7 is significantly less than the mean sorting value of Core #16 (i.e., the sorting of Core #7 is better than that of Core #16).