Late Wisconsinan and Holocene Environments of the Central Great Plains: A Status Report for Geoarchaeological Investigations

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Our intent in this paper is to summarize that which is currently known about the paleoenvironmental history of the central Great Plains, i.e., the states of Kansas, Nebraska, and easternmost Colorado (Figure 1). Objectives are to summarize the current status of knowledge regarding environmental changes in the central Great Plains during the late Quaternary and to briefly identify gaps that exist in the data base. Such a summary provides necessary context for understanding the cultural history of the region. Archaeologists have long recognized that the data base is not inherited in a pristine state, i.e., the archaeological record preserved in the landscape is an incomplete rendering of past cultural activities that left fossil remains vulnerable to a variety of post-depositional processes. Consequently, the record must be extracted and the extent of modification be appreciated.

Less is known about environmental conditions in the central Great Plains during the late Pleistocene and Holocene than many other regions of North America, due primarily to an inability to widely apply the more traditional investigative tools such as palynology and dendroclimatology. Consequently, that which is known has been and is being derived using some of the newer approaches to late-Quaternary environmental reconstruction such as stable isotope analysis, opal phytolith analysis, and rock magnetism. Those sedimentological contexts being explored include loess, eolian sand sheets and dunes, alluvial fill, and to a lesser extent isolated lake and peat deposits. The loess deposits of the region, which represent some of the thickest and most complete loess accumulations in North America, hold the potential to provide a particularly promising avenue for the pursuit of the paleoenvironmental record.

Only the Wisconsinan Stage of the Pleistocene and the Holocene are addressed, since this is the period for which the most data are available and of greatest relevance to a prehistoric cultural context. Further, the establishment of radiocarbon chronostratigraphies permits correlations with the cultural overlays. First, however, the climatic and physiographic settings are presented as a backdrop.

CLIMATIC SETTING

The principal climatic features of the central Great Plains are its continentality and Rocky Mountain-induced rainshadow. The Prairie Wedge, which dominates the study region, is a consequence of the zonal westerly airflow crossing the western mountains and penetration of modified Pacific air mass (Borchert 1950, 1971). Isohyet patterns exhibits a longitudinal orientation, with mean annual precipitation decreasing from over 40 inches (1000 mm) at the southeastern margin to less than 15 inches (380 mm) in the western and northern parts (Figure 2). Winters are typically cold with relatively little precipitation, mostly as snow; summers are hot with increased precipitation, chiefly associated with collision of Pacific (mP) and Arctic (cA) air masses with warm, humid air masses from the Gulf of Mexico. Since it determines the carrying capacity of the region, drought is the most significant climatic element of the Great Plains environment from the ecological, historical, and prehistoric standpoints (Weakley 1943; Barry 1983; Wedel 1986). Vegetation is mostly prairie grassland, due to the subhumid-semiarid, markedly seasonal climate. The mean tropical Atlantic airflow (mT) that influences the grassland east of about the 100th meridian in normal summers has tended to give away during the summers of drought years to continental flow (Figure 3).

The prairie crosses the region from north to south in three broad zones (Figure 4). In the west, the grama-buffalograss prairie consists of short grasses, while the bluestem prairie with its tall grasses and many forbs prevails in the east. Between them lies the mixed prairie with tall, medium, and short grasses (Küchler 1964; 1974). In the sand sheets of southeastern Colorado, west-central Nebraska and central Kansas, edaphic conditions promote the existence of a sandsage-bluestem prairie. The sensitivity of prairie composition and boundaries to short-term climatic variation during the historical period is well documented for the region (Tomanek and Hulett 1970). Similarly, long-term prairie expansion and contraction, presumably in response to climatic variation, is documented at the prehistoric time scale (e.g., Watts and Wright 1966; Grüger 1973; Bernabo and Webb 1977; Bradbury 1980). The consequence of short- and long-term climatic variations within the central Great Plains and attendant changes in the vegetation probably had measurable impact on prehistoric peoples, but the magnitude of such is, however, certainly open to question (cf. Wedel 1961, Reeves 1973, Johnson 1990).

According to Borchert (1950), regional distinctiveness of the grassland climate lies basically in the precipitation. Low snowfall and low rainfall in the region are typical of winter. There is a greater risk of a large rainfall deficit in summer within the grassland than in the bordering regions of forests. The short-grass steppe receives markedly less rainfall than the remainder of the continental

United States east of the Rockies during the summer. The grassland is distinguished from the forest region to the north by fewer days with precipitation, less cloud cover, and lower humidity, on the average, during July and August. It is also characterized by large positive departures from average temperature and by frequent hot winds during summer.

PHYSIOGRAPHIC REGIONS

The Great Plains physiographic region lies east of the Rocky Mountains and extends from southern Alberta and Saskatchewan to nearly the United States-Mexico border (Figure 1). The central Great Plains is a large region of generally low relief sloping eastward from the Rocky Mountains toward the Missouri and Mississippi Rivers. Multiple continental glaciations, starting perhaps as early as 2.5 million years ago (Boellstorff 1978), caused reorientation of the Missouri River system southeastward to the Mississippi River, resulting in many stream captures and other geomorphic changes (Wayne et al. 1991). Each time the ice blocked eastward-flowing rivers, proglacial lakes formed, spilled across divides, and developed new courses around the glacial margin. The present course of the Missouri River through North and South Dakota is chiefly along a late Illinoian ice margin. The Platte River evolved through spasmodic uplift of the Chadron arch (Stanley and Wayne 1972) and several early and middle Pleistocene glacial advances into eastern Nebraska and northeastern Kansas (Aber 1991). In the middle Pleistocene, the Platte River joined with the glacially diverted Missouri River and formed a wide alluvial plain across east-central Nebraska and northeast Kansas. Quaternary erosion of the central Great Plains, which largely is drained by the Missouri River, has been mostly by fluvial processes. However, the channel network in much of the Missouri River basin was the result of drainage rearrangements by glaciation.

In extreme southeastern Kansas, a small portion of the Ozark Plateau extends into Kansas. The streams of this region have carved the thick, flint-bearing Mississippian limestones into the present-day topography. To the west of the Ozark Plateau is the broad Cherokee Plain, a region developed on thick shale beds of the Cherokee Formation of middle Pennsylvanian age. The low-gradient, shallow streams have planed the surface of this low-relief region.

West and north of the Cherokee Plain, the topography consists of a series of parallel northeast-southwest trending cuestas. Cuesta topography is developed as a series of ridges having a sharp slope on the east side and a gentle slope on the west side. A series of relatively erosion-resistant limestone strata exposed at the surface descend gently westward until they dip under the outcrop of thick overlying shale. The shale underlying the capping limestone is less resistant to weathering and erodes to form an abrupt east-facing escarpment.

To the west of the Osage Cuestas is a band of grass-covered limestone hills, the Flint Hills, that constitute a preserve of the Kansas tall-grass prairie. The Flint Hills are located at the eastern edge of huge expanse of grass-covered plains that extends continuously westward to the Front Range of the Rocky Mountains, northward into Canada, and southward into northern Texas. At the western margin, the Flint Hills dip gently under younger rocks which, on the north, slope gently westward under the Smoky Hills escarpment. To the south, these strata dip below the McPherson-Wellington Lowlands. Extending from north of Salina, Kansas southward to the Oklahoma border, the McPherson-Wellington Lowlands mark the outcrop belt of the thick Wellington shale.

Along the southern border of Kansas, from Harper and Kingman counties to eastern Meade County, erosion has exposed the Red Hills. The badlands topography of the Red Hills is unique to the central Great Plains. In some areas, erosion-resistant dolomite caps the red Permian strata resulting in buttes and mesas.

Extensive areas of grass-covered sand dunes lying south of the Arkansas River constitute the Great Bend Sand Prairie. A similar region exists south of the Cimarron River in the southwestern corner of Kansas. During the Late-Pleistocene and Holocene, strong winds eroded fine sediments from alluvial surfaces of the Arkansas River and transported them to the dune area, which covered hundreds of square miles. North of the Great Bend Sand Prairie, Cretaceous rocks, exposed over a large portion of western Kansas, constitute the Smoky Hills, named such because of their dark shales.

From Saskatchewan to northern Texas, lies the High Plains. Viewed from a broad perspective, the whole of the High Plains surface is upheld by a huge wedge-shaped, alluvial apron consisting of sediment derived from erosion of the eastern Rocky Mountains. These sediments, the Ogallala Formation, represent Miocene stream deposition similar to that presently occurring in the Arkansas

River to the south. The Ogallala Formation is composed not only of river-borne sands and gravels, but also of loess, volcanic ash beds, and diatomite deposits.

The eastern part of the central Great Plains is an area of rolling hills that was invaded by one or more glacial ice masses during the Pleistocene. With the abandonment of the classic Nebraskan and Kansan nomenclature for glacial stages, the intrusions of ice into this region are designated as pre-Illinoian, with the exception of extreme northeastern Nebraska, which experienced glaciation during the Wisconsinan Stage. The use of the traditional terms has been confused and has become ambiguous, e.g., coring "Nebraskan" till yielded three separate tills separated by distinct soils (Hallberg 1986). Despite the lack of a chronology and nomenclature, glaciers occupied the eastern fifth of Nebraska and northeastern Kansas. Glaciers apparently advanced into Kansas twice from two different directions, the second of which did so between 600 ka and 700 ka (Aber, 1991).

Central and western Nebraska is mantled by extensive deposits of Wisconsinan to late-Holocene eolian sands known as the Sand Hills. The age and origin of this spectacular eolian feature are yet uncertain. Ahlbrandt et al. (1983) suggested that the dunes are late-Holocene features, possibly derived from older, unconsolidated sediment that mantled the Great Plains. In contrast, Wells (1983) regarded the Sand Hills as a coarse, upwind facies of a single late-Pleistocene sand-silt unit.

There are three dune fields in northeastern Colorado (Figure 5). The relatively small Greeley dune field is located immediately north of the South Platte River, and the larger Fort Morgan dune field lies south of the river. The Wray dune field, the largest, is on the High Plains to the east and southeast of the other two. The Fort Morgan and Wray sands were probably derived from sediment of the South Platte River under northwest winds during late-Holocene time. The source and direction of movement of the Greeley sand hills is, however, uncertain (Muhs 1985).

LATE WISCONSINAN STAGE

Due to its relative youth, the Wisconsinan Stage has the greatest chronostratigraphic resolution. Based on the chronology from Illinois, five substages of the Wisconsin have been traditionally recognized: the Altonian (70,000-28,000 yr B.P.), Farmdalian (28,000-22,000 yr B.P.), Woodfordian (22,000-12,500 yr B.P.), Twocreekan (12,500-11,000 yr B.P.), and Valderan (11,000-

5,000 yr B.P.) (Willman and Frye 1970; Frye and Willman 1973). This chronology of substages has, however, limited stratigraphic application in Nebraska, Kansas and eastern Colorado, and has therefore not been adopted literally.

In Nebraska, Reed and Dreeszen (1965) identified four Wisconsinan units: the Gilman Canyon Formation (an upland loess with soil development), Peoria Formation (fluvial sand and silt in valleys and loess on the uplands), Brady Interstadial soil, and Bignell Formation (dune sand and loess). For the Wisconsin of Kansas, Frye and Leonard (1952) recognized early Wisconsinan alluvial deposits and the Sanborn Formation. The late Wisconsinan units of the latter included the Peoria loess, Brady soil and Bignell loess (Figure 6). Since these early statements of stratigraphic succession, the Bignell loess has been assigned to the Holocene.

During the 1960s, the record of past climate was based primarily on continental deposits, but these were rarely continuous sedimentary records, and consequently the picture of past climatic variations that developed was therefore incomplete (Bradley 1985). In the next decade, studies of marine sediments revolutionized our understanding of climatic variations and enabled models of the causes of climatic changes to be tested. Undoubtedly, studies of marine sediments have provided data bases which continue to expand in quantity and quality (Ruddiman 1985). However, the 1980s experienced a renewed focus on continental records of climate, which complement the perspective provided by marine sediment (COHMAP members 1988). Continental deposits often provide more detailed information about short-term (high-frequency) changes of climate than most marine records (Broecker et al. 1988).

Geomorphology and Stratigraphy

From Nebraska, loess deposits in the central Great Plains extend west across eastern Colorado and south across most of Kansas. The thickest deposits of loess are adjacent to and underlie the Nebraska Sand Hills (Kollmorgen 1963; Ahlbrandt et al. 1980). The oldest laterally extensive loess unit in the region is the Loveland loess, found at least as far west as the Colorado/Kansas state line. In Kansas, exposures of Loveland loess are patchy and are found mostly on ridges near drainages (Welch and Hale 1987). The Peoria loess is the thickest and most laterally continuous loess deposit

(Frye and Leonard 1951), whereas the overlying Holocene Bignell loess is found discontinuously in the central Great Plains and is not identified east of the Missouri River.

Until recently, little age control existed for the timing of loess deposition in the central Great Plains; age assumptions were based largely on the classical continental glaciation sequence, similar to that used for the loess stratigraphy in the Mississippi and Missouri River Valleys. However, younger loesses and associated paleosols exposed at a number of localities in Kansas and Nebraska have been systematically radiocarbon and, to a lesser extent, thermoluminescence-dated (e.g., Souders and Kuzila 1990; Johnson 1993a; Martin 1993; May and Holen 1993; Feng et al. 1994a, 1994b; Maat and Johnson 1996). According to these age estimates, the Gilman Canyon Formation was deposited from at least 40 ka to about 20 ka, Peoria loess about 20 ka to 10.5 ka, and Bignell loess from at least 9 ka to about 5.5 ka.

Unconformably overlying the Loveland loess and the Sangamon soil that caps it is the loess of the Gilman Canyon Formation (Reed and Dreeszen 1965). The upper two-third of this loess is organic-rich and contains one or more cumulic A horizons that represent a period of slow accumulation and pedogenesis. Similarities in stratigraphic position, soil development and molluscan assemblages indicate it may be equivalent to the Farmdale interstadial soil (Johnson 1990, 1993a). Recent age estimates also support correlations with the Roxana silt of the Upper Mississippi River Valley (Leigh and Knox 1994) and the Pisgah Formation of western Iowa (Forman et al.1992a). A transitional zone of loess, characterized by upwardly decreasing organic content and increasing color value (brightness), represents a slowly accelerating rate of loess deposition and separates the Gilman Canyon Formation from the overlying Peoria loess.

Peoria loess

Late Wisconsinan loess deposits mantle much of upland surface of the region covering the central Great Plains from the North Dakota/South Dakota border to that of Kansas and Oklahoma and provide a terrestrial record of late Quaternary climate. Thickest deposits lie adjacent to the Missouri River and its major tributaries (Ruhe 1983).

Leverett (1899) first proposed the name Peoria for an interglacial period between the Iowan and Wisconsinan glacial stages. When Alden and Leighton (1917) demonstrated the Peoria was younger than the Iowan, usage shifted to that of a loess, rather than of a weathering interval. Within the Midcontinent, several names have been used for post-Farmdalian loess. Ruhe (1983) preferred the term "late Wisconsin loess" because of the uncertainties in the stratigraphic equivalency from one region to another.

The Peoria loess is typically eolian, calcareous, massive, light yellowish-tan to buff silt that overlies the Loveland loess or an approximate equivalent of the Gilman Canyon Formation. Based on conventional and accelerator radiocarbon ages, deposition of the late-Wisconsinan Peoria loess in Kansas and Nebraska began about 19.5-21 ka. In cold and arid conditions of the late Wisconsin, a high depositional rate was probable; at the Bignell Hill type section in southwestern Nebraska, for example, the sedimentation rate for the late-Wisconsinan Peoria loess averaged 5.7 mm/year. This rapid deposition rate seems to be comparable to marine records and rapid enough to preserve high-resolution data for the late-Wisconsinan environmental changes. The rate of accumulation for the Peoria loess was certainly variable, but apparent annual laminae are present at many localities near the Platte River valley of Nebraska, including the Bignell Hill and the Eustis ash pit sites (Johnson 1993b). Loess accumulation rates decreased as the regionally-expressed Brady soil began developing between 10.6 and 10.1 ka.

The lack of any well developed, buried soils or other unconformities suggests that Peoria loess in the region represents a continuous deposit and that the faunal zonation reflects a change in the rate of deposition. Evidence that Peoria loess deposition was episodic has emerged from western Iowa (Daniels et al. 1960; Ruhe et al. 1971), central Kansas (Arbogast 1995) and southwestern Illinois (McKay 1979), where deposits exhibit 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 loess thickness decreases gradually with distance to the south and southeast of the Platte River valley. Except for the loess of the Loess-Drift Hill area in southeast Nebraska, loess south of the Platte was deposited rather evenly on a nearly level surface of old alluvial sands and gravels. In the Loess-Drift Hill area of southeast Nebraska and in most of the area north of the Platte River, the loess mantles a previously dissected and hilly topography.

Little is known about the environmental conditions in the central Great Plains during Peoria loess deposition. Early studies, however, postulated that loess likely accumulated under dry conditions (e.g., Schultz and Stout 1948). Evidence from modern depositional environments suggests, however, that well-vegetated rather than barren surfaces favor loess deposition (Martin 1993). Additionally, the rich landsnail fauna of Peoria loess in the Great Plains implies deposition on a vegetated surface (Leonard 1952; Ostlie 1986; Wells and Stewart 1987a).

Recent findings indicate that trees were present in the central Great Plains during Peoria loess deposition, although the distribution and density of tree cover is unknown. Wells and Stewart (1987b) recovered *Picea glauca* (white spruce) cones, needles, and wood from Peoria loess at several sites in south-central Nebraska; radiocarbon ages on the wood range from 14,700 to 13,000 yr B.P. (Johnson 1989). Wells and Stewart (1987b) also report charcoal in Peoria loess at two locales in north-central Kansas. At the Coyote Canyon site in south-central Nebraska, bands of charcoal near the base of the Peoria loess afforded additional radiocarbon ages which range from 21,250 to 19,730 yr B.P. (Martin 1993). At the nearby Sindt Point site, Johnson also reports an additional age of 21,440 yr B.P. for *Picea* charcoal (Table 1).

A upland pollen record from southeastern Kansas suggests that a *Populus* (aspen) parkland was present during the late Pleistocene (Fredlund and Jaumann 1987). Watts and Wright (1966) conclude that *Picea* was the dominant vegetation cover in the Nebraskan Sand hills until 12,500 yr B.P., when it was gradually replaced by *Pinus* (pine) and herbaceous vegetation. Recent surveys of the Midcontinent pollen record by Webb et al. (1983) and Baker and Waln (1985) postulate parkland vegetation with treeless openings on the central Great Plains during the late Pleistocene.

Ruhe (1983) noted three major features of late-Wisconsinan (Peoria) loess: it thins downwind from the source area, decreases in particle size systematically away from the source area, and is strongly time transgressive at its base. The last feature is problematic and causes correlation problems. Ruhe (1969) realized a decrease in the age of the soil under the loess from 24,500 yr B.P. near the Missouri River to about 19,000 yr B.P. eastward across southwestern Iowa. A decrease from 25,000 to 21,000 yr B.P. was noted for the base of the loess along a transect in Illinois (Kleiss and

Fehrenbacher 1973). The top of the loess also seems to be time transgressive, ranging from about 12,500 yr B.P. in Illinois (McKay 1979) to about 14,000 yr B.P. in central Iowa (Ruhe 1969).

Despite the attention given to the Peoria loess in central Great Plains, the source of the silt is not completely certain. From their review of available data, Welch and Hale (1987) concluded that a single source was not likely for all loess deposits in Kansas and that the loess was derived from a combination of three sources: glacial outwash river flood plains, present sand dune areas, and fluvial and eolian erosion of the Ogallala Formation. The Platte River undoubtedly contributed massive quantities of loess during glacial stages, as presumed (Swineford and Frye 1951). Loess is thickest immediately south of the Platte River Valley, which suggests that the alluvium in the valley was the source of the loess, at least for those deposits adjacent to the valley (Kollmorgen 1963). Some local thickening of loess occurs to the southeast of the Platte River wherever streams enter from the Sand Hills to the northwest. With prevailing northwesterly winds, these locally thick deposits are probably partially derived from alluvium brought into this valley by these streams. In addition, nonglacial rivers in western Kansas and Nebraska probably contributed substantially more to the volume of loess in the area. Local loess deposits in excess of 23 m have been measured along the southeastern bluffs of the Arikaree and Republican Rivers (Swineford and Frye 1951). Swineford and Frye (1951) concluded that the Arkansas River carried too sandy a sediment load to act as a major loess source and suggested that most of the loess deposited south of the Arkansas River in southwest Kansas was derived from northern sources.

In Nebraska and Kansas, radiocarbon and thermoluminescence dating indicates that Peoria loess in those areas correlates temporally with the Peoria loess of Iowa, Illinois, and Indiana (e.g., Johnson et al. 1993a; May and Holen 1993; Martin 1993; Maat and Johnson 1996). However, much of the loess in Kansas and Nebraska occurs upwind of or distant from late-Wisconsinan continental glacial outwash sources. In addition, some of the thickest deposits of loess in Nebraska occur upwind of the Platte River (Swinehart 1990). Flint (1971) pointed out that the volume of loess on the Great Plains is surprisingly high if it was all generated from glacial outwash derived from the Rocky mountains. At the present time, the source of loess in the Kansas and Nebraska portion central Great Plains is unknown, and more than one source may be involved (Welch and Hale 1987).

Welch and Hale (1987) concluded that loess in northwest, north-central, west-central, central, southwest, and south-central Kansas was derived from regional sand-dune areas in central and western Nebraska (Sand Hills area), whereas loess in southwest Nebraska, eastern Colorado, southwest and south-central Kansas from alpine glacial-outwash sediments of the flood plains of the Platte and Arkansas Rivers and from flood plain sediments of nonglacial rivers such as the Arikaree, Republican, Solomon, Saline, Smoky Hill, Pawnee, and Cimarron.

Trace element analysis (e.g., cerium, strontium, yttrium, zirconium) is being employed to gain insight into loess source and paleowind directions (Johnson et al. 1993b). Concentrations of many of the elements considered decrease south/southeastward away from the Platte River Valley and Sand Hills of Nebraska. Superimposed on the overall trend is an increase in concentrations at sites adjacent major river valleys due to a 'refreshening effect' (Johnson and Muhs 1996).

Peoria loess deposition appears to relate to the formation of the Sand Hills of Nebraska. Kutzbach and Wright (1985) argue that the Sand Hills of Nebraska is the source for much of the Peorian loess in the region, and that formation of the huge transverse dunes within the Sand Hills must have occurred during a period of extreme aridity. Although strong circumstantial evidence links the Sand Hills to the adjacent body of Peoria loess, this hypothesis is not universally accepted (e.g., Ahlbrandt et al. 1983).

The question of the Sand Hills formation and Peoria loess deposition is germane to the investigation because, if the arguments of Kutzbach and Wright (1985) are correct, the Woodfordian, or at least some portion of it, was relatively xeric. This presumed Woodfordian aridity conflicts with at least some paleobotanical evidence in the region, particularly that related to the widespread occurrence of spruce and other taiga-like plant taxa (Table 1). These taxa, typically associated with the modern boreal forest, do not suggest moisture stress. The Woodfordian macrofossil record from the central Great Plains includes several well documented and dated occurrences of *Picea* remains, indicating a cool, mesic environments.

Leonard (1952) subdivided the Peoria loess of Kansas into four zones on the basis of the molluskan fauna assemblages present. The *basal zone* is equivalent to a leached interval above the Gilman Canyon Formation and is void of molluskan material. The *lower molluskan zone*, or Iowan, produced an assemblage containing 14 species, two of which are diagnostic of the zone. A *transitional zone*, located between the upper and lower faunal zones contains elements of both assemblages and does not imply any abrupt changes in the depositional environment, although the depositional rate may have slowed somewhat. The *upper molluskan zone*, or Tazewellian, contains 26 species, 14 of which do not occur in the lower zone. Because of the relative youth of the Peoria loess, little of the upper zone has been removed from the upland.

Although readily visible stratigraphic breaks, such as the Jules soil recognized in Illinois (Frye and Willman 1973; Ruhe 1976; McKay 1979) and the soil zones in Iowa (Ruhe et al. 1971), have not yet been widely identified in Kansas and Nebraska, evidence of one or more stable or vegetated surfaces is common. One of the few indications of soil development recognized is that of a Bt horizon in the Medicine Creek valley (May and Holen 1993); interestingly, the soil has a probable Paleoindian association (May 1990, 1991). Other indications of soil development in the Peoria loess come from the magnetic data obtained at Fort Riley, Kansas (Johnson 1996b).

Many of the age determinations were made from *Picea* remains, indicating a cool, moist environment. For example, a radiocarbon age of 18,830 yr B. P. on *Picea* charcoal was obtained from the Woodfordian/Peoria-age deposits near Bloomington, Franklin County, Nebraska (May and Holen 1993). Although radiocarbon data documents the burial of vegetative material throughout the Woodfordian, two temporal clusters of ages appear from the limited data: 18-17 ka and 14-13 ka. The 18-17 ka time interval represents the Last Glacial Maximum, and the 14-13 ka interval represents the time of major deglaciation (Ruddiman 1987). By interpreting ice-core data from Greenland, Paterson and Hammer (1987) recorded a dramatic decrease in atmospheric dust content from about 13 ka; this period of reduced atmospheric dust may relate to the time of relative surface stability and tree establishment. May (1989) identified deposition of the Todd Valley Formation in the South Loup River of central Nebraska at about 14 ka; the Todd Valley was subsequently buried by loess. Furthermore, Martin(1990) identified entrenchment in the Republican River of the south central Nebraska at about 13 ka, after which valleys were filled with late Peoria loess.

Eolian Sand Deposits

Perhaps the most controversial of the geologic aspects of the Sand Hills is the age of eolian activity that produced the dunes. Traditionally, they have been considered to be as old as the late Pleistocene Peoria loess. Lugn (1968), Wright et al. (1985) and others have hypothesized that deposition of loess and the development of large sand dunes occurred contemporaneously. Smith (1965) hypothesized two main periods of eolian activity in the late Pleistocene. The first period formed the large crescentic dunes, and, following a period of dune stabilization, the second episode rejuvenated the earlier dunes and produced linear dunes. Recent attempts to model climatic changes of the last 18,000 years (COHMAP Members 1988) have modeled the Sand Hills as dating primarily from the late Pleistocene. However, no stratigraphic evidence has been reported that would lend support to these hypotheses. Some eolian sand was deposited during the last glacial episode, but the sand dunes of the Nebraska Sand Hills appear to have formed during the episodes of aridity and eolian activity occurring within the Holocene (Ahlbrandt and Fryberger 1980).

Landscape stability has also been episodic on the Great Bend Sand Prairie in the last 20,000 years (Arbogast 1995). During the late Wisconsin, Peoria loess accumulated periodically, and the prevailing northwest winds created lunettes on southern margins of playa-lake basins. A brief period of soil formation occurred during the Last Glacial Maximum about 18 ka (Figure 7). Floral materials recovered from the late-Wisconsinan sediments suggests that the climate was more mesic than at present (Johnson 1991b).

Alluvial Deposits

Much of the chronology of late-Wisconsinan landform evolution for the region was compiled in the 1940s and early 1950s, prior to the use of radiocarbon dating (e.g., Lugn 1935; Schultz and Stout 1945; Frye and Leonard 1951), and focused to a large degree on the upland rather than valley deposits. Additionally, erosion has removed a large part of late-Quaternary record from most drainage basins in the central Great Plains (Knox 1983). Accordingly, a comprehensive sedimentation and erosion chronology for the region during the late Wisconsin is lacking.

It is becoming increasingly apparent that entrenchment occurred in the channels of the Kansas River basin sometime during the late Wisconsin. A basal soil buried within the fill of both tributary and major stream valleys of the Kansas River basin has an age of 10,500-10,000 yr B.P. (Johnson and

Martin 1987; Johnson 1987; Johnson and Logan 1990), thereby providing a minimum age on the entrenchment. May (1989) has radiocarbon dated the Todd valley fluvial sand of central Nebraska to about 14,000 yr B.P., although Condra et al. (1950) had postulated a much earlier Wisconsinan age. Martin (1990; 1993) recognized a late-Wisconsin fill in the Republican River valley that was largely removed through entrenchment about 13,000 yr B.P. A radiocarbon age of 14,700 yr B.P. was obtained on spruce wood situated above crossbedded fluvial sand and gravel at the North Cove site located in that same reach of the Republican River valley (Wells and Stewart 1987b; Johnson 1989). At the Prairie Dog Bay site in the Republican River valley, the stratigraphy and radiocarbon ages suggest downcutting before 11,800 yr B.P. (Martin 1990, 1993). Speculation about the cause of this entrenchment centers on an increase in effective moisture as climatic conditions ameliorated towards the end of the Pleistocene. Spring deposits dating to this time at the North Cove site possibly formed during the increase in moisture (Johnson 1989, Martin 1990).

Brice (1964) studied alluvial fills and terraces in the valleys of the North Loup, Middle Loup, and South Loup rivers of central Nebraska, and identified two major terraces in the Loup valleys. The Kilgore terrace occurs as remnants 26 to 30 m above stream level along the South Loup River. Brice suggested that valley fill underlying the Kilgore terrace is Peorian (late Wisconsin) in age. The adjacent Elba Terrace, which stands 11 to 12 m above stream level, is the most prominent and extensive terrace in the main valleys, with fill dating to the late Pleistocene and Holocene.

Climatic Proxies

Two general quantitative methods have been applied to the reconstruction of past climates. The first is to determine past climate through the analysis of local or regional field data with the aid of transfer functions. The other method uses large-area climate modeling with the boundary conditions determined by calculation or from field data. Neither supplants the other, for the reconstructions have different spatial scales and degrees of precision. Most models of past climates also require inputs that can only be obtained from field investigations (Smiley et al. 1991). Transfer functions refer to a quantitative relation between a climatic indicator, such as $\delta^{13}C$ data from buried soils, as an independent variable, and a climatic element or complex of elements, expressed as a dependent variable. The use of analogs for estimating past climates involves considerable uncertainty, brought about both by the complex mix of factors that constitute climate and by the complex response of most proxy climatic indicators in the record. In a sense, the use of analogues involves the construction of a mental transfer function based on the assumption of appropriate modern analogue selection (Smiley et al. 1991). Because each source of paleoenvironmental data records a somewhat different aspect of climate, comparing reconstructions based on two or more environmental sensors can broaden and deepen our understanding of past climate changes.

With few suitable species and settings for dendroclimatological study and few natural wet environments to preserve fossil pollen for palynological studies, the nature and timing of late-Quaternary vegetation and climate change in the central Great Plains remain poorly understood. The region's late-Quaternary climate and vegetation conditions are inferred from the palynological records obtained from sites peripheral to the central Great Plains or from limited pollen records available at a limited number of sites in the region (Fredlund and Jaumann 1987; Fredlund 1995). However, some of the climatically-sensitive parameters that have recently been examined in the central Great Plains include fossil pollen, opal phytoliths, stable carbon isotopes, and rock magnetism. In addition to an expanding proxy base, recent research has indicated that the extensive loess deposits of the region contain an extractable climatic proxy record comparable to the marine isotopic record.

Fossil Pollen and Botanical Macrofossils

Several factors in the interpretation of Great Plains fossil pollen assemblages warrant consideration. Any interpretation of pollen assemblages for vegetational reconstruction must be based on appropriate analog studies of modern vegetation and pollen (Fredlund and Jaumann 1987). Additionally, in the central Great Plains region where ideal wet depositional sites are rare, differential pollen preservation is a problem. Modern analogs are a basis for late-Quaternary environmental reconstruction only where pollen deterioration has not significantly biased the informational content of the fossil pollen assemblage (Delcourt and Delcourt 1983). For example, differential preservation has been shown to be responsible for tremendous over representation of *Pinus* in some situations, while elsewhere rendering *Populus* invisible. Poor pollen preservation is therefore the limiting factor for many of the late-Quaternary records in the central Great Plains. Although temporally and spatially limited, several sites in the region have produced a picture of past environments (Figure 8).

By the time ice lobes in Iowa and the Dakota had reached their maxima at about 14ka (Clayton and Moran 1982), *Picea* had begun to spread its range northward into the Des Moines area (Baker and Waln 1985). By 12ka, spruce forest was replaced along its southern margin by prairie in southern South Dakota. About 11ka, *Quercus* (oak), *Populus, Fraxinus* (ash), and other hardwoods, which were probably confined to the central Unites Stated in glacial time, expanded their northern ranges and mixed with *Picea* in the eastern part of the northern Great Plains and the Midwest. This admixture of trees has no close analogues in the present day, but the vegetation is presumed to have been open and dominated by spruce, with some hardwoods and no pine.

The western limit of the forest is not known. At two sites in the glaciated region of northeastern South Dakota (Pickerel Lake: Watts and Bright 1968; Medicine Lake: Radle 1981), deciduous tree pollen replaced that of *Picea* about 11ka, and prairie taxa appear at about 10ka. Pollen data from east-central North Dakota indicate a similar sequence. From the eastern fringe of the central Great Plains, in Iowa and Missouri, pollen and macrofossil evidence suggests that open jack-pine forest of the Farmdalian period yielded rapidly to open white-spruce forest around 22ka (Fredlund and Jaumann 1987). A similar record of Woodfordian spruce forest comes from Muscotah Marsh in northeastern Kansas. According to Grüger (1973), a somewhat open vegetation, with pine, spruce, and birch as the most important tree species and local stands of alder and willow, changed about 23,000 yr B.P. to a spruce forest, which prevailed in the region until at least 15,000 yr B.P. Because of a hiatus in the sedimentary record, vegetation changes resulting in the spread of a mixed deciduous forest and prairie present in the region from 11,000 to 9,000 yr B.P. remain unknown.

According to Wells and Stewart (1987b), the central and northern Rocky Mountains harbor extant populations of most of the boreal-subalpine species thus far recovered from Pleistocene sediments in the central Great Plains. Moreover, even within the Northern Plains, there are numerous refuges for Pleistocene-relict species of trees, landsnails and small mammals on forested ecological islands surrounded by steppe, an outstanding example being the Black Hills of South Dakota. Cones and needles from Harlan County, south-central Nebraska, enable the positive identification of the spruce as *Picea glauca* (Johnson 1989), the boreal white spruce of the neoarctic taiga that now grows from Alaska to Newfoundland and along the eastern flank of the Rocky Mountains to Montana, with outliers to the East on the Great Plains in the Cypress Hills of Saskatchewan and Black Hills of South Dakota.

The Rosebud site, near the northern edge of the Sand Hills on the Nebraska-South Dakota border, provides a pollen record of late-Pleistocene vegetation. The pollen and plant macrofossil records indicate that a boreal forest existed at that location about 12,600 yr B.P., and that soon afterward a pine forest and subsequently prairie vegetation rapidly replaced the spruce (Watts and Wright 1966). Seeds and leaves of aquatic macrophytes at the site suggest that a fresh, open-water basin existed when spruce was prevalent, and that conditions changed to a species-poor, alkaline reed swamp with the change to prairie vegetation. This vegetation and limnologic history implies change from a cooler, probably somewhat moister climate to one of increased aridity and higher temperatures that characterizes the Sand Hills today. The pollen record of prairie vegetation at Rosebud does not significantly differ from that of modern surface samples in this area. The rapid disappearance of *Picea* pollen and its immediate replacement by *Pinus* and prairie herb pollen suggest a depositional hiatus, which makes it difficult to interpret subsequent vegetational history. It is clear is that prairie vegetation existed sometime after 12,600 yr B.P., and that the lake subsequently dried up and either the upper pollen-bearing sediments were destroyed, or intermittent fluvial deposition with poor pollen preservation occurred.

The association of nonarboreal taxa from sand pits near Wichita, Kansas indicates a substantial presence of steppe or grassland taxa on the late-Pleistocene landscape of south-central Kansas (Jaumann 1991). Not only do these taxa represent a significant portion of the pollen spectra, but they also occur in consistent numbers and presently comprise the most important herbaceous taxa of the North American grasslands. Some of these taxa include Graminaeae (grass family), Asteraceae (sunflower family), *Artemisia* (sage), *Iva* cf. *xanthifolia* (marsh elder), *Xanthium* (cocklebur), *Amorpha* cf. *canescens* (lead plant), *Phlox* cf. *pilosa* (prairie phlox), *Petalostemon* cf. *purpureus* (purple prairie clover), *Potentilla sp.* (cinquefoil), *Ambrosia type* (ragweed), *Chenopodium/Amaranthus* (goosefoot, pigweed), herbaceous Rosaceae (rose family), Fabaceae (bean family), *Epilobium* cf. *angustifolium* (willow-herb), Euphorbiaceae (spurge family), Cannabaceae (hemp family), and *Tradescantia* (spiderwort).

The closest vegetation type showing such a compositional mix can be found along the southern rim of the boreal forest on the Canadian Prairies. Mapping of the southern limits of coniferous trees indicate that the southern natural distribution of *Picea glauca, Picea marina* (black spruce), *Larix laricina* (tamarack), *Pinus banksiana* (jack pine), and *Pinus contorta* (limber pine) is confined to a narrow transitional zone between the taiga and aspen parkland (Jaumann 1991). Mosaics of grasslands and forests characterize the aspen parkland. The fossil plant communities recorded in the Wichita sand pits and Mt. Hope Sand Company pit pollen assemblages look very much like the vegetation types in this narrow transitional zone, which prominently extend eastward into the prairie or aspen parkland.

According to Fredlund (1995), the high relative frequency of *Artemisia* pollen in the Farmdalian record from Cheyenne Bottoms in central Kansas indicates that one or more species of sage were an extremely important element in the upland grassland-steppe. This vegetation assemblage does not, however, appear to be exactly analogous to the modern sagebrush steppe of the northwestern High Plains. The pollen evidence suggests that the regional vegetation, although dominated by grassland-steppe, was not totally treeless. Most of the arboreal elements present are boreal or taiga-like in their modern distribution. The most common trees of the Pleistocene vegetation in the region were not, however, coniferous. The low percentages of both *Picea* and *Pinus* pollen could be the result solely of long-distance transportation; this is especially likely for *Pinus* which could represent forests as far away as 400 km. In the case of *Picea,* however, it is more likely that local populations of trees were scattered along river valleys or fire-protected escarpments. It is extremely unlikely that the *Pinus* and *Picea* pollen signals from the Farmdalian portion of the record represent coniferous parklands or savannas, rather it is more likely that these low pollen percentages represent small populations of conifers limited to edaphically mesic and fire-protected situations.

Opal Phytoliths

Growing plants absorb water containing dissolved silica through their roots. Microscopic amorphous silica bodies are subsequently produced by the precipitation of hydrated silicon dioxide ($SiO_2 \cdot nH_2O$) within the plant's cells, cell walls, and intercellular spaces. Silica bodies with shapes characteristic of specific plants or group of plants are called opal phytoliths. Phytolith is derived from the Greek words *phyton*, meaning plant, and *lithos*, meaning stone, and opal is the common name for hydrated silicon dioxide. Opaline bodies formed in plants without specific shapes are simply plant opal or biogenic opal.

It is well known that three different photosynthetic pathways exist among plant species: C_3 (Calvin-Benson cycle), C_4 (Hatch-Slack cycle) and CAM (Crassulacean Acid Metabolism). Twiss (1987) suggested that grass-opal phytoliths could serve as indicators of C_3 and C_4 pathways in grasses. On the Great Plains, two grass subfamilies commonly employ the C_4 pathway: the Panicoideae and the Eragrostoideae. The panicoids include such common prairie grasses as the bluestems (*Andropogon* spp.), panicums (*Panicum* spp.), and Indian grass (*Sorghastrum nutans*), as well as domesticated grasses, e.g., sorghum and corn. The grama grasses (*Bouteloua* spp.) and buffalo-grass (*Buchloe doctyloides)* of the Chloridoideae tribe of the Eragostoideae subfamily are the two most important of these grasses in the arid southwestern region of the Great Plains. Pooideae grasses such as the bromes (*Bromus* spp.), wheatgrass (*Agropyron* spp.), fescues (*Festuca* spp.), and many of the cereal grains, including wheat (*Thriticum aestivum*) and oats (*Avena* spp.), are C_3 pathway types. The overall pattern, where pooids (C_3) dominate the cool north-central Great Plains, panicoids on the moist, warm eastern and southeastern margins, and chloridoids primarily in the western and southwestern Great Plains, is consistent with the pattern expected from general C_3 and C_4 adaptations of grasses.

Few workers have reported opal phytolith data from sites in the central Great Plains. Among them, Fredlund et al. (1985) tabulated the abundance and type of phytoliths from a vertical loess section at the Eustis ash pit in south-central Nebraska. Pooid phytoliths were the most abundant forms, followed by significant vertical variation in the chloridoid and panicoid types. They concluded that increases in the chloridoid type in paleosol complexes indicated that the soil forming periods must have been warmer and drier than the periods of loess accumulation. The phytolith assemblages from the soil of the Gilman Canyon Formation is unique at the Eustis ash pit: nowhere in the entire 620,000-year record of loess accumulation at the site has anything similar been recorded. The high relative frequencies of panicoid-class phytoliths are even higher than those found in the tall-grass, panicoid-dominated prairies today. In general, the phytolith evidence of warmer soil-forming periods and cooler episodes of increased dust accumulation fits the traditionally accepted models for loess deposition and other proxy records.

Ongoing opal phytolith analysis of the Peoria loess is producing a climatic signal consistent with that of the carbon isotope data (Johnson et al. 1993a). Phytolith data can be represented as a composite parameter, the aridity index (Figure 9). A cool, mesic climate is apparent by the occurrence of arboreal phytolith types and C_3 grass types in the loess of the lower Gilman Canyon Formation and the Peoria loess.

Stable Carbon Isotopes

There are few quantitative techniques in use today for paleoecological reconstructions in terrestrial depositional systems. One recently adopted approach to quantitative reconstructions is to estimate the proportion of C_3 (cool-season) to C_4 (warm-season) plants once present at a site using carbon isotopes from the bulk carbon content in sediments, primarily in buried soils.

The natural difference in the stable carbon isotopic composition of C_3 and C_4 plant species provides an opportunity to assess the long-term stability of plant communities and climate of a given region (Troughton et al. 1974, Stout et al. 1975). The basis of this approach is that during photosynthesis C_4 plants discriminate less against $^{13}CO_2$ than C_3 plants (Vogel 1980; O'Leary 1981). This difference in carbon isotope fractionation during photosynthesis results in a characteristic carbon isotope ratio in plant tissue that serves as a diagnostic indicator for the occurrence of C_3 and C_4 photosynthesis. The $\delta^{13}C$ values of C_3 plant species range from approximately -32 to -20‰, with a mean of -27‰, whereas $\delta^{13}C$ values of C_4 species range from -17 to -9‰, with a mean of -13‰. Thus, C_3 and C_4 plant species have distinct, non-overlapping $\delta^{13}C$ values and differ from each other by approximately 14‰ (Nordt et al. 1994).

The stable isotope ratios for $^{12}C/^{13}C$ are measured by mass spectrometry, and the isotopic data are expressed as the difference, or delta value (δ), between the sample or standard. The δ value for a carbon isotope in soil is defined as

$$\delta^{13}C\ \text{soil} = (\delta^{13}C_{C4})(x) + (\delta^{13}C_{C3})(1-x),$$

where $\delta^{13}C_{C4}$ is the average of $\delta^{13}C$ values of C_4 plants (-13‰); ($\delta^{13}C_{C3}$) is the average of $\delta^{13}C$ values of C_3 plants (-27‰); and x is the proportion of carbon from C_4 plant sources. Isotopic composition of soil organic matter or pedogenic carbonate in soils with a high-respiration rate is a direct indicator of the fraction of the biomass using the C_3 or C_4 photosynthetic pathways. Humus from buried soils probably represents organic matter from the last few hundred years before burial, given the short residence times typical for humus in most modern soils (Birkeland 1984).

Teeri and Stowe (1976) found that the strongest correlation with the percent C_4 in the continental United States was given by the normal July daily minimum temperature, with a correlation coefficient of 0.97. This temperature was a better predictor of the percent C_4 than either the normal July average temperature or the normal July maximum temperature. Based on their analysis, the percent of C_4 species in a grass flora in the continental United States is most accurately predicted by a linear combination of the normal July minimum temperature, mean annual degree-days and the log of the length of the freeze-free period.

For the Gilman Canyon Formation, $\delta^{13}C$ values exhibit a good correlation with coincident phytolith data (Figure 9). $\delta^{13}C$ data acquired in association with the correction of radiocarbon ages for the Peoria loess in Kansas and Nebraska indicate that C_3 plants were dominant during most of Peoria loess deposition (Figures 10 and 11). This reflects the cooling associated with the Last Glacial Maximum within early-middle Peoria time (ca. 18 ka). Conversely, C_4 plants were dominant for most of the Gilman Canyon time of pedogenesis (Figure 9), indicating that vegetation and thus climate during Gilman Canyon time was similar to present warm, semiarid conditions in the central Great Plains (Johnson 1993b).

Site specific factors should be borne in mind when interpreting $\delta^{13}C$ data from soil humates. For example, the 17,000 yr B.P. buried soil on the north flank and crest of the dune at Wilson Ridge in the Great Bend Sand Prairie (17,180 ± 240, Tx-7824: 16,520 ± 200, Tx-7825) yielded a $\delta^{13}C$ value of -11.9‰ (Figure 11). During the Last Glacial Maximum, the dune temporarily stabilized and a soil formed. The $\delta^{13}C$ value suggests that warm-season or edaphic plants dominated, a finding contradictory with regional late-Wisconsinan mesic climatic conditions. Following landscape stability, the soil was buried by sand, presumably during another period of increased aridity and prevailing northwesterly winds (Arbogast 1995).

Rock Magnetism

Loess is perhaps the closest terrestrial analogue to marine sediments in that both result from more-or-less continuous deposition of fine-grained sediment. A great deal of attention, therefore, has been given to magnetic measurements of loess-paleosol sequences, particularly in China and in Europe (e.g., An et al. 1990, 1991; Kukla et al. 1988; Kukla 1977). Although much research has focused on paleomagnetic events such as excursions, recent studies in China have suggested that bulk magnetic susceptibility varies systematically in loess sections and can be visually correlated with the well-known marine oxygen curve (Kukla 1987), indicating that changes in the magnetic susceptibility of loess constitute a terrestrial proxy climatic signal. Further direct relationships between magnetic susceptibility and oxygen isotope variations have recently been demonstrated within some marine cores (Heller et al. 1991). Ongoing research is demonstrating that rock magnetism may be used to successfully reconstruct the climatic sequences of the central Great Plains (e.g., Park et al. 1993; Farr et al. 1993; Johnson and Park, 1996).

Magnetic susceptibility measures magnetization temporarily induced in a soil or loess by an artificially applied, low-amplitude magnetic field. The strength of the susceptibility signal depends on the concentration and grain size of the magnetic minerals. Magnetic susceptibility intensities measured at the Eustis ash pit and Bignell Hill sites in southwestern Nebraska, the Beisel-Steinle site in north-central Kansas, and Barton County landfill site in central Kansas indicate that magnetic intensities are strong in the Gilman Canyon Formation soil, i.e., susceptibilities are nearly twice that of the unweathered Peoria loess (Figure 12). Weaker intensities associated with the Peoria loess indicate that either the depositional rate was faster than in times of soil development or that biologic/soil forming activity was weaker in Peoria time, presumably the latter. Susceptibility curves from the region correlation well despite the distance from each other, testifying to the regional synchroneity and direction of change in climate.

Magnetic analysis from the Eustis ash pit produced a climatic signal consistent with that of the carbon isotope data (Figure 13). Weaker intensities throughout the Peoria time likely reflects the more moist and cooler C_3 plant types and the cooling associated with the Last Glacial Maximum. The terrestrial plant ecology of the Gilman Canyon Formation apparently is characterized by primarily C_4-

type grasses, or warm, possibly dry climate (Johnson 1993b). From the magnetic susceptibility data, it is evident that an environment suitable for more active pedogenesis existed .

Recent research on the Fort Riley military reservation in east-central Kansas has produced magnetic data supporting the regional late-Quaternary climatic model. The frequency dependence data extracted from a 13m-thick loess mantle on the bluff top adjacent to the Kansas River valley exhibits the higher values anticipated for soils and lower values from the relatively unweathered loess, as well as good correlation to the $\delta^{13}C$ curve (Figure 14). The Manhattan Airport site consists of Pleistocene and Holocene alluvial fill within a high terrace of the Kansas River valley. The bulk of the fill at this site is Pleistocene, but, on the basis of a single radiocarbon age, the upper 1 to 1.5 m is Holocene (Figure 15). Physical attributes from about 2.75m to the bottom of the trench indicated the Sangamon soil, but the relatively weak magnetic signal, particularly for susceptibility, is probably due to poor drainage conditions of the alluvial surface during that time. The Gilman Canyon soil does not appear magnetically, but the 19,990 yr B.P. age, a terminal Gilman Canyon age, should identify the top of that soil forming period. Soils present presumably represent the alluvial, or valley phases of those expressed regionally on the uplands (e.g., Sangamon, Gilman Canyon).

Climatic Modeling

Kutzbach (1987) summarized the behavior of the North American jet streams during the late Wisconsin. In July, the split flow around the North American ice sheet modeled at 18 ka persisted to 15 ka, with almost no changes having occurred. By 12 ka, two important changes appeared: first, the northern branch of the jet moved south, over and along the southern flank of the ice sheet, and merged with the southern branch over the northeastern United States. The second change was the reduced intensity of the North Atlantic extension of jets. The jet at 9 ka followed about the same track as at 12 ka, but with weakened intensity. At 6 ka and thereafter, only a single jet core was simulated over Alaska and Canada, and winds were weak compared to reconstructions of earlier periods. Specifically, the modern single jet core of July follows generally the same track as the northern branch of the split jet in July during the glacial maximum.

In January, like July, the split flow around the North American ice sheet and the intense North America/North Atlantic jet cores at 18 ka persisted to 15 ka with almost no change. At 18 ka the simulated temperatures over the continent were much lower than at present, especially over the elevated and highly reflective ice sheets (COHMAP Members 1988). By 12 ka, the flow had adjusted to single core of high velocity winds that followed the west-coast-ridge, east-coast-trough pattern of today, and the jet maximum was almost as strong as at 18-15 ka.

The north central region, from the Rockies to the Appalachians and from immediately south of the ice sheet to 40°N, had summer temperatures of 16°C at 18-15 ka, about 7°C below present. Precipitation was less than present at 18-15 ka (colder, with storm track shifted south of the region) and precipitation-minus-evaporation was slightly increased at 18-15 ka (evaporation decreased more than precipitation).

PLEISTOCENE/HOLOCENE TRANSITION

The last deglaciation was a period of intense and rapid climatic changes that affected the global climate from about 20,000 to 5,000 yr B.P. Paleoclimatologists have reconstructed global variations, including chemical composition of the atmosphere (30% increase in CO_2 and CH_4, decrease in dust content, etc.), temperature of the atmosphere and surface of the ocean (mean global change of about +4°C), and major reorganization of the ocean circulation and sea-level rise of about 120 meters, followed by slow rebound of the continents below the ice caps.

The transition between the Last Glacial Maximum and the present inter- or postglacial episode has drawn much attention from investigators for many decades. At first, the last deglaciation was believed to have been a simple, unidirectional shift, but more recent detailed studies revealed that it was a two-step process (Duplessy et al. 1981; Broecker et al. 1989). During the last deglaciation, intervals of rapid warming between about 13 ka and 11 ka and at about 10 ka were separated by a distinct, brief, cool climate episode occurring between about 11 ka and 10 ka.

Between about 12ka and 9ka, the climate and vegetation of central North America underwent dramatic changes (Wright 1970; Watts 1983; Webb et al. 1983). Spruce trees had been replaced by widely distributed deciduous trees in northeastern Kansas, and deciduous trees persisted until about

9ka when grasslands expanded (Webb et al. 1983). It is clear that megafaunal extinction and dissolution of disharmonious faunas began about 12ka, and the mesic conditions under which the regionally-expressed Brady soil developed persisted until about 8ka, when the modern climate first appeared. Changes in vegetation and faunal assemblages at this time reflect a shift to warmer and drier conditions with increased seasonality (COHMAP Members 1988) and stronger zonal air flow at the surface (Kutzbach 1987). This was a time of major atmospheric circulation change within the central Great Plains, as well as elsewhere.

Geomorphology and Stratigraphy

The beginning of the Holocene, about 10 ka (Hopkins 1975), is a time of dramatic environmental change and attendant stratigraphic discontinuities. In general, this boundary is considered only geochronometric without specific stratigraphic reference, although a stratotype in Sweden has been proposed for the boundary (Mörner 1976); the Swedish unit has a reported age of 10,000±250 yr B.P. (Fairbridge 1983). According to Richmond and Fullerton (1986), a stratigraphic boundary of regional extent of the Pleistocene-Holocene boundary age has not been identified in the United States, and that major climatic or environmental changes at 10,000 yr B.P. are documented only locally (Watson and Wright 1980). This contention seems faulty, at least on the regional scale in that research of the last several years in the central Great Plains has identified the Brady soil (Schultz and Stout 1948) as a major pedostratigraphic marker (e.g., Johnson and Martin 1987, Johnson and Logan 1990, Johnson and May 1992).

Brady soil

Classically, the Brady soil was associated with the upland loess deposits, but recent investigations have identified a contemporaneous soil in upland eolian sands and in alluvial valley fill (Johnson and May 1992). It therefore appears that the Brady soil development represents a time of extensive, broad-scale landscape stability. The Brady soil represents the most important break in the sedimentation recorded since development of the cumulic soil of the Gilman Canyon Formation, and also marks the position of a distinct faunal discordance (Frye and Leonard 1955). At least the early and perhaps all of the Brady soil-forming interval coincides with the Younger Dryas cold interval of the North Atlantic region.

The Brady soil was first named and described by Schultz and Stout (1948) at the Bignell Hill type locality, a loess sequence exposed along a roadcut in the south valley wall of the Platte River of western Nebraska. The soil is developed within the Peoria loess and is overlain by the Bignell loess. The name was subsequently adopted by researchers in Kansas (Frye and Fent, 1947; Frye and Leonard, 1951; Frye et al., 1949). The soil is regionally extensive only in the northwestern and west central parts of Kansas, but even there it occurs discontinuously on the landscape. Frye and Leonard (1951) and Caspall (1970, 1972) recognized Brady development in northeastern and other parts of Kansas. Without the overlying Bignell loess, the Brady soil does not exist; the modern surface soil has incorporated post-Bradyan loess fall into its profile. The Brady soil is typically dark gray to gray-brown and better developed than the overlying surface soil within the Bignell loess. Strong textural B horizon development and carbonate accumulation in the C horizon are typical, although it occasionally displays evidence of having formed under poorer drainage conditions than have associated surface soils (Frye and Leonard, 1951). Feng (1991) noted that the Brady soil, as expressed in Barton County, is strongly weathered both physically and chemically.

Until recently the age of the Brady soil had been uncertain, even at the type section: Dreeszen (1970) reported two ages of 9160 and 9750 yr B.P., both of which were believed to be too young because of contamination, and Lutenegger (1985) reported an age of 8080 yr B.P. without any stratigraphic context. Johnson (1993a) reported two ages of 10,670 and 9240 yr B.P. on the lower and upper 5 cm, respectively, of the Brady A horizon at the type section. Souders and Kuzila (1990) dated a core at a site in the Republican River valley and reported an age of 10,130 yr B.P. Similar ages from the eolian phase have been obtained in south-central Nebraska and north-central and central Kansas (Table 2). According to age data, soil development began at about 10.5 ka and ended 9-9.5 ka, suggesting a soil forming interval of greater than 1000 years.

Arbogast (1995) obtained several Brady era radiocarbon ages from soils buried within the eolian sand of the Great Bend Sand Prairie. Following a period of instability after a short period of stability during the Last Glacial Maximum, soil formation occurred at the Pleistocene/ Holocene boundary, which correlates temporally with the loessal Brady soil. Two radiocarbon ages of 10,330 ± 100 and 10,360 ± 100 yr B.P. were obtained at Wilson Ridge, a lunette in the Great Bend Sand Prairie.

The Brady soil is also well expressed in an alluvial facies, i.e., an isochronous alluvial soil found throughout the region is temporally equivalent to the Brady soil identified within loess of the uplands. Since a large number of radiocarbon ages have been obtained from alluvial fill in the central Great Plains (Johnson et al. 1996), the patterns of alluviation, erosion and particularly soil formation during the late Pleistocene through Holocene have become relatively well established.

In northwestern Nebraska, Agenbroad (1978) examined alluvial deposits at the Hudson-Meng Site (25SX115) in Whitehead Creek Valley. Four stratigraphic units were identified beneath a loess-mantled terrace. The lowermost unit, designated Unit 1, consists of alluvium and contains a bone bed and many artifacts; charcoal from the bone bed yielded a radiocarbon age of 9820 yr B.P., and bone apatite and collagen yielded ages of 8990 and 9380 yr B.P., respectively. Agenbroad suggested that the site was buried by alluvial sands and silts sometime after about 9,000 yr B.P., and that a soil developed at the top of Unit I (ca. 4800 yr B.P.) during Altithermal time.

The two ages of 8,274 yr B.P. and 9,880 yr B.P. determined from alluvial fill (Fill 2A) at archaeological sites Ft-50 and Ft-41 on Harry Strunk lake in southwestern Nebraska (Schultz et al 1951; Libby 1955) were the first radiocarbon determinations obtained on the Brady soil. At Cooper's Canyon, which is southeast of Elba, Nebraska, May (1990, 1991), in a reinvestigation of a site studies by Brice (1964), reported radiocarbon ages on humates in the silt and clay fractions of buried soils, of which two ages agree well with both those in the fill from other localities in the Loup River Basin and eolian facies. Ages included, 10,290 yr B.P. from the lowest 10 cm of Cooper's Canyon gley soil, and 9250 yr B.P.on the uppermost 10 cm of the Brady soil (Figure 16).

In Kansas, Holien (1982) derived a radiocarbon age of about 10.5 ka from a well developed soil situated in the lower part of Newman Terrace fill along the lower Kansas River. Johnson et al. (1996) obtained an age of 9,820 yr B.P. from a soil buried in alluvial fill at the Ade site within the Saline River valley near Salina, Kansas.

Clovis Level Soil of the Colorado High Plains

The Dutton and Selby sites lie on the High Plains of eastern Colorado in shallow, internally drained, surface depressions. The Dutton site contains a Clovis level between two clay-rich paleosols formed in reworked Peoria loess that fills the depression. A collagen age of 11,710 ± 150 yr B.P. on mammoth bone from near the E-B horizon boundary of the lower soil may be related to the Clovis level (Reider 1990). Camel bone ages from the bone bed in Peoria loess, in conjunction with the Clovis or perhaps largely pre-Clovis level, provide evidence that the lower soil formed between about 13,600 yr B.P. and the time of Clovis occupation (ca. 11,400 yr B.P.). Both paleosols are regarded as late Pleistocene, but the upper soil (and to a lesser extent the lower soil) were altered by Holocene pedogenesis. Stratigraphic correlation with the Brady soil has not, however, been made at this point.

Climatic Proxies

Fossil Pollen and Botanical Macrofossils

The most detailed description of the nature of Late Pleistocene/Holocene environmental changes in the central Great Plains comes from palynological studies undertaken along the eastern and northern periphery of the region. At Muscotah and Arlington Marshes in northeastern Kansas, Grüger (1973) documented spruce forest from 23,000 to 15,000 yr B.P. followed by the spread of a mixed deciduous forest and prairie, which was present in the region from 11,000 to 9,000 yr B.P. The nature and duration of the climatic changes which precipitated vegetation changes are not, however, certain because of a hiatus in sedimentation. Fredlund and Jaumann (1987) suggested that pollen records represent an expansion of an aspen parkland-like community across the Great Plains.

According to Wright (1989), pollen records from the Great Plains can not show the effects of minor climatic fluctuations like the Younger Dryas because climate had become too warm by 11 ka to permit introduction of spruce. General circulation model results also show that the temperature for winter was deeply depressed far across Eurasia but was little changed in North America (Mathewes et al. 1993). The critical vegetation change identified by Shane and Anderson (1993) in east-central North America involves the recurrence of spruce, which is limited in its southern range by summer rather than winter temperatures. The southerly position of the polar front across the North Atlantic could have resulted in a southward displacement of the jet stream and associated storm tracks, thus enhancing the cyclonic storms that could deliver cold northwesterly winds, not only to the Maritime Provinces, but inland to the Ohio area as well (Wright 1989).

Another source of paleoenvironmental information comes from peat beds and logs, radiocarbon dated from 10,500 to 8,400 yr B.P., buried in valley fills associated with the North Loup River (Bradbury 1980). The peat is buried by alluvium which is in turn mantled by dune sand. The stratigraphic association of these deposits and the presence of marsh plants like *Equisetum* (horsetail) indicate that locally, fluvial processes and riparian environments, similar to those that exist today, were followed by sand movement (Bradbury 1980). Most recently, Ponte et al. (1994) dated a peat recovered in a core from the central Sand Hills and radiocarbon dated it to 12,260 yr B.P.; the peat contained 70% *Picea* pollen, indicating that the spruce forests of the late Wisconsin existed farther south into the Sand Hills than previously reported.

Stable Carbon Isotopes

Temporal changes in $\delta^{13}C$ data derived from carbon contained within soil and sediment (Figures 10, 11, and 14) are sufficiently large to show major shifts in vegetation during the late Wisconsin. The interval between 12,000 and 9000 yr B. P. can be interpreted as transitional between the cooler and more xeric late Pleistocene to warmer and drier Holocene. Based on a slight decrease in the $\delta^{13}C$ values from the Brady soil at six sites in the region, climatic conditions shifted to more xeric conditions (C_3 to C_4) from the beginning to the end of the Brady time, a period of major landscape stability and pedogenesis (Table 3).

The isotopic data agree with that of other climatic proxies for the region. The fossil pollen record from Muscotah Marsh in northeastern Kansas indicates that spruce had essentially disappeared from the region by about 10,500 yr B.P. As this decline occurred, deciduous tree species increased until about 9,000 yr B.P. From a site in central Texas, Nordt et al. (1994) interpreted the time between 11,000 and 8000 yr B.P. as transitional between late-Pleistocene conditions and warmer and drier Holocene conditions based on a slight increase in the abundance of C_4 plant biomass using stable carbon isotopic data.

Rock Magnetism

The Eustis ash pit, Beisel-Steinle site and Barton County landfill site each produced magnetic susceptibility curves characterized by a pronounced increase in the upper Peoria loess as the depositional rate decreased dramatically and Brady pedogenesis began (Figure 12). Frequency dependence of susceptibility exhibited a notable but not dramatic increase in the basal Brady soil for the Sumner Hill site on Fort Riley, Kansas (Figure 14), suggesting perhaps that the intensity of Brady pedogenesis varied spatially according to microclimatic conditions.

Magnetic susceptibility intensities measured at the Bignell Hill Site in Nebraska and the Beisel-Steinle and Barton County landfill sites in central Kansas indicate that magnetic intensities are very high in the Brady soil, e.g., susceptibility intensities ($80\text{-}100 \times 10^{-8} m^3/kg$) are nearly twice that of the unweathered Peoria loess ($40\text{-}50 \times 10^{-8} m^3/kg$) and slightly higher than the modern soil (Figure 17). On a hemispheric scale, the abrupt decrease in atmospheric dust noted in the Greenland ice core at about 10,750 yr B.P. (Paterson and Hammer 1987) reflects decreased loess transportation and deposition, and probably increased Brady-age pedogenesis associated with relative terrestrial stability.

Climatic Modeling

Significant deglaciation did not begin until 14 ka and ended by 6 ka. This conclusion is validated by maps of ice area, by marine $\delta^{18}O$ records, and by terrestrial and marine records (Ruddiman 1987; Crowley and North 1991). With increased summer insolation during the termination, the mass imbalance of ice sheet would have increased. Ice sheet decay may also have been affected by a number of processes. For example, CO_2-induced air temperature changes were apparently sufficiently large to cause disintegration of the extensive marine-based ice sheet on Eurasia. Broecker et al. (1988) suggested that changes in the coupled ocean-atmosphere circulation in the North Atlantic were responsible for the changes.

The structure of deglaciation within this 8,000-year interval is uncertain. There is evidence supporting: (1) a smooth deglaciation model with fastest ice wastage centered on 11 ka; (2) a two-step deglaciation model with rapid ice wasting from 14 to 12 ka and 10 to 7 ka, and a mid-deglacial pause with little or no ice disintegration from 12 to 10 ka; and (3) a Younger Dryas deglaciation model with two rapid deglacial steps as in (2) above, interrupted by a mid-deglacial reversal with significant ice growth from 11 to 10 ka.

The critical data supporting the smooth deglaciation model are maps of Laurentide ice area based on radiocarbon-dated glacial deposits. Although there are subtle suggestions of more rapid retreat at or near the time of the two steps mentioned above, these curves indicate a steady progressive retreat of North American ice, with significant oscillations in retreat rate only at local spatial scales. Some marine $\delta^{18}O$ curves also show a smooth progressive decrease toward Holocene values.

The step deglaciation model is also supported by some marine $\delta^{18}O$ records (Mix 1987). In addition, the distinctive patterns of change in sea-surface temperature of the North Atlantic Ocean and in Greenland ice-core $\delta^{18}O$ values also show abrupt step-like warmings at 10 ka and approximately 13 ka; these warmings might be associated with step-like decreases in Laurentide ice volume. Regionally integrated rates of pollen change in eastern and central North America also show a rapid change centered on 13.7 and 12.3 ka. (Ruddiman 1987).

The Younger Dryas deglaciation model is suggested by sea-surface temperature cooling between 11 and 10 ka in the North Atlantic Ocean. At least early and perhaps all of Brady pedogenesis coincides with an abrupt and brief cool interval correlative with the classic Younger Dryas cold interval of the North Atlantic region.

Younger Dryas

The Younger Dryas, as the last glacial cold spell, was an abrupt and well defined event (Dansgaard et al.1989, Broecker et al. 1988), which has been absolutely dated at about 11 ka to 10 ka. (Table 4). In this short period of time, the return to near-glacial conditions interrupted the Pleistocene/ Holocene climatic transition, during which most of the Northern Hemisphere ice sheets melted. A leading explanation for the Younger Dryas cooling depends primarily on a mechanism for cooling of North Atlantic waters, rather than on the radiation distribution or directly on the presence of the ice sheets (Wright 1989). During deglaciation, large quantities of meltwater flowed from the melting Laurentide Ice Sheet. Appreciable evidence exists to suggest that the Younger Dryas coincided with changes in the routing of meltwater between the Mississippi and St. Lawrence drainage basins (Broecker et al. 1988, Broecker et al. 1989, Lehman and Keigwin 1992, Taylor et al. 1993). The influx of fresh water to high latitudes of the North Atlantic has been suggested as inhibiting the generation of dense, saline North Atlantic deep water, which, in turn, led to a reduction in heat transport to the North Atlantic (Broecker et al. 1988). For this reason, the Younger Dryas was recorded much more distinctly in Europe and Greenland than in North America, and is thought to be confined to amphi-Atlantic region. General circulation model results also support that conclusion, i.e., they show that temperatures for winter were deeply depressed far across Eurasia but were little changed in North America (Rind et al. 1986).

Many recent studies, however, demonstrate that the varied climatic deterioration was felt well beyond the North Atlantic (e.g., An et al. 1993, Mathewes et al.1993, Kudrass et al.1991, Engstrom et al.1990, Wright 1989). In their study of pollen and chemical stratigraphy in southeastern Alaska, Engstrom et al. (1990) suggested that a significant climatic reversal occurred in this region between about 10,800 and 9,800 yr B.P. The temporary return of tundra after full development of lodgepole pine parkland is regarded as a clear response to climatic reversal, even though it is not contemporaneous with any known readvance of glaciers in the area or elsewhere in the Pacific Northwest. More recently, Mathewes et al. (1993), in their study of the British Columbia coast, also reported a shift from forest to open, herb-rich vegetation after 11,000 yr B.P., in response to colder and wetter conditions identified by a pollen-climate function. Shane and Anderson (1993) argued that the recurrence of spruce between about 11,000 and 10,400 yr B.P. supports the interpretation of regional temperature decrease in the Till Plains region of Ohio, Indiana, Michigan and Illinois.

Johnson and Park (1996) suggested that the timing of a magnetic susceptibility reversal within the Brady soil forming interval at the Bignell Hill type section matches with the Younger Dryas cooling record from the oxygen isotope data of a Greenland ice core. This minor, but notable drop in the susceptibility intensity occurs immediately below or within the lowermost Brady soil, which may indicate climatic degradation comparable to the Younger Dryas cold spell (Figure 17). Expanded use of AMS dating may provide the chronological framework needed to develop estimates of time and rates of changes during Younger Dryas time within the region.

HOLOCENE

Geomorphology and Stratigraphy

Bignell Loess

The Bignell loess was first described and named at the same type locality as the Brady soil (Schultz and Stout 1945). It is typically a gray or yellow-tan massive, calcareous silt, seldom more than 1.5 m thick. Although Bignell loess is often less compact and friable than the underlying Peoria loess, no certain identification can be made without the presence of the Brady soil (Caspall 1970). The Bignell loess does not form a continuous mantle, but is most prevalent and thickest adjacent to river valleys, particularly the south side, and often occurs in depressions on the Peoria surface. Of the loesses comprising the late-Quaternary stratigraphy of the central Great Plains, the Bignell is the only one that appears to have been deposited during a warm, nonglacial climate.

Eolian Sand Deposits

Holocene history documented for the sand sheets of Nebraska, Colorado and Kansas has indicated significant activity. Global climate change, resulting in shifting temperature and precipitation patterns, has been the focus of many of the studies focusing on the sand sheets of the central Great Plains (e.g., Forman et al. 1992b, Yuhas 1993, Arbogast 1995). For example, numerous years of drought during the spring growing season is an effective mechanism for reducing vegetative cover and resultant dune destabilization. In fact, during historic drought, the coverage of native short-grass vegetation was reduced, and soils were extensively eroded by eolian activity (Tomanek and Hulett 1970).

A record of late Holocene dune activity comes from the Nebraska Sand Hills through the research of Ahlbrandt and Fryberger (1980) and Ahlbrandt et al. (1983). The latter work, producing the first stratigraphically controlled radiocarbon ages from this large sand sea, reported that the most recent period of dune activity was not during the Wisconsinan glaciation, but rather during the late Holocene (ca. 3,000-1,500 yr B.P.). Their conclusions were based on data from seven sites: three with maximum-limiting radiocarbon ages of about 3,000 yr B.P. and four with maximum-limiting ages of about 10,000-5,000 yr B.P. They correlate their age estimates for stabilization of the dunes around 1,500 yr B.P., based on archaeologic and pollen evidence, with the interstade between the Triple Lakes and Audubon glacial advances in the Colorado front Range reported by Benedict (1973). Further evidence for a Holocene age of dunal development offered by Swinehart (1990) was a radiocarbon age of 13,160 yr B.P. obtained in alluvium 3-4.5m below a 52-85m-high barchan dune in the central Sand Hills. Vibracores from fens in Cherry County in the central Sand Hills collected by Ponte et al. (1994) indicated multiple peat layers. Radiocarbon ages outlined two major periods of eolian activity during the middle Holocene and two subsequent periods at about 3,500-2,800 yr B.P. and after 1,000 yr B.P.

The Great Plains region of northeastern Colorado is also an area of extensive sand dunes. Parabolic dunes in the region provide primary paleoclimatic information: dunes are elongate parallel to prevailing winds, causing the limbs of parabolas, anchored by vegetation, to point up wind. The dominant northwest-to-southeast orientation indicates that winds from the northwest shaped the landforms. Such strong prevailing winds on the High Plains are associated with air masses originating from the North Pacific or Canadian Arctic, and they preclude significant influence of tropical or subtropical air masses (Borchert 1971). These dunes exhibit evidence for a late-Holocene dry period (Muhs 1985), i.e., soils developed on these dunes have morphological and textural properties similar to soils on stabilized dunes in the Nebraska Sand Hills with maximum limiting radiocarbon ages of about 3,000 yr B.P. Forman and Maat (1990), using thermoluminescence and radiocarbon dating, obtained ages of 7-9ka on soils buried in dunes near Hudson, Colorado. Forman et al. (1992b) documented a succession of paleosols buried by eolian sand during the Holocene, indicating that there were four possible periods of eolian sand deposition in the Holocene-about 9500 to 5500 yr B.P., 5500 to >4800 yr B.P., 4800 to >1000 yr B.P., and <1000 yr B.P., separated by relatively short intervals. Using radiocarbon dating, archaeological data and other information, Madole (1994) observed that the sand sheet of northeastern Colorado were mobilized within the last 1,000 years. Stratigraphic evidence from Nebraska and northeastern Colorado indicates extensive sand sheet reactivation and dune formation during the late Holocene, with significant mobilization during the last 1000 years in Colorado. Global climate change, resulting in different temperature and precipitation patterns, has been the focus of these studies of sand sheet activity in the central Great Plains (e.g., Forman et al.1992b, Yuhas 1993, Arbogast 1995).

Johnson (1991a) and Arbogast (1995) documented periods of dune activity in the Great Bend Sand Prairie of Kansas. The most intensive period of dune formation in the region apparently occurred between 9 and 6 ka, an interval of sand mobility widely recognized on the Great Plains. In the late Holocene, loess accumulated episodically on relatively flat landscapes, while sand sheets and dunes were mobilized from about 5,700 - 4,800, 2,300 - 1,700, 1,600 - 800, and 200 yr B.P. The orientation of parabolic and barchan dunes indicates that prevailing winds during the Holocene have been generally southwesterly.

Recent geomorphic research in the Great Plains (Holliday 1987; Forman and Maat 1990; Swinehart 1990) has indicated that the middle Holocene, or Altithermal (Antevs 1955), was an episode of decreased precipitation and increased erosion. A number of studies suggest an Altithermal age for dune sand on the Great Plains. On the Southern Plains, Holliday (1985, 1989) identified two periods of dune sand movement at about 6,500-5,500 and 5,000-4,500 yr B.P. These latter episodes of dune sand movement have been correlated with similar-aged dune deposits in Bailey County, Texas (Gile 1979). Thus, geomorphic evidence strongly supports Benedict and Olson's (1978) reconstruction of Great Plains paleoclimate based on archaeological data.

It has been proposed that the Altithermal was the most likely time during which the large dunes of the Sand Hills formed (Swinehart 1990). Following about 2,000 years of stabilization, the climate became dry enough to allow reactivation of much of the sand in the eastern part of the Sand Hills.

Alluvial Deposits

During the last decade, a great deal of attention has focused on the development of alluvial chronologies in the central Great Plains, typically in connection with geoarchaeological investigations. As a consequence, this research has resulted in a number of studies and a sizable radiocarbon data base; well over 400 radiocarbon ages have been obtained from alluvium in Kansas and Nebraska (Johnson et al., 1996). Only a sampling of the many studies is presented below.

Much of the research in Nebraska has focused on the Loup River basin. Brice (1964) recognized two major terrace systems in the basin and obtained early Holocene radiocarbon ages of 10,500, 9,000, and 8,500 yr B.P. on fill beneath the lower of these terraces, the Elba. In a recent re-examination of the Elba terrace, May (1990, 1991) secured radiocarbon ages ranging from nearly 11,000 to 4,670 yr B.P. from the Cooper's Canyon area (Figure 16). On the South Loup River, May (1986, 1989, 1992) recognized four alluvial fills, with the oldest one dating between about 10,200 and 4,700 yr B.P., thereby correlating temporally with the Elba terrace of the North Loup. Elsewhere in the basin, Ahlbrandt et al. (1983) dated organic accumulations in alluvial sands at 8,410 yr B.P. from a site on the Dismal River.

In the Kansas River basin, alluvial geomorphic studies have a relatively long history, beginning in the 1950s. The first dating of alluvial stratigraphy on the Kansas River proper was done by Holien (1982), who obtained an age of 10,450 yr B.P. on a soil buried within lower Newman terrace fill at the Bonner Spring site. Subsequent radiocarbon dating of Newman fill at this locality (Johnson and Martin 1987, Johnson and Logan, 1990) and others (e.g., Bowman 1985) produced more early Holocene ages. The lower Holliday terrace has dated about 4,300 yr B.P. and younger (Johnson and Logan 1990).

Many studies have been conducted elsewhere in the Kansas River basin on the many tributaries. Some of the first radiocabon dating was carried out on samples collected from the Republican River basin by Schultz and his colleagues at the University of Nebraska: from varied locations they secured early to middle Holocene ages from buried soils. The most recent research in the basin was conducted by Martin (1990, 1992), who concluded from dating various alluvial fills that the majority of the fill was deposited less than about 4,600 yr B.P.

Several geoarchaeological studies were done in conjunction with cultural resource management projects focusing on federal impoundments within the Kansas River basin. Mandel (1987), in a study of the lower Wakarusa River, recognized two terraces, the lower of which produced radiocarbon ages of about 2,900 yr B.P. and less. A study of the alluvial history of the Smoky Hill River in the vicinity of Kanopolis Lake (Mandel 1988, 1992) revealed a striking absence of early and middle Holocene fill in small valleys, and middle Holocene fill in the main valleys and in alluvial fans.

In their study of Wolf Creek basin, Kansas, Arbogast and Johnson (1994; Figure 18) observed that alluviation of early-Holocene flood plains in this small basin was episodic, with at least one period of flood plain stability and soil formation about 6,800 yr B.P. During the middle Holocene (ca. 6,500-5,300 yr B.P.), lateral erosion and entrenchment flushed most early-Holocene fill from the main valley of Wolf Creek and the lower reaches of its larger tributaries. Following the interval of mid-Holocene erosion, sediment accumulated on flood plains between 5,300 and 3,000 yr B.P. Late Holocene alluviation was episodic, with intervening periods of flood plain stability and soil formation about 1,800, 1,500, and 1,200 yr B.P.

A number of studies have been conducted in the Arkansas River basin area of south-central and southeastern Kansas. Mandel examined terraces and associated fills in the Neosho (Mandel 1992, 1993) and Verdigris Rivers (Mandel 1993), obtaining radiocarbon ages on fill to about 4,200 yr B.P. The most extensive study in the Arkansas River basin was that of the Pawnee River basin by Mandel (1988, 1991, 1994). Two terraces were recognized in the higher order tributaries, with fill of the high terrace dating between about 10,000 and 5,000 yr B.P., and that of the low terrace to 3,000 yr B.P. and younger. Of the three terraces present in the lower part of the system, the lowermost one has Holocene fill and the others are Pleistocene. Holocene valley fills in the Pawnee Basin appear to lack soil development from about 7,000 to 5,000 yr B.P.

The alluvial record is temporally and spatially fragmented, i.e., the history of valley and stream evolution stored in alluvium is scattered and wrought with gaps. So, it is only by assembling this fragmentary information that one obtains a unified perspective on the record of stream evolution. Out of the many studies conducted in recent years, a pattern of change is emerging. Large stream valleys appear to contain, more or less, alluvial fill dating throughout the Holocene, whereas small stream valleys typically contain only fill dating in the late Holocene. This model has an intuitive basis in that the probability of survival of early and middle Holocene fill in smaller streams is greatly diminished by the limited storage capacity for alluvium and the relatively high stream gradients, large area in hillslope, and associated peaked flood waves. Exceptions to this pattern do, of course, exist (e.g., Lime Creek, Nebraska: May 1996; Wolf Creek: Arbogast and Johnson 1994), but are likely due to locally unusual valley width and other discernable factors.

A first approximation of this alluvial model was presented by Johnson and Martin (1987) in an examination of radiocarbon ages obtained from alluvial fill in the central Great Plains. In recent years, the model has evolved with a vastly expanded data base and been articulated recently by Mandel (1995). He noted that fill in small valleys appears to be less than 4,000 years old, and that the missing early and middle-Holocene record is frequently preserved at the lower end of small stream valleys as terrace fill or alluvial fans.

From the alluvial chronologies, it is obvious that regional synchroniety of stream behavior exists in the central Great Plains (Johnson and Martin 1987, Johnson and Logan 1990, Mandel 1995). When erosion and sedimentation are considered in a stream hierarchical sense, patterns of coincidence appear, such as similar times of flood-plain stability and attendant soil formation. A frequency distribution of over 400 radiocarbon ages from alluvium of Kansas and Nebraska (Figure 19) provides an indication of the synchroniety. The high frequencies of the last 5,000 years reflect the age of the alluvium in large and small streams, whereas those prior to about 8,000 represent the ages from the large valleys alone. Alluvial fans ages account for many ages within the 4,000 to 8,000 year range (Mandel 1995). The greatest frequency of ages occurs about 1,200 yr B.P., a time when pronounced low terrace stability and soil development occurred throughout the stream systems. Another notable feature of the distribution is that when the ages obtained from alluvial fans are not considered, very few alluvial ages fall within the 5,000 to 7000 yr B.P. period. This paucity of ages suggests little flood-plain stability and/or preservation of alluvium from that interval, which coincides with the Altithermal climatic episode. Stream activity of this dry period may have been characterized by rapid sedimentation, thereby precluding soil development, in response to low-frequency, high-intensity convectional storms (Knox 1976, 1983).

Regional synchrony in Holocene fluvial behavior suggests that climatic fluctuation is the dominant external variable in stream systems (Wendland 1982; Knox 1976, 1983). Changes in climate during the Holocene were frequent and episodic (e.g., Wendland and Bryson 1974; Kutzbach 1985; COHMAP members 1988), resulting in discrete periods of stream stability and instability (Knox 1983). The concept of a middle Holocene, or Altithermal (ca. 7,000-5,000 yr B.P.) cultural hiatus on the Great Plains has become well-entrenched within the archaeological literatures. Of the various theories put forth to explain the hiatus (Reeves 1973), fluvial erosion or aggradation sufficient to dramatically alter the record for the region during the interval 7,000-5,000 yr B.P. is most pertinent (Johnson 1987, Mandel 1995). Some argued that the similarity in the alluvial stratigraphic record from eastern humid portions of the region to the more arid western areas, as well as with chronologies further afield indicates that regionally anomalous erosion and deposition do not explain the hiatus completely; rather, the increased dryness during the Altithermal was likely sufficient to reduce populations on the Plains (Wedel 1961; Knox 1978; Wendland 1978). However, the rapidly expanding alluvial radiocarbon and stratigraphic data base for the region is indicating that much of the cultural record, namely that of the Archaic period, is buried, often deeply, or lost to erosion.

Climatic Proxies

Fossil Pollen

Palynological documentation of vegetation and climatic change within the Holocene presents some special challenges (Fredlund and Jaumann 1987). These problems are, at least in part, the result of the taxonomic limitation of pollen analysis. Many major grassland pollen types encompass entire families of plants (Fredlund 1991), and, consequently, large changes within grasslands can occur but not be readily apparent within the pollen record (Wright et al. 1985). This taxonomic limitation explains the lack of clear palynological definition of the middle-Holocene climatic drying in the central Great Plains. Because of the limited records and inability to differentiate grass pollen, little Holocene vegetational change is apparent in the fossil pollen record (Baker and Waln 1985).

Abundant palynological evidence exists for middle-Holocene eastward migration of the prairie/forest ecotone. Several palynological studies from areas peripheral to the central Great Plains document middle-Holocene expansion of the prairie (e.g., Brush 1967, Watts and Bright 1968, Durkee 1971, Van Zant 1979). Barnosky et al. (1987) subsequently documented the eastward ecotonal shift between about 8,000 and 6,000 years ago through a review of data from the northern Great Plains. Using pollen/climate transfer functions, Bartlein et al.(1984) estimated that precipitation in the Minnesota area was about 20% less during the middle Holocene than it is today, but that temperature was only slightly higher.

In Nebraska, a paleoecological record comes from Sears' (1961) study of Hackberry Lake in the north-central part of the Sand Hills. A radiocarbon age indicates that organic deposition began at this site about 5,040 yr B.P., and the sediments also record a fluctuating dominance of prairie vegetation that persists to the present, but with no discernible record of the Altithermal. Since the sand dunes that enclose the Hackberry Lake basin are well-preserved barchan and barchanoid-ridge dunes that indicate prevailing wind directions to the southeast, this site appears to represent a post-Altithermal stabilization of the dunes. On the southwestern margin of the Sand Hills at Swan Lake, Wright et al. (1985) analyzed a core with a basal radiocarbon age of about 8,000 yr B.P. Sedimentation in Swan Lake appeared to be continuous to the present, and pollen analysis indicated a prairie vegetation with minor fluctuations of herbs and grasses throughout this time, but no Altithermal signal.

Two sites in Kansas provide palynological information for the Holocene: Muscotah Marsh (Grüger 1973) and Cheyenne Bottoms (Fredlund 1995). The Holocene portion of the record at Muscotah Marsh in north-central Kansas contains unconformities and lacks close-interval radiocarbon ages, but clearly portrays middle Holocene prairie expansion and contraction. At Cheyenne Bottoms in central Kansas, the Holocene is markedly different from the late-Pleistocene Farmdalian grassland-steppe assemblage: lower *Artemisia* percentages and lower relative frequencies of arboreal pollen types characterize the Holocene. These differences suggest that the Holocene regional upland vegetation in the Holocene lacked the sage component which was so important during the Farmdalian. The Holocene vegetation also lacked diversity of tree and shrub taxa regionally present during the Farmdalian. Of all tree and shrub pollen taxa identified, only *Ulmus* (elm) and *Celtis* (hackberry) are more common during the Holocene. Fredlund (1995) divided the Holocene into four microzones based on changes in the local pollen signal. The latest Pleistocene-earliest Holocene zone (>9,690 yr B.P.), through its abundance of diatoms and gastropods, suggests increasing moisture at the site. The soil developed above this zone appears to correlate temporally with the Brady soil. The high relative frequencies of *Cheno-Am* type pollen throughout the Holocene are associated with the existence of mudflats periodically exposed as fluctuations of water levels occurred within the basin. In the middle Holocene (ca. 8,500 to 3,700 yr B.P.), frequencies of *Cheno-Am* pollen types decreased significantly, suggesting more stable, perhaps lower, water levels. The increase in *Ambrosia* (ragweed) pollen during the middle Holocene indicates less fluctuating and lower water levels. The late Holocene (> 3,700 yr B.P.) was characterized by a return to fluctuating water levels and exposed mudflats.

The timing of the Holocene dry/warm interval appears to vary geographically. In Minnesota the maximum of Altithermal warmth and dryness occurred between about 8,000 and 4,000 yr B.P., peaking at 7,200 yr B.P. (Wright 1976). In the northwestern United States most sites register greatest drought in the early Holocene, although at some sites it was delayed until the middle Holocene, concurrent with the Midwest (Barnosky et al. 1987). In the Southern High Plains, widespread eolian activity began in some areas by 9,000 yr B.P. and culminated 6,000-4,500 yr B.P., probably because of warmer, drier conditions that reduced vegetation cover (Holliday 1989).

Stable Carbon Isotopes

A gradual shift to drier and warmer conditions occurred during the late Pleistocene. Using stable oxygen and carbon isotopes from lacustrine and soil carbonates collected at Fort Hood in north-central Texas, Humphrey and Ferring (1994) demonstrated that mesic conditions continued until 7500 yr B.P., except for a brief drying period between about 12,000 and 11,000 yr B.P. The slow replacement of cool-season plants by warm-season plants at Fort Hood agrees with an extended warming and drying climatic transition during the early Holocene.

By the middle Holocene, drying had reached a maximum according to most studies. Northwestern Texas was experiencing conditions of maximum temperatures, minimum precipitation, and eolian activity between 6000 and 4500 yr B.P. (Holliday 1985, 1989; Pierce 1987). $\delta^{13}C$ values derived from paleosols in this region revealed a shift from -23‰ in the early Holocene to -15‰ in the middle Holocene (Haas et al. 1986), i.e., a shift in dominance from cool-season C_3 grasses to warm-season C_4 grasses. Based on enriched $\delta^{13}C$ values in soil carbonate from their Texas study, Humphrey and Ferring (1994) identified a middle-Holocene xeric episode, although the $\delta^{18}O$ values from these same carbonates did not indicate a significant temperature change.

Despite an aberrant value, limited $\delta^{13}C$ data from the Sargent site, an upland loess exposure in southwestern Nebraska, suggest a gradual increase in dryness through the Holocene (Figure 20); this is interpreted as a shift in the abundance of C_4 species from slightly under 50% during the late Pleistocene to 80 - 90% in the middle Holocene. $\delta^{13}C$ data derived from the correction of radiocarbon ages obtained from soils buried in alluvial fill of the central Great Plains (Johnson et al. 1996) also indicate a gradual increase in C_4 plants from about 12,000 yr B.P. through the Holocene, but these data are relatively noisy, however, due to the edaphic conditions encountered on bottomlands.

Rock Magnetism

An extended period of loess deposition is indicated by decreased weathering, i.e., decreased magnetic susceptibility. Magnetic susceptibility intensities measured at the Eustis ash pit-Bignell Hill section, Nebraska, and the Beisel-Steinle and Barton County landfill sites in Kansas (Figure 12) indicate high values in the Bignell loess (averaging 70-100 x $10^{-8}m^3/kg$), when compared to the Peoria loess, and are only slightly lower than those of the Brady soil. Because the magnetic signal of the Bignell loess is only slightly diminished from that of the Brady soil and far greater than that of the Peoria loess (Figure 12), the former likely has as part of its source pre-weathered sediment from the adjacent, exposed Brady soil surface. Weaker intensities from mid-Bignell loess indicate, however, that either the depositional rate was faster than the rest of Holocene or pedogenic and/or biologic activity was weaker in Altithermal time, presumably the latter.

Magnetic data from the Sumner Hill and Manhattan Airport sites in east-central Kansas also display reduced magnetic intensities in the Holocene Bignell loess. Frequency dependence data from Sumner Hill suggest that pronounced surface instability (rapid loess input or throughput) existed at this bluff-top position during much of the Holocene (Figure 14). The Manhattan Airport site, consisting of Pleistocene and Holocene alluvial fill, displays a steady upward decline that likely reflects an increase in alluviation (Figure 15). The Holocene fill may have originated from an unnamed tributary entering from the west, rather than from the Kansas River proper; contributions of Bignell loess may also be present.

The DB site, located on the loess-mantled bluff overlooking the Missouri River valley in northeastern Kansas, yielded cultural material dating from the early Holocene (Johnson 1996a). A buried soil, believed to be the Brady, is truncated down to its Bt horizon, and the surface soil, developed in overlying loess, is welded to the soil below. The frequency dependence curve derived from the site increases from the unaltered Peoria loess below at about 1.3m and exhibits two bulges in the upper meter, the lowermost being the buried soil and the upper being the surface soil (Figure 21). Cultural material is associated with the buried Bt horizon as well as with the surface soil.

Tree Rings

Variations in tree-ring widths from one year to the next have long been recognized as an important source of chronological and climatic information. The mean width of a ring in any one tree is a function of many variables, including the tree species, tree age, availability of stored food within the tree and of important nutrients in the soil, and a whole complex of climatic factors, including sunshine, precipitation, temperature, wind speed, humidity, and their distribution through the year (Bradley 1985). The tree is essentially a filter or transducer which, through various physiological processes, converts a given climatic input signal into certain ring-width output which is stored and can be studied in detail, even thousands of years later (e.g., Yapp and Epstein 1977, Fritts 1983).

Unfortunately, the tree-ring record extracted from the central Great Plains covers only the last few hundred years, but does provide us with an impression of the recent variability in climate. Information on latest Holocene drought episodes comes from the ring sequences in logs buried at the Ash Hollow site in western Nebraska (Weakley 1962). According to that record, droughts longer than 15 years occurred in 1276-1313, 1438-1455, 1512-1529, 1539-1564, 1587-1605 and 1688-1707 A.D. In the North Platte area of western Nebraska, Weakley (1943), in a study of red cedar and ponderosa pine, found 13 more or less severe droughts lasting 5 years or more during the past 400 years. Drought appeared to recur at ill-defined intervals of from 15 to 25 years.

Climatic Modeling

Using a modified version of the Blytt-Sernander scheme of climatic episodes, Bryson et al. (1970) produced a model that subdivided the Holocene into the pre-boreal, Boreal, Atlantic, sub-Boreal, sub-Atlantic, Scandic, neo-Atlantic, and Pacific episodes. For example, during the Atlantic episode (8450-4680 yr B.P.), the wedge of modified Pacific air that characterizes the grassland climate was expanded northeastward into central Minnesota and eastward towards the Atlantic seaboard (Bryson et al. 1970).

According to recent model simulations, by around 9 ka summer insolation had increased but was still secondary in influence to the shrinking Laurentide ice sheet (COHMAP members 1988). The glacial anticyclone persisted in eastern North America, but was much smaller than at 12 ka. With the Pacific subtropical high gaining strength adjacent to the west coast of North America, northwesterly winds replaced westerly winds along the coast in the Northwest. The Midcontinent was still cooler and more moist than at present in July. By the early Holocene (9 ka), the ice had wasted appreciably, the jet stream was no longer split, orbital parameters were favoring increased temperatures, and zonal flow was dominating (Kutzbach 1987).

For the Altithermal, i.e., 6 ka, model results produced mean summer temperatures 2° to 4°C higher than present (COHMAP members 1988) and annual precipitation up to 25 % less than at present in the region (Kutzbach 1987). Surface westerly winds in the midcontinent were stronger than today, with warmer and drier conditions prevailing. Since 6 ka, simulation indicates that westerly flow has weakened and summer temperatures have decreased.

NATURE OF THE RECORD

A substantial body of knowledge exists concerning the paleoenvironmental history of the central Great Plains. This body of knowledge is, however, somewhat awkward tosynthesize because of its uneven distribution both geographically and chronologically, and its derivation from many different types of proxy data using many different types of methods. Nonetheless, when making a comparison with the level of knowledge two to three decades ago, we certainly have a much better impression of late-Quaternary environments today.

To extract the paleoclimatic signal from proxy data, the record must first be calibrated, which involves using modern climatic records and proxy materials to understand how, and to what extent, proxy materials are climate-dependent (Bradley 1985). It is assumed that the modern relationships observed have operated, unchanged, throughout the period of interest. All paleoclimatic research, therefore, must build on studies of climate dependency in modern-day natural systems, but not all environmental conditions in the past are represented in the modern times. Obviously, situations existed during glacial and early postglacial times which defy characterization by modern analogs (Martin and Martin 1987). Accordingly, one must be aware of the possibility that erroneous paleoclimatic reconstructions may result from the use of modern climate-proxy data relationships when past conditions have no analog in modern world.

Although there is considerable paleoenvironmental data from the central Great Plains, very little can be interpreted quantitatively, e.g., in terms of temperature or precipitation. Quantitative paleoclimatology seems to be conceptually quite difficult, with the few limited attempts thus far based on tree rings (e.g., Weakley 1962), pollen (e.g., Webb et al. 1993), phytolith morphology (Johnson et al. 1993a), and stable carbon isotopes (e.g., Nordt et al. 1994; Fredlund 1993; Johnson et al. 1993a). However, the application of mathematical techniques such as transfer functions to proxy data sets of the region holds the promise of allowing quantitative reconstruction of climate by relating quantitative data from modern environments to that of past environments. The use of such transfer functions requires systematic modern baseline data for both the proxies and modern climate.

A major problem in environmental reconstruction is that studies tend to produce paleoenvironmental data which lack the necessary chronological control. Such control for paleoenvironmental information must exist, or it is almost impossible to make spatial correlations; the data cannot, therefore, contribute to regional understanding. Until recent years this had been a serious problem for the central Great Plains, but chronostratigraphies are rapidly developing for the alluvial and eolian deposits. The late-Pleistocene vegetation cover of the region has provided only scattered wood and charcoal in loess, eolian sands, and alluvium; the grass cover of the Holocene has furnished even fewer datable materials. For radiocarbon control in this region, researchers have frequently used humates preserved in buried soils. Despite general acceptance of radiocarbon dating by Quaternary scientists, some debate still exists regarding the accuracy of humate-derived ages. Thermoluminescence dating has been employed in several studies (e.g., Oviatt et al. 1988, Feng et al. 1994, Maat and Johnson 1996). Together, radiocarbon and thermoluminescence dating are, however, producing consistent ages, resulting in an ever increasing resolution in the chronology for late-Quaternary deposits of the central Great Plains.

Table 1. Wood and charcoal ages from late-Wisconsinan deposits in Kansas and Nebraska.

Location	Material	Sample	Age	Source
Kansas				
Mt. Hope Sand Co. Sedgewick Co.	*Picea* litter	Dic-3101	19,340±200	Jaumann 1991
GMD5 site 9 Edwards Co.	*Picea* charcoal	TX-6479	17,970±330	Johnson 1991a
Coon Creek Graham Co.	*Picea* charcoal	GX-9355G	17,930±550	Wells and Stewart 1987b
Courtland Canal Jewell Co.	*Picea* charcoal	Beta-9320	14,450±140	Wells and Stewart 1987b
Nebraska				
North Cove Harlan Co.	*Picea glauca*	Beta-12286	14,700±100	Wells and Stewart 1987b
Coyote Canyon Harlan Co.	*Picea* charcoal	TX-7294	21,250±530	Martin 1993
Coyote Canyon Harlan Co.	*Picea* charcoal	TX-7295	19,730±300	Martin 1993
Sindt Point Harlan Co.	*Picea* charcoal	Tx-7711	21,440±200	Johnson unpublished data
Bloomington Franklin Co.	*Picea* or *Larix*	Beta-42015	18,830±180	May and Holen 1993
South Loup River Buffalo and Grant Co.	*Abies balsamea* fragments	Tx-6128 Beta-27758	14,080±190 13,160±450	May 1989, Swinehart 1990

Table 2. Radiocarbon ages from the Brady soil.

Site	Age (BP)	Lab. No.	Source
Nebraska			
Bignell Hill			
	8,080±180	n.a.	Lutenegger 1985
	9,160±250	W-234	Dreeszen 1970
	9,750±300	W-1676	Dreeszen 1970
	9,240±110	Tx-7425	Johnson 1993a
	10,670±130	Tx-7358	Johnson 1993a
North Cove			
west	10,550±160	Tx-6319	Johnson 1989
	10,220±140	Tx-6112	Johnson 1989
	10,270±160	Tx-6320	Johnson 1989
east	11,530±150	Tx-6321	Johnson 1989
	11,025±90	PITT-824	Martin 1993
Prairie Dog Bay			
	10,140±110	DIC-3310	Cornwell 1987
	10,360±130	Tx-5909	Martin 1993
	10,370±70	PITT-824	Martin 1993
	11,780±60	PITT-0961	Martin 1993
	9,020±95	PITT-825	Martin and Johnson 1995
Naponee			
	10,130±140	Beta-33939	Souders and Kuzila 1990
Kansas			
Speed	8,850±140	Tx-6626	Johnson 1990
	10,050±160	Tx-6627	Johnson 1990
Barton County			
	9,820±110	Tx-7045	Feng 1991
	10,550±150	Tx-7046	Feng, 1991

Ages represent the eolian phase only. An alluvial phase has been well documented throughout the Kansas River basin (Johnson and Martin 1987, Johnson and Logan 1990) and adjacent river systems, such as the Loup River of central Nebraska (Brice 1964, May 1990) and the Pawnee River (Mandel 1991, 1994) and Walnut River (Mandel 1990) of the Arkansas River system.

Table 3. $\delta^{13}C$ values and radiocarbon ages derived from the Brady soil A horizon (upper and lower 5cm).

Location	Age range	$\delta^{13}C$ (‰) top	bottom	Source
Nebraska				
Bignell Hill	9240-10,670	-17.40	-19.3	Johnson and May 1992
North Cove	10,360-10,550	-16.30	-18.9	Johnson 1989
Sargent Site	9920-10,620	-16.1	-20.1	Dort 1996
Elba Valley	9250-10,290	-15.4	-20.0	May 1991
Kansas				
Speed roadcut	8850-10,050	-18.8	-17.5	Johnson 1993a
Barton County	9820-10,550	-18.6	-19.0	Feng 1991

Table 4. Ages of the Younger Dryas.

Beginning	End	Area	Source
	10,720±150	Greenland Icecore	Dansgarrd et al. 1989
11,000	10,000	Sulu Sea, SE Asia	Kudrass et al. 1991
11,000	10,300	Ohio	Shane 1987
11,000	10,000	EN32-PC4, Orca Basin	Broecker et al. 1988
10,800	10,000	Alaska	Engstrom et al. 1990
11,290	10,170	British Columbia	Mathewes et al. 1993
11,000	10,000	Atlantic Canada	Mott et al. 1986
11,200	10,500	North Atlantic	Lehman and Keigwin 1992
	10580-10950	China	An et al. 1993
11,010±170	10,390±130	South Portugal	Bard et al. 1987

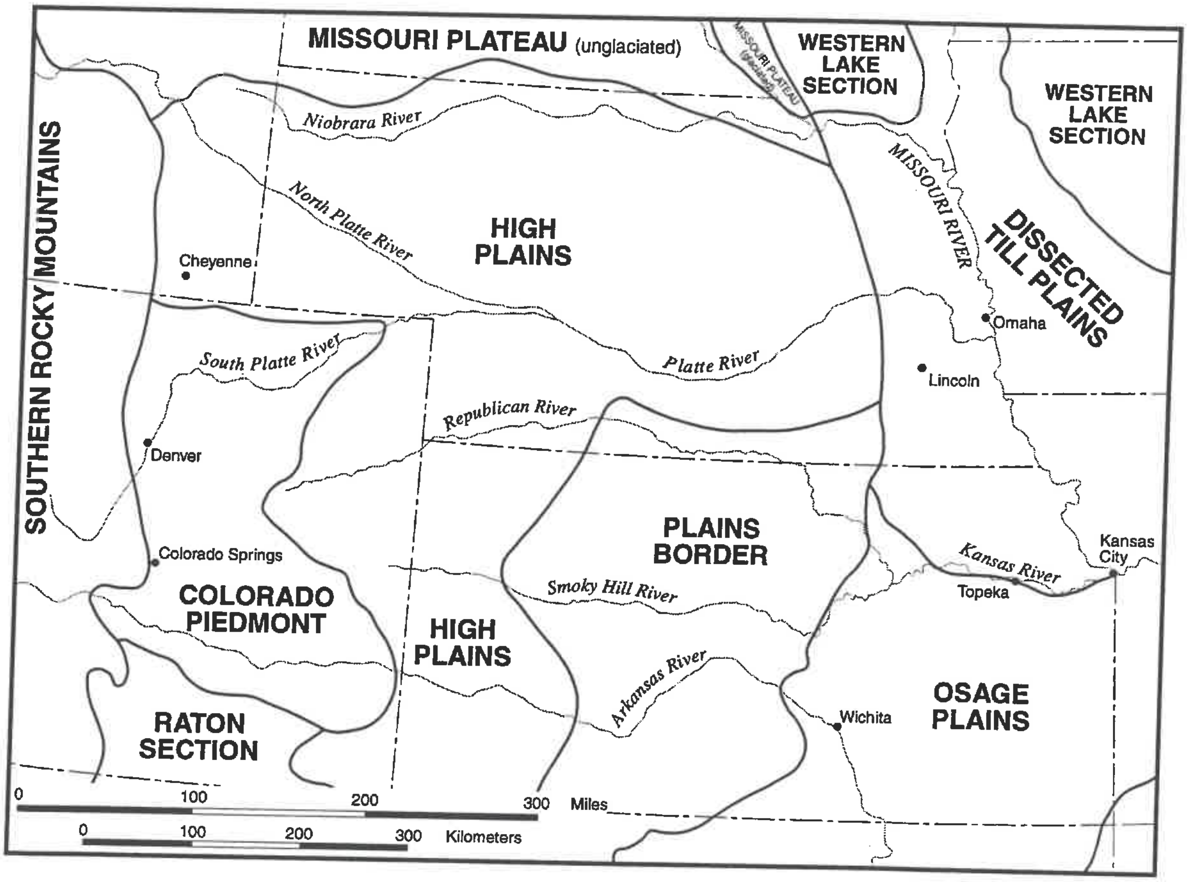

FIGURE 1. Physiographic map of the central Great Plains.

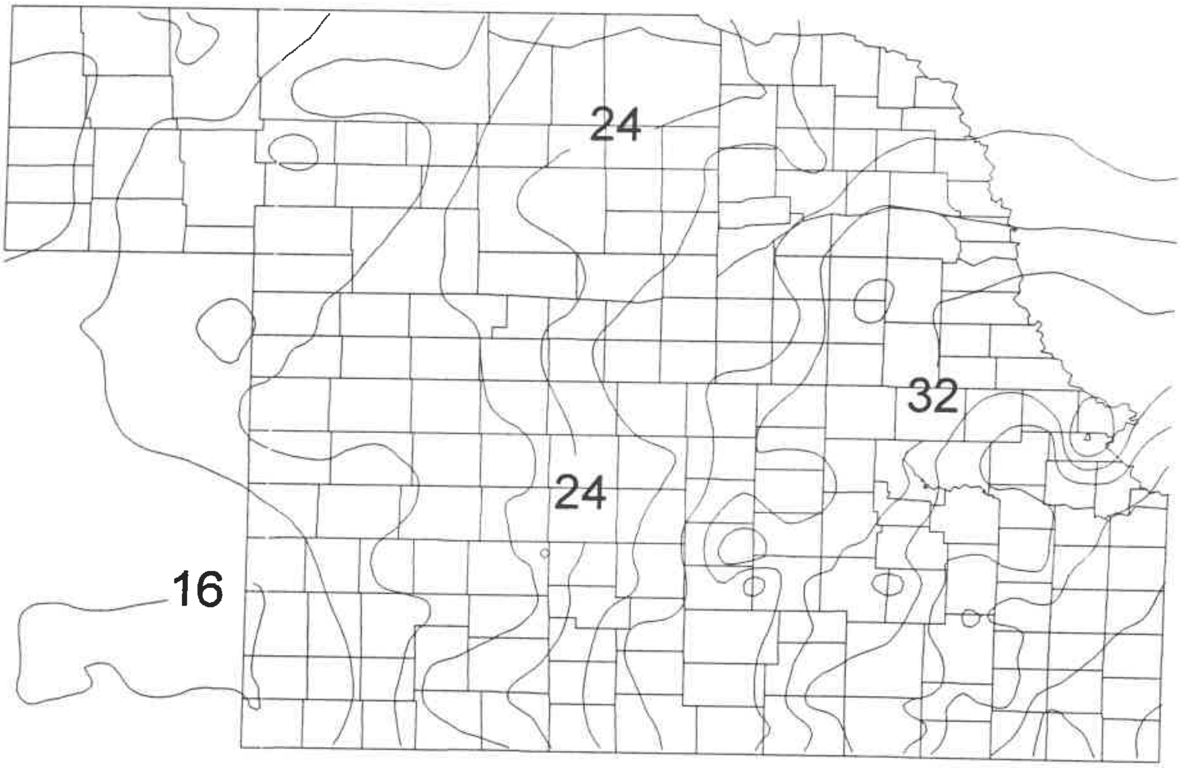

FIGURE 2. Mean annual total precipitation (inches) for the central Great Plains. (United States Department of Commerce 1983)

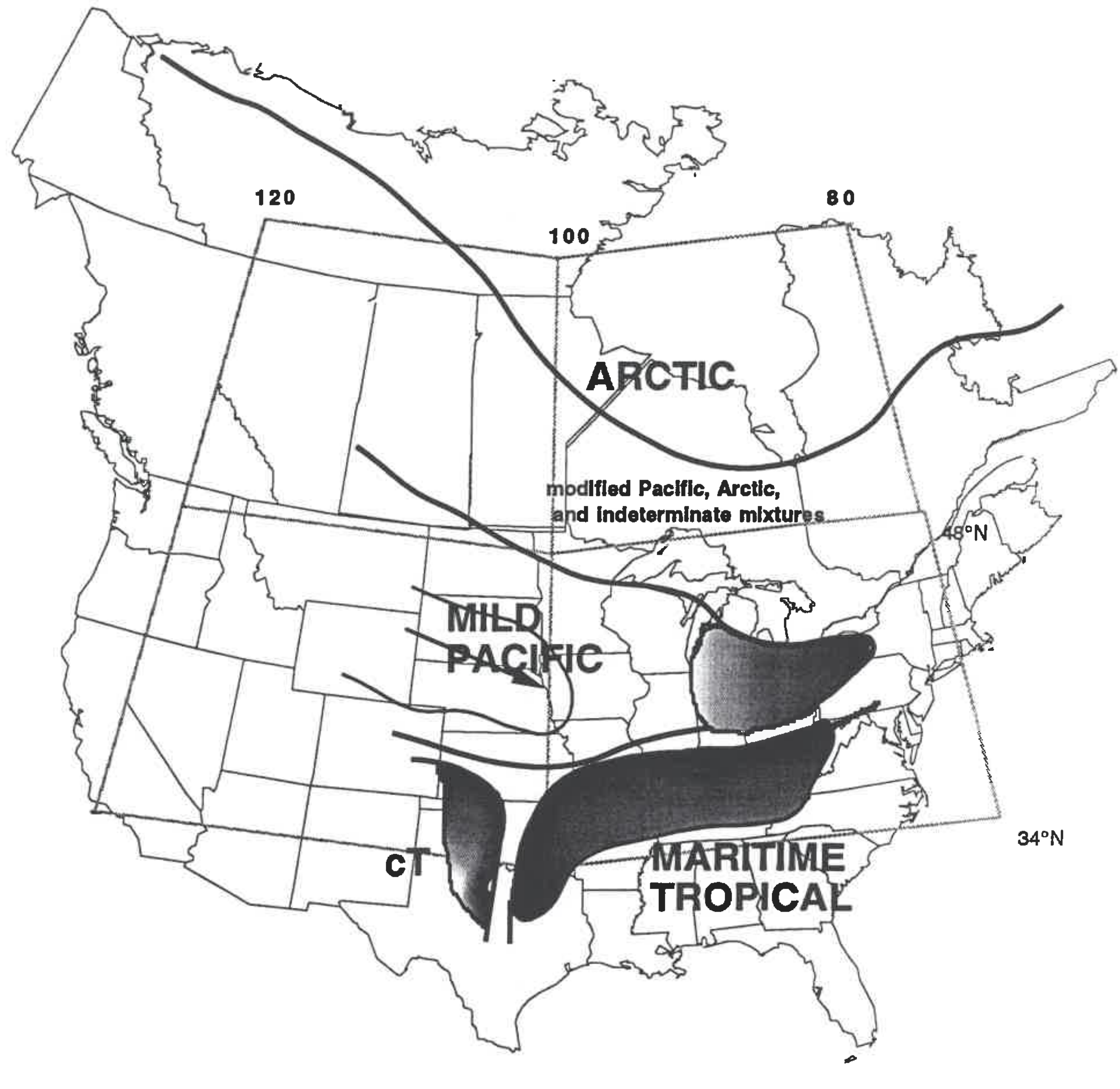

FIGURE 3. Regions dominated by the various air mass types. The shaded regions are occupied more than 50% of the time by the indicated air mass. (after Bryson 1966)

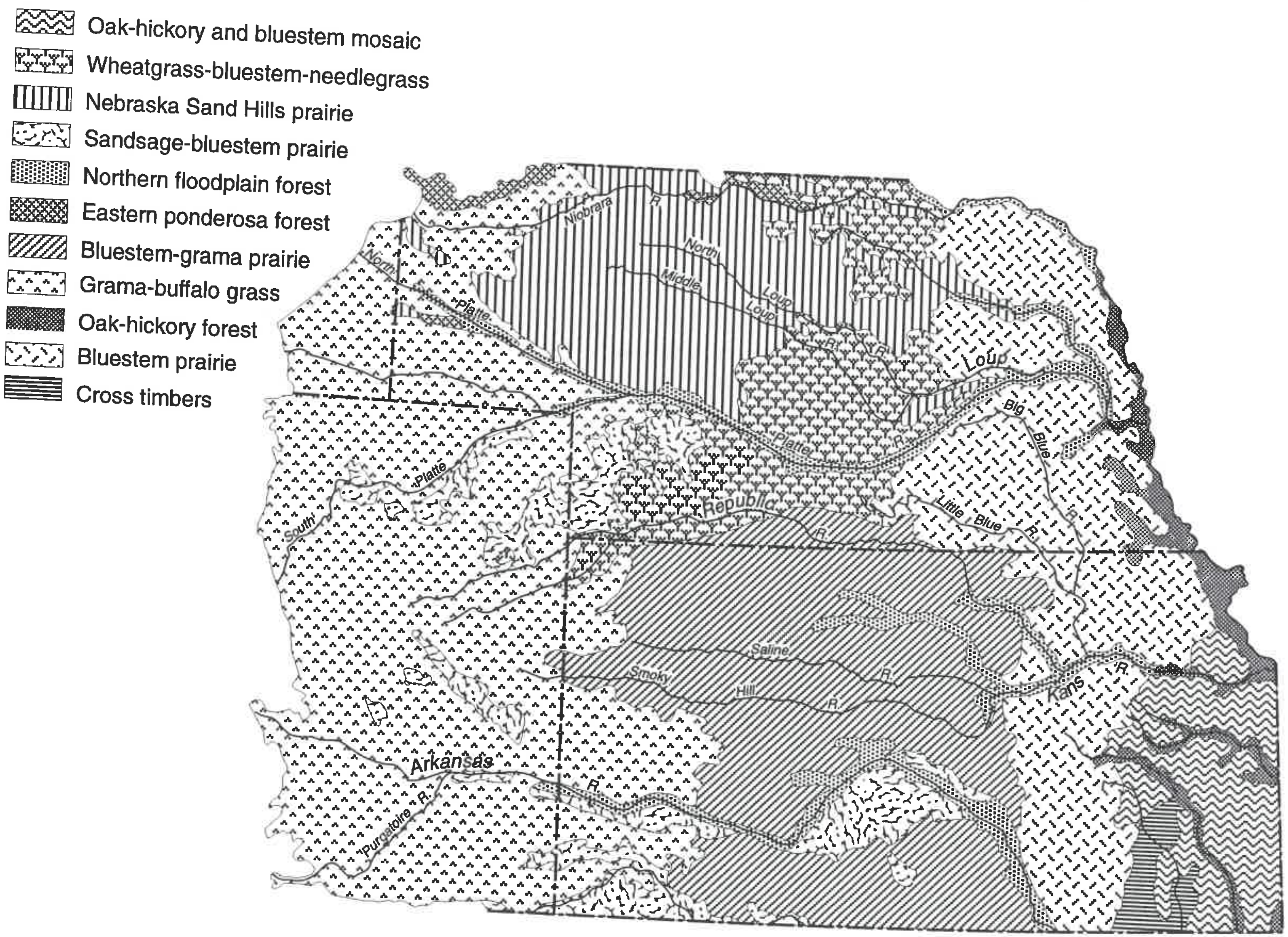

FIGURE 4. Native vegetation of the central Great Plains. (after Küchler 1964)

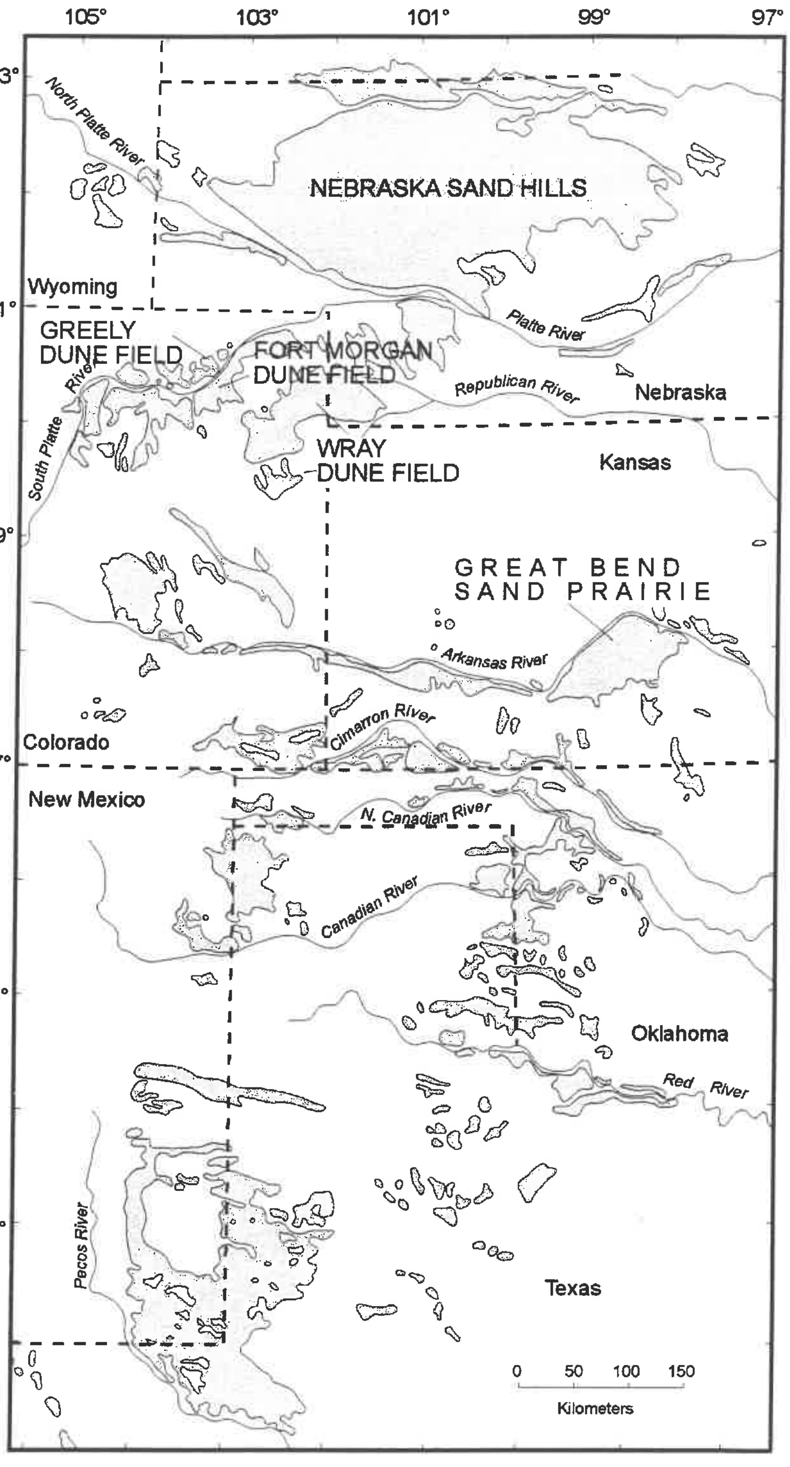

FIGURE 5. Distribution of dune fields and sand sheets in the central Great Plains. (after Muhs and Holliday 1995)

Nebraska

Time-Stratigraphic		Classification: Rock-stratigraphic: Eolian	Fluviatile	Glacial	Soils	Terrace Surfaces	
Wisconsinan	Late	Bignell Loess and Dunesand	Bignell Formation: Silt / sand-gravel	Absent		2a 2b	2
	Medial	Peoria Loess and Dunesand	Peoria Formation: Silt / Todd Valley Sand	Hartington Till	Brady	3	
	Early	Gilman Canyon Formation	Gilman Canyon Formation	Absent	Unnamed		3
Sangamonian	Late	Loveland Loess	Loveland Loess: Silt / sand-gravel	Absent	Sangamon	4	4

Kansas

Time-stratigraphic units	Rock-stratigraphic units: Northeastern area		Southeastern area		Central and Western area	
Recent stage	Eolian and fluvial deposits					
Wisconsinan Stage	Bignell Formation	Fluvial deposits	Bignell Formation	Fluvial deposits	Bignell Formation	Fluvial deposits
	Brady Soil					
	Peoria Formation G.C.F	Fluvial deposits	Peoria Formation	Fluvial deposits	Peoria Formation	Fluvial deposits
Sangamonian Stage	Sangamon Soil					
Illinoian stage	Loveland Formation	Fluvial deposits	Loveland Formation	Fluvial deposits	Loveland Formation Crete Formation	
	Yarmouth Soil					
Pre-Illinoian	Loess Cedar Bluffs Till Fluvial deposits Nickerson Till Atchinson Formation	Fluvial deposits	Fluvial deposits		Sappa Formation Grand Island Formation	
	Affon Soil					
	Loess Iowa point Till David City Formation	Fluvial deposits	Fluvial deposits		Fullerton Formation Holdredge Formation	

FIGURE 6. Late-Quaternary stratigraphic succession in Kansas (Bayne and O'Connor 1968) and Nebraska. (Reed and Dreeszen 1965)

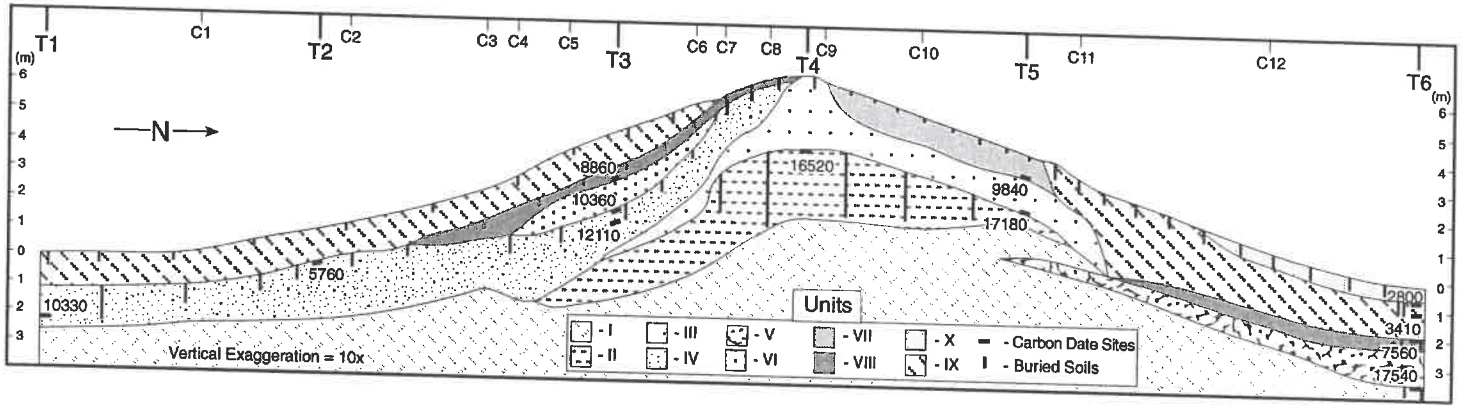

FIGURE 7. Cross-sectional stratigraphy of the Wilson Ridge lunette, Kansas. T and C designations along the top indicate the location of trenches and cores. (after Arbogast 1995)

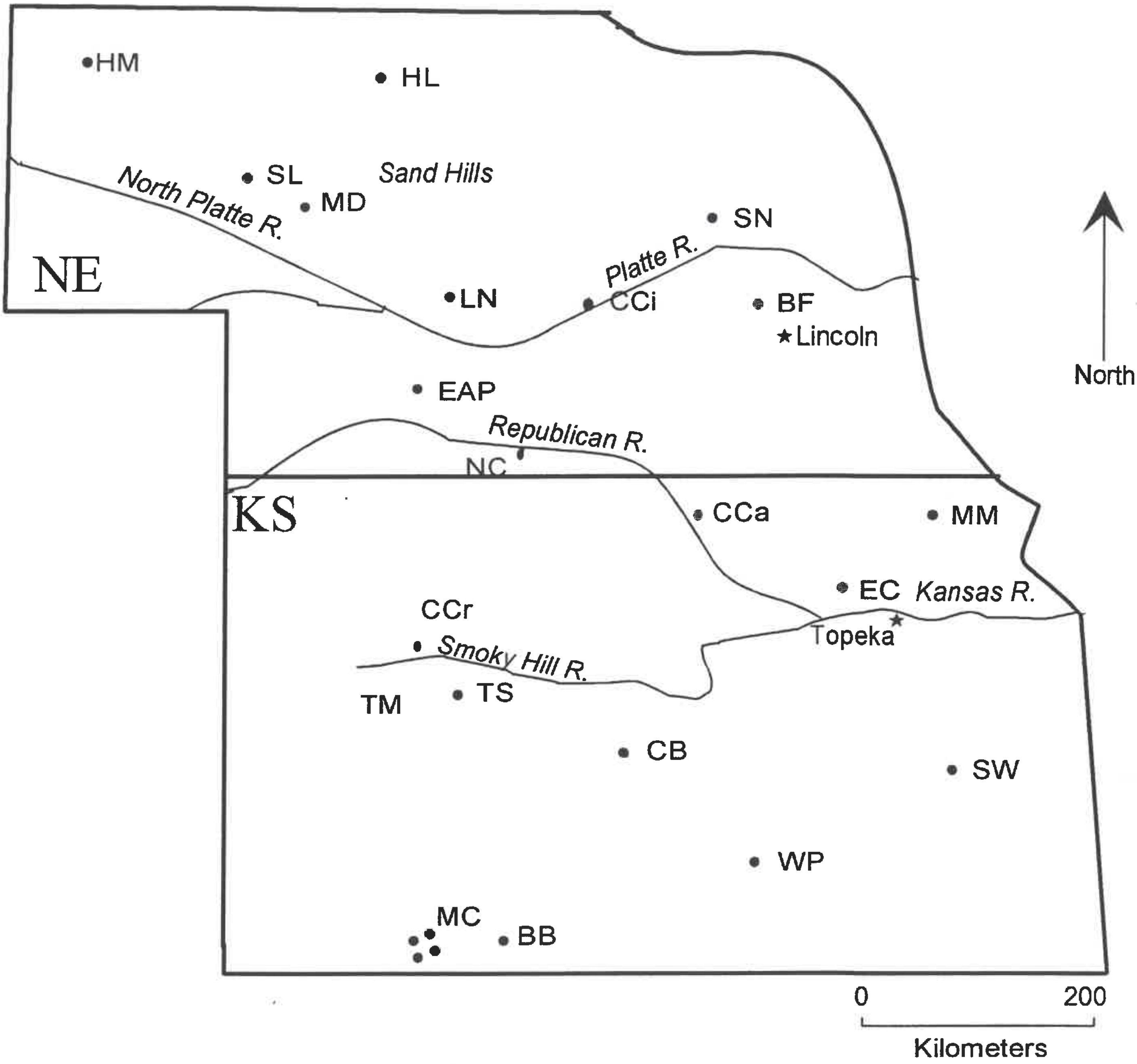

FIGURE 8. Late-Quaternary pollen and macro-botanical sites in the central Great Plains. (after Fredlund and Jaumann 1987) (Site key on page 60)

Key	Site, State	Depositional Environment	Time range	References
BB	Big Basin, KS	karst sink	Late Holocene	Schumard 1974
BF	Barteck farm, NE	prairie fen	Pre-Wisconsin	Kapp 1965, 1970
CB	Cheyenne Bottom, KS	marsh	Wisconsin	Fredlund 1991
CCa	Courtland Canal, KS	loess/colluvium	Woodfordian	Wells and Stewart 1987b
CCi	Central City, NB	loess/colluvium	Woodfordian	Martin unpub. data
CCr	Coon Creek, KS	colluvium/loess	Woodfordian	Wells and Stewart 1987b
EC	Elbow Creek, KS	alluvium/paleosol	late Holocene	Kurmann 1985
EAP	Eustis ash pit, NE	loess/paleosol	pre-Wisconsin to Holocene	Johnson 1993b
HL	Hackberry Lake, NE	interdunal lake	Holocene	Sears 1961
HM	Hudson-Meng, NB	colluvium	early Holocene	Agenbroad 1978
LN	Litchfield, NB	alluvium	Farmdalian(?)	Rogers 1985
MC	Meade County, KS	karst sinks, springs/alluvium	pre-Wisconsin	Hibbard 1970, Kapp 1965, 1970
MM	Muscotah Marsh, Arrington Marsh, KS	bogs/marshes and alluvium	Farmdalian Holocene	Grüger 1973
MD	McPherson County Drill Hole, NE	lake(?)	Farmdalian	Swinehart 1990
NC	North Cove, NE	spring deposits	Woodfordian	Johnson 1989
SL	Swan Lake, NE	interdunal lake	early Holocene	Wright et al. 1985
SN	Schuyler, NE	loess/colluvium	Woodfordian(?)	Fredlund and Jaumann, 1987
SW	Sanders', KS	spring-fed bog	Farmdalian	Fredlund and Johnson 1985, Fredlund and Jaumann 1987
TM	12 Mile Creek, KS	alluvium	early Holocene	Rogers and Martin 1984
TS	Trapshoot site, KS	loess/paleosol	Woodfordian (?)	Stewart and Rogers 1984
WP	Wichita peat, KS	bogs(?)/alluvium	Woodfordian	Jaumann 1991

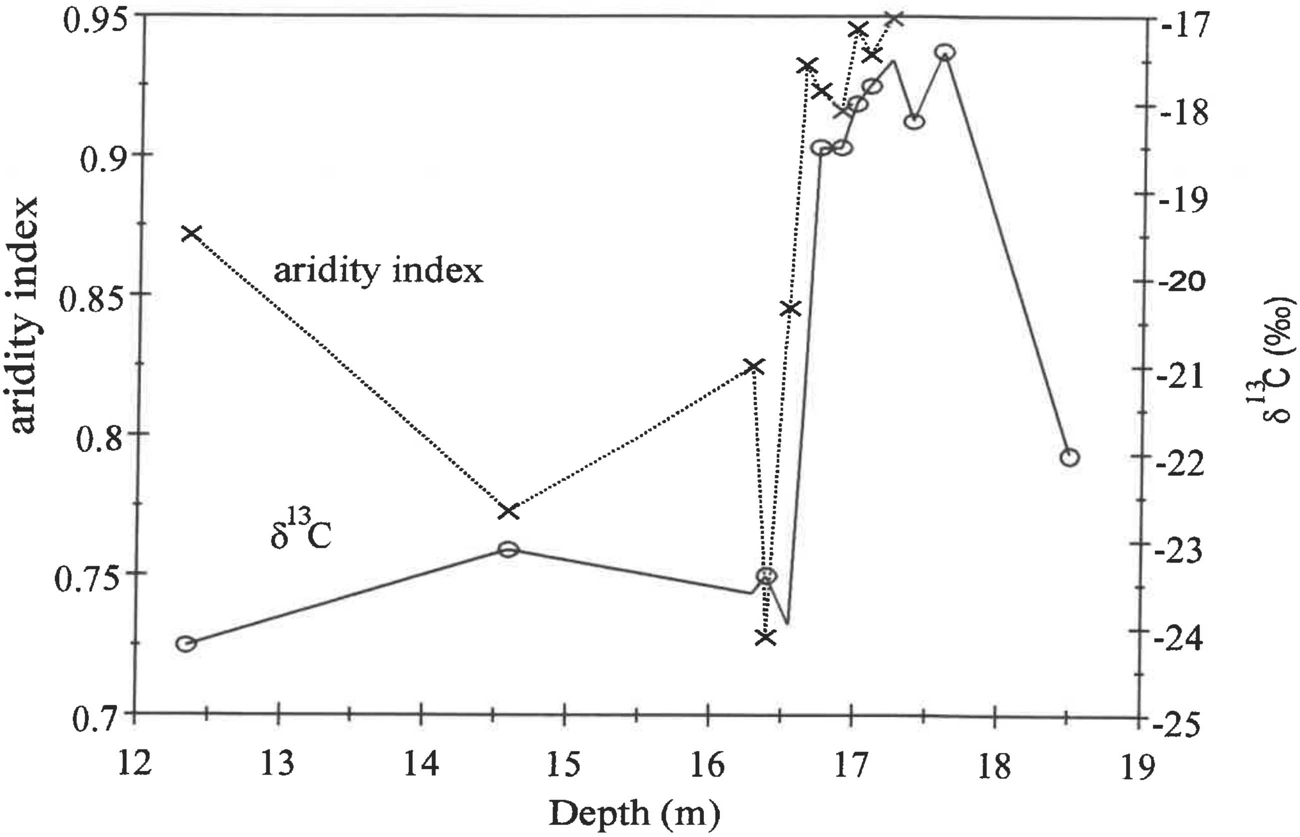

FIGURE 9. Aridity index and $\delta^{13}C$ curves from the Eustis ash pit, Nebraska. The abrupt change in both parameters at about 16.5m depth is indicative of the shift from the warm-season grasses of the Gilman Canyon Formation soil to the cool-season grasses of the overlying Peoria loess. (from Johnson 1993b)

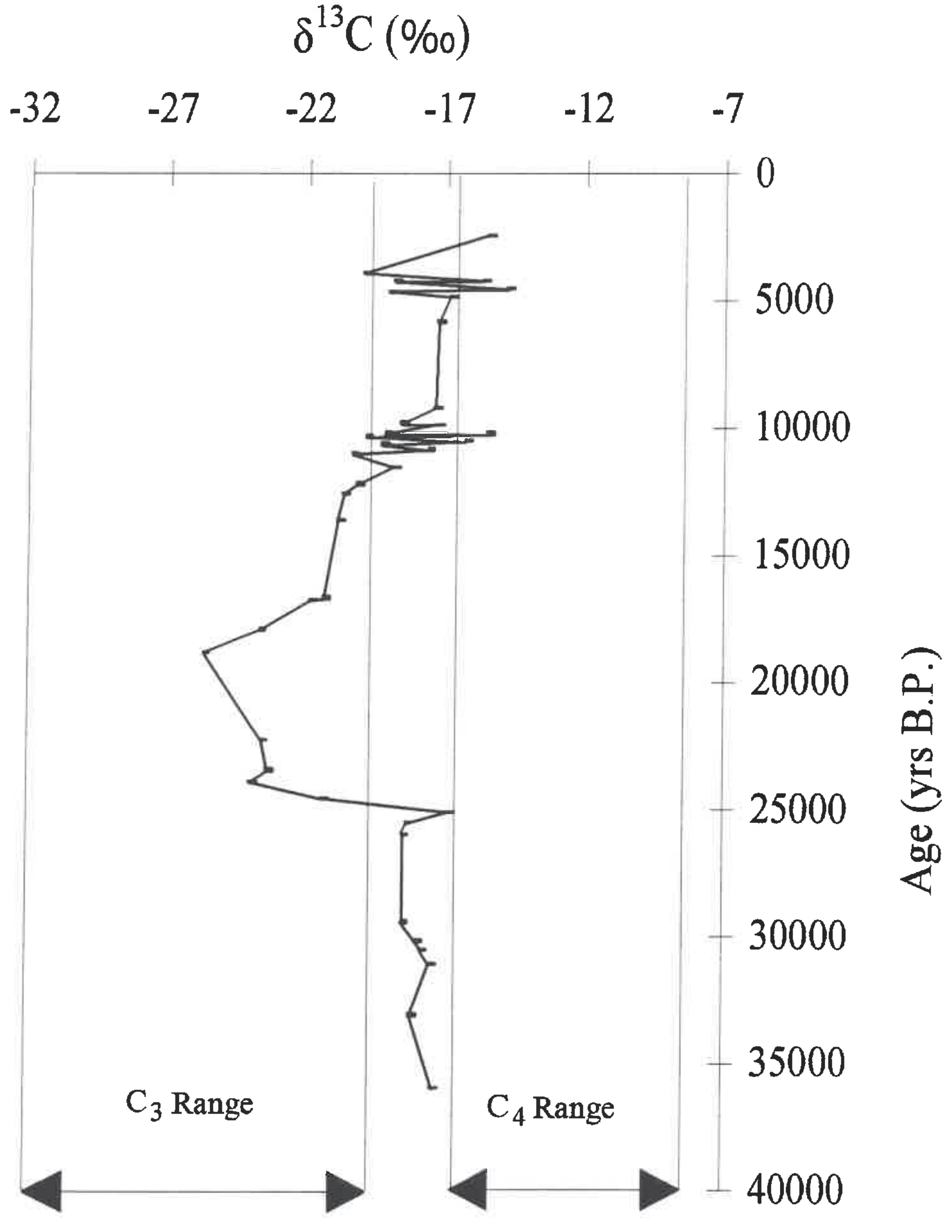

FIGURE 10. $\delta^{13}C$ values obtained from loess deposits of Kansas and Nebraska. The ranges in C_3 (cool-season) and C_4 (warm-season) plants indicate mixed communities for much of the last 35,000 years. (after Johnson 1993b)

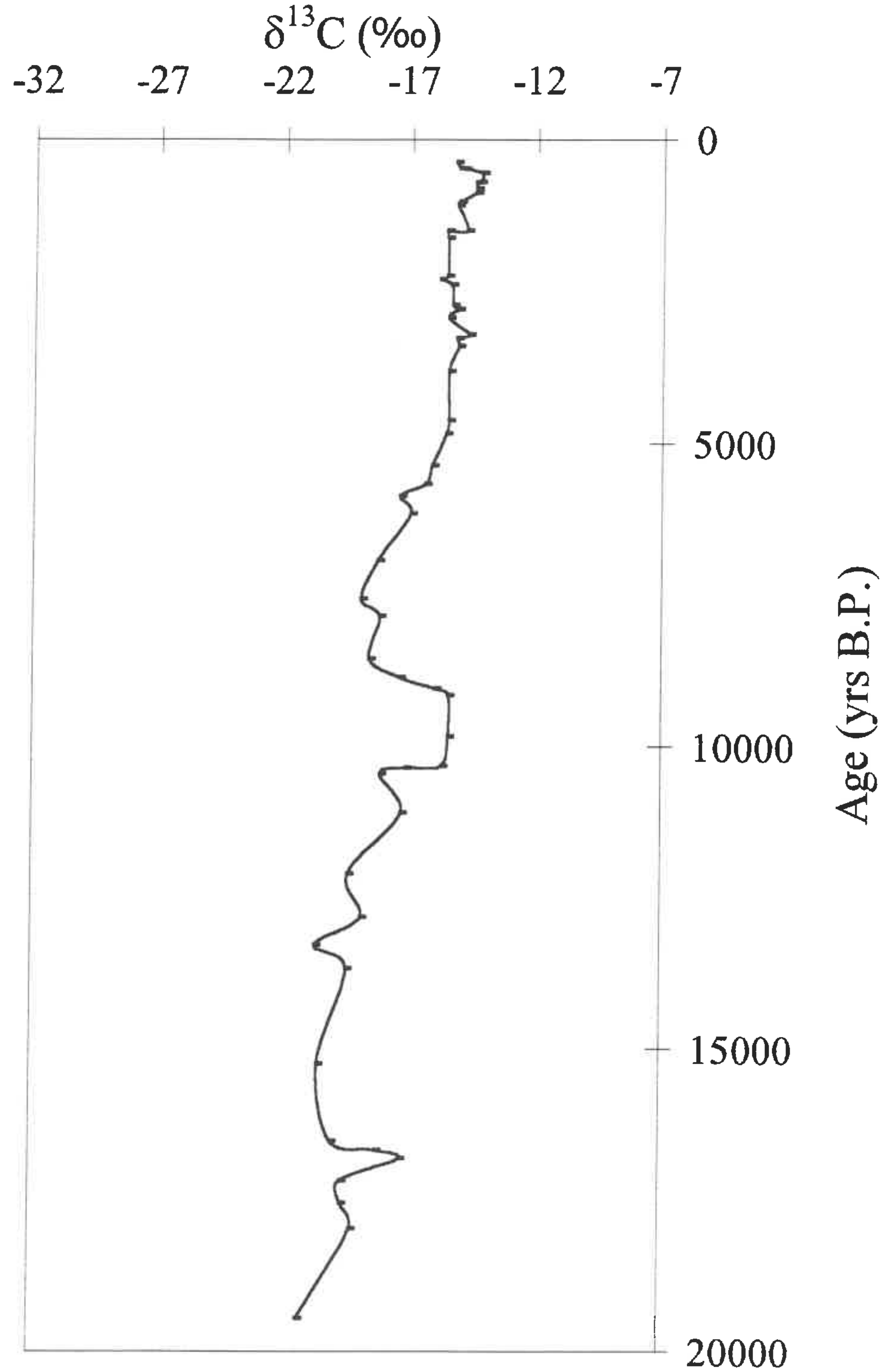

FIGURE 11. $\delta^{13}C$ values from the Great Bend Sand Prairie, Kansas for the past 21,000 years; data are smoothed. (data from Arbogast 1995)

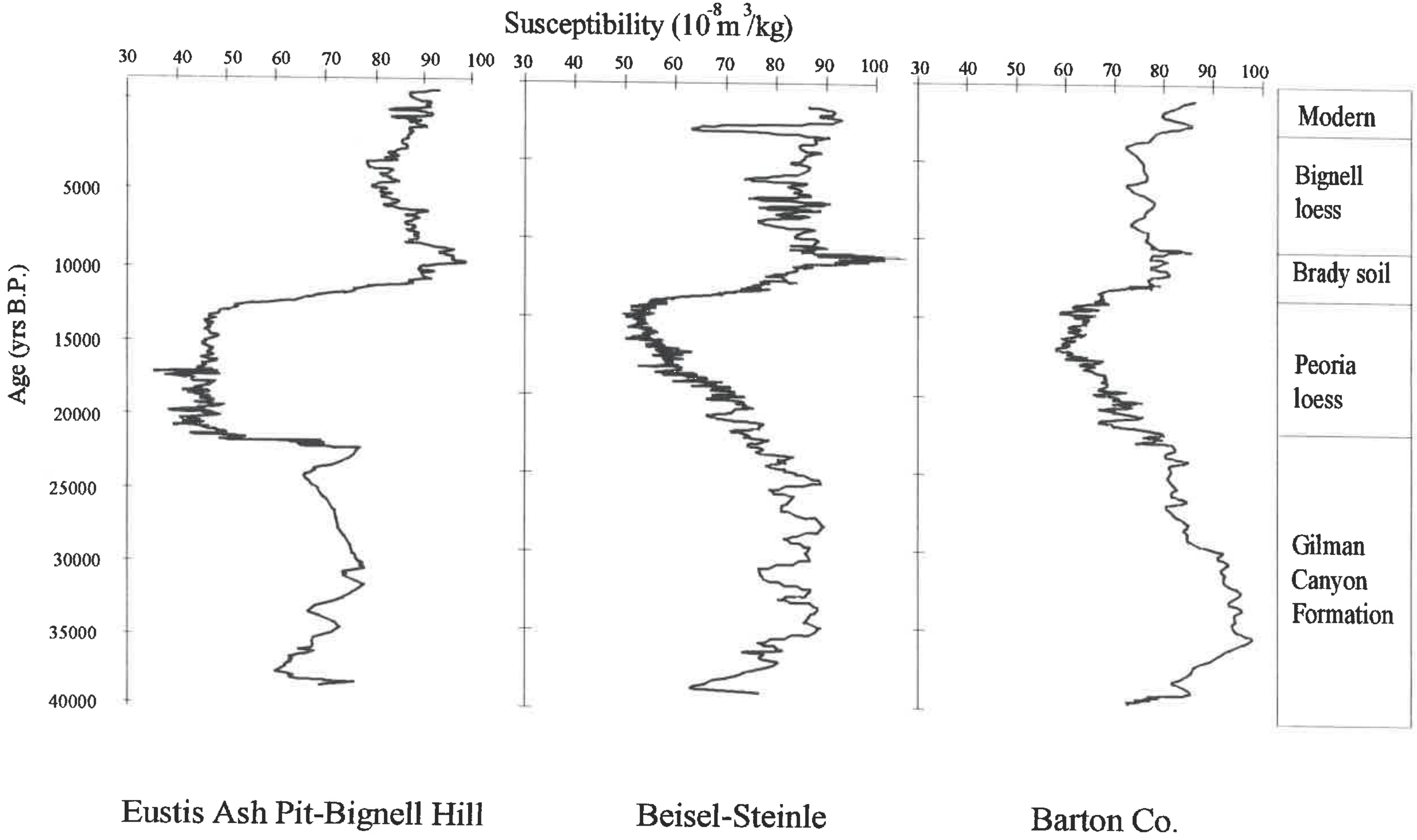

FIGURE 12. Magnetic susceptibility from sites in south-central Kansas (Barton County, Beisel-Steinle) to southwestern Nebraska (Eustis ash pit, Bignell Hill). The Gilman Canyon Formation and Brady soils are well represented in the curves, as is the Last Glacial Maximum. (Johnson 1993b)

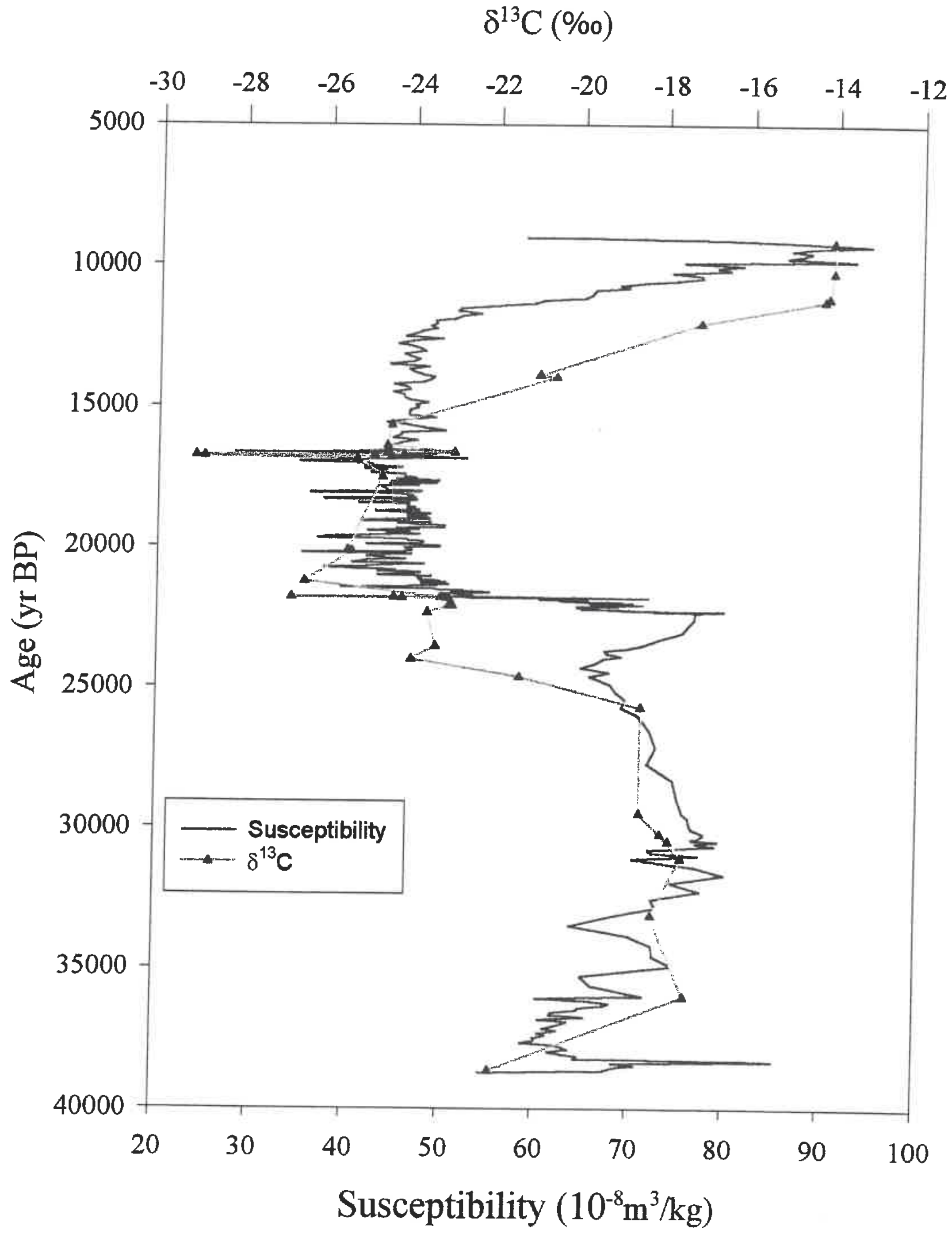

FIGURE 13. Well-expressed agreement between the stable carbon isotope and magnetic susceptibility curves from the Eustis ash pit, Nebraska. (Johnson 1996c)

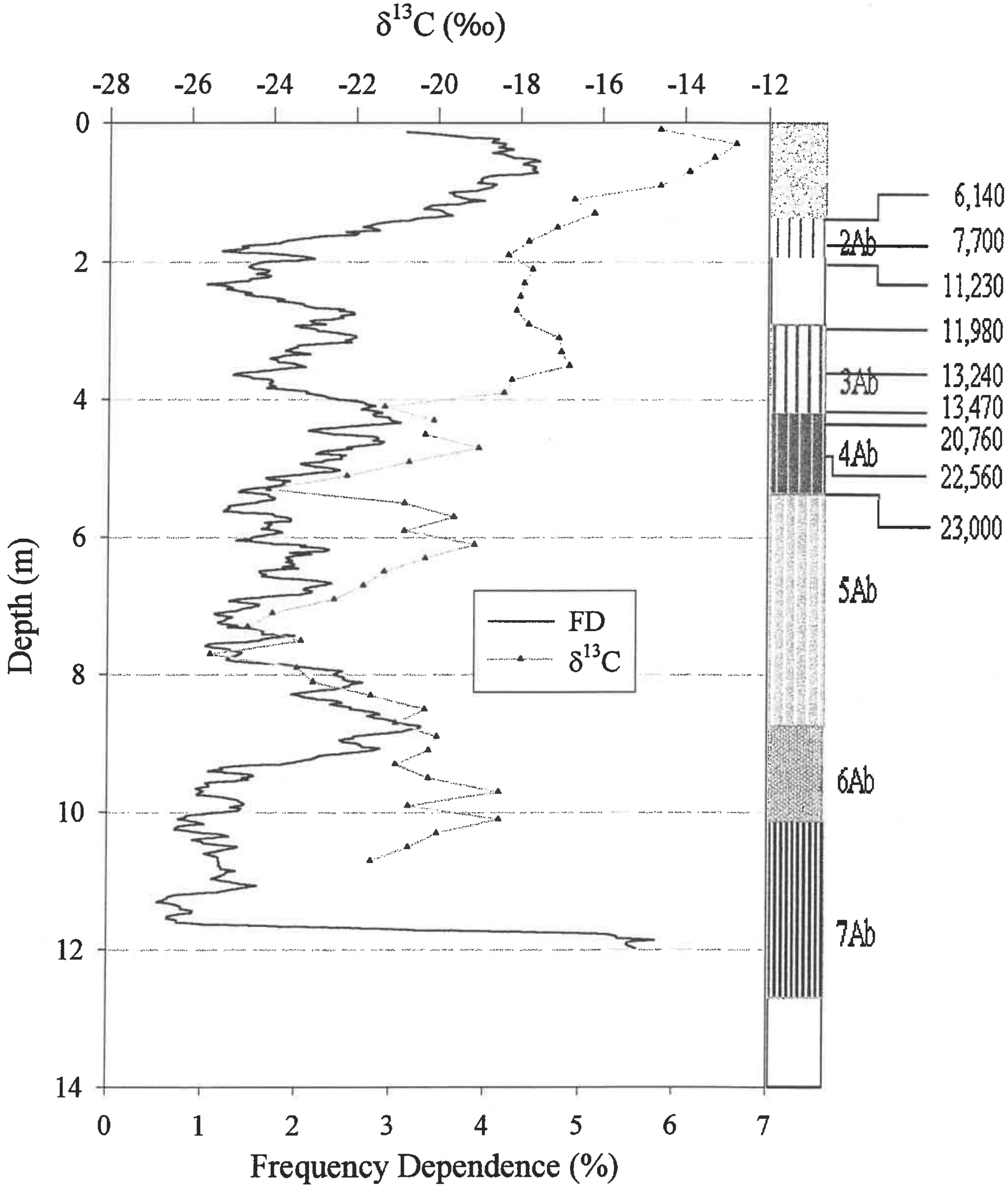

FIGURE 14. Magnetic frequency dependence and $\delta^{13}C$ curves from Sumner Hill on Fort Riley, Kansas. Humate-derived radiocarbon ages are indicated on the right. (Johnson 1996b)

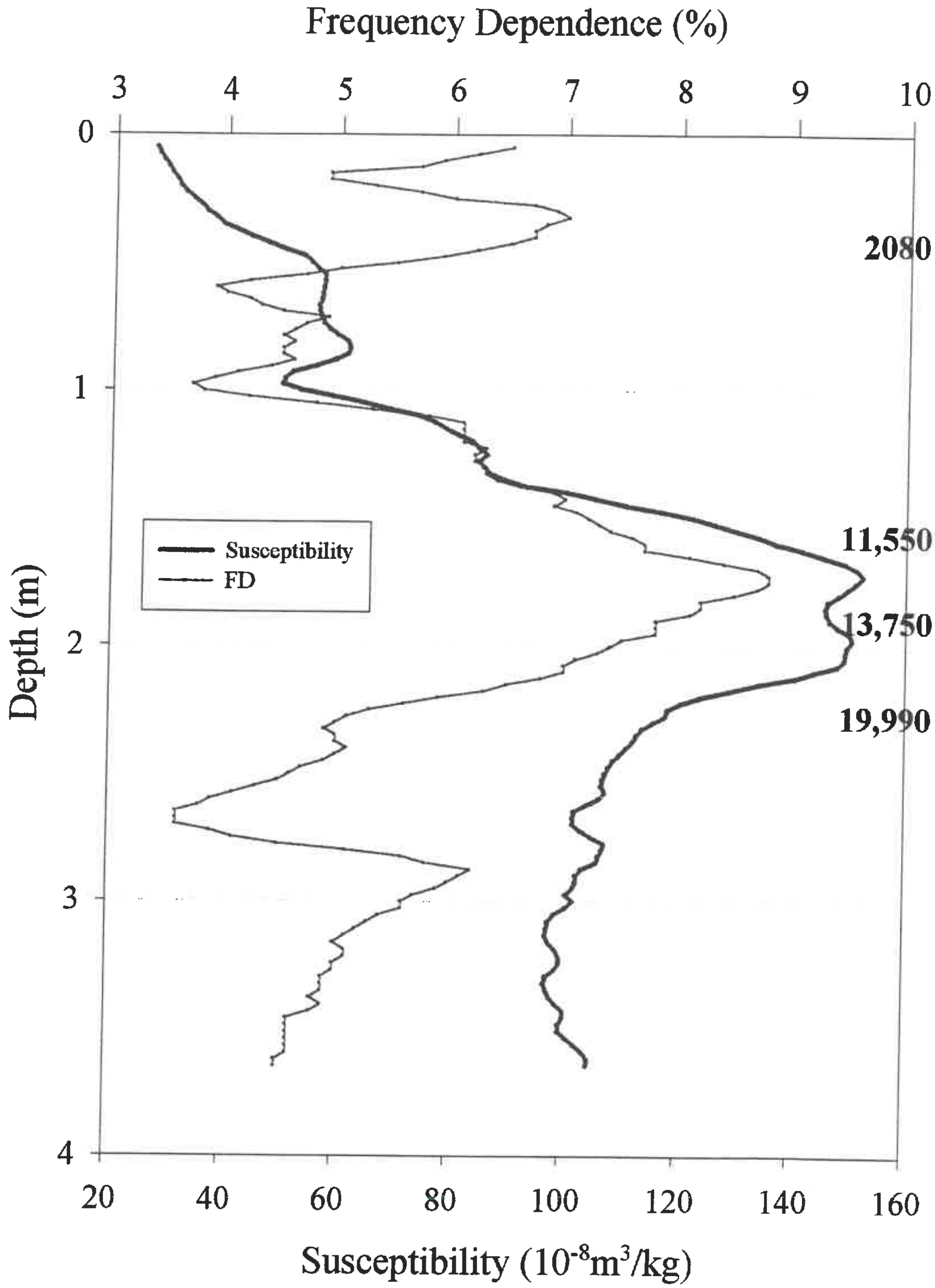

FIGURE 15. Magnetic susceptibility and frequency dependence from the Manhattan Airport site on Fort Riley, Kansas. Humate-derived radiocarbon ages are indicated on the right. (Johnson 1996b)

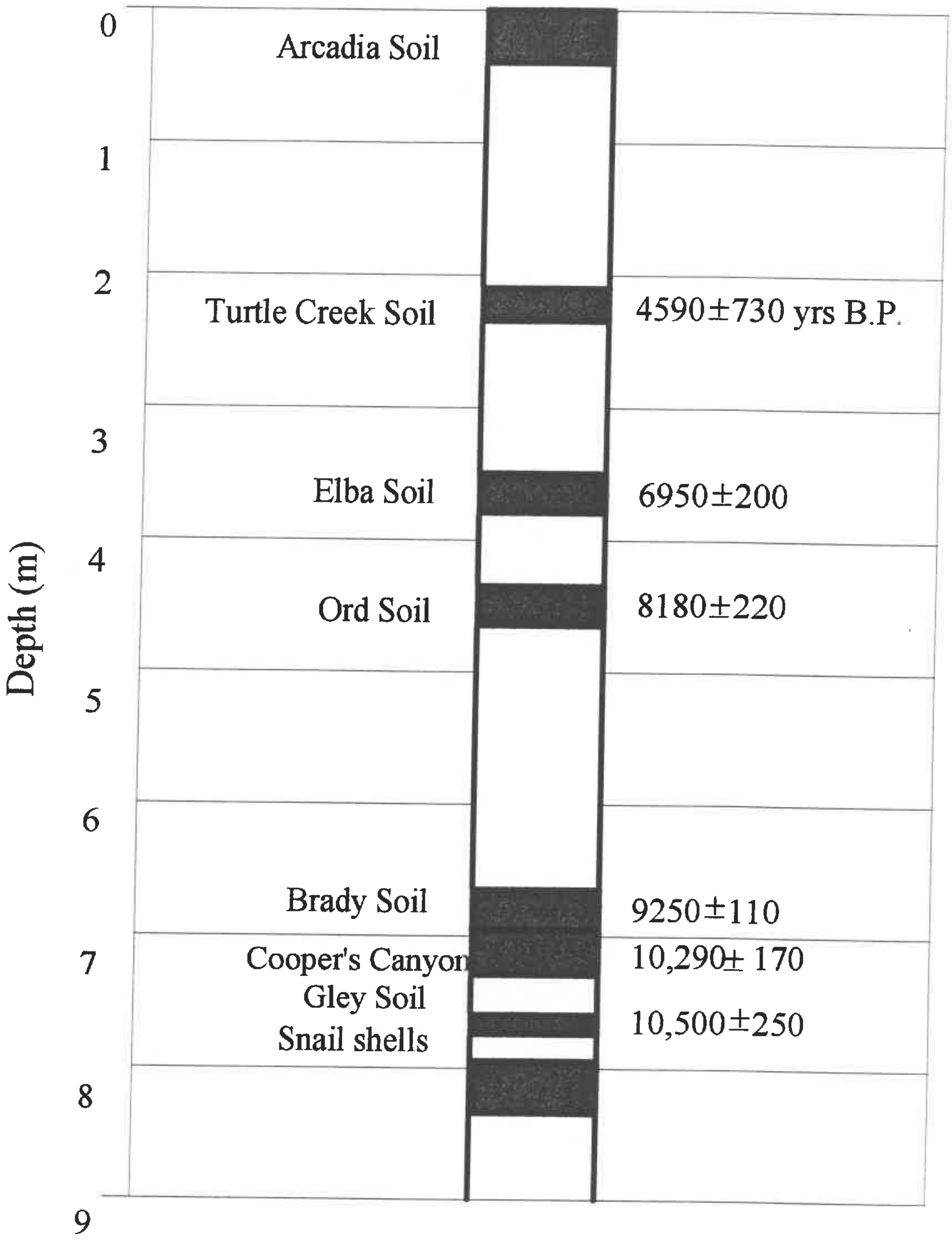

FIGURE 16. Cooper's Canyon section, located near Elba, east-central Nebraska (May 1990, 1991).

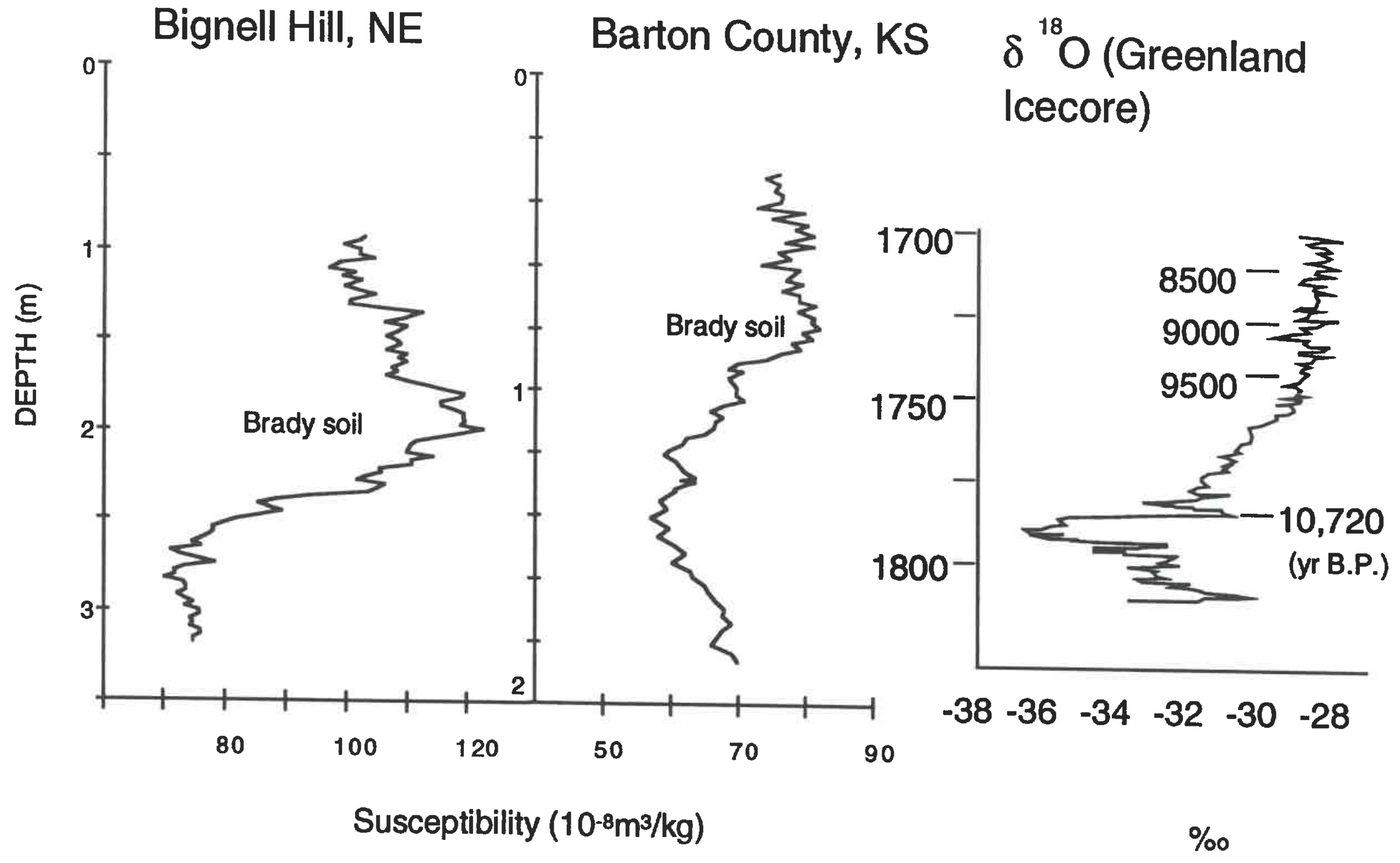

FIGURE 17. Magnetic susceptibility from two loess sites, and the $\delta^{18}O$ curve from a Greenland ice core. A Younger Dryas-type climatic fluctuation is apparent at the Pleistocene/Holocene boundary (magnetic data from Johnson et al. 1993, $\delta^{18}O$ data from Dansgaard et al. 1989).

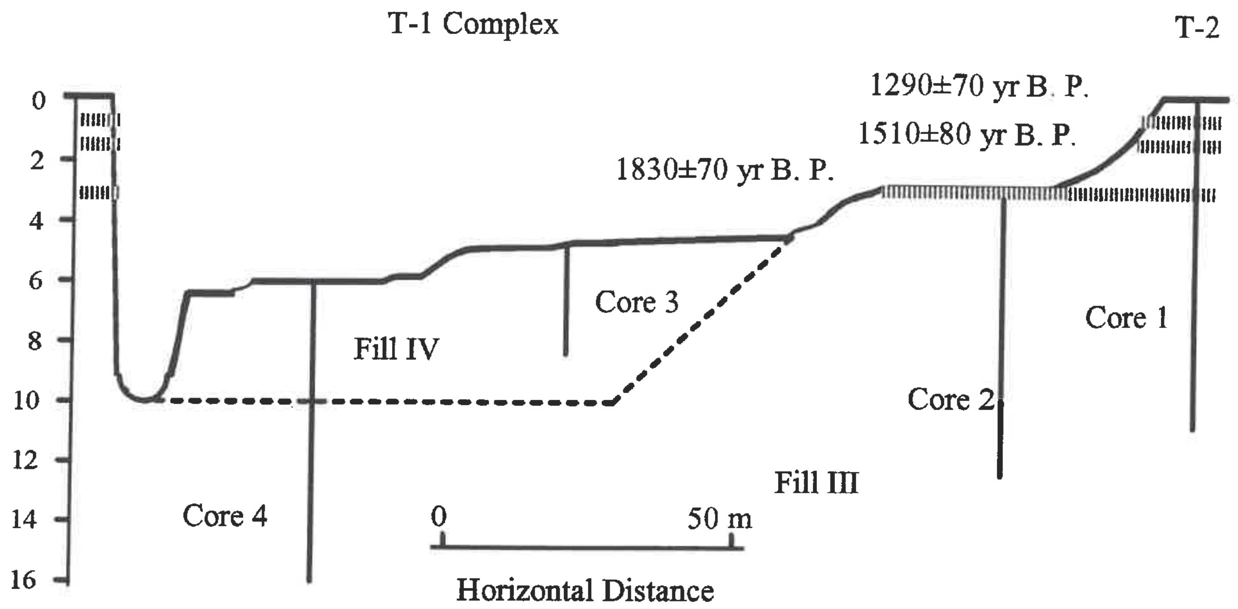

FIGURE 18. Late-Holocene alluvial stratigraphy from Wolf Creek in west-central Kansas. The T-1 surface complex has been created through cutting of the T-2 and through cutting and filling. (after Arbogast and Johnson 1994)

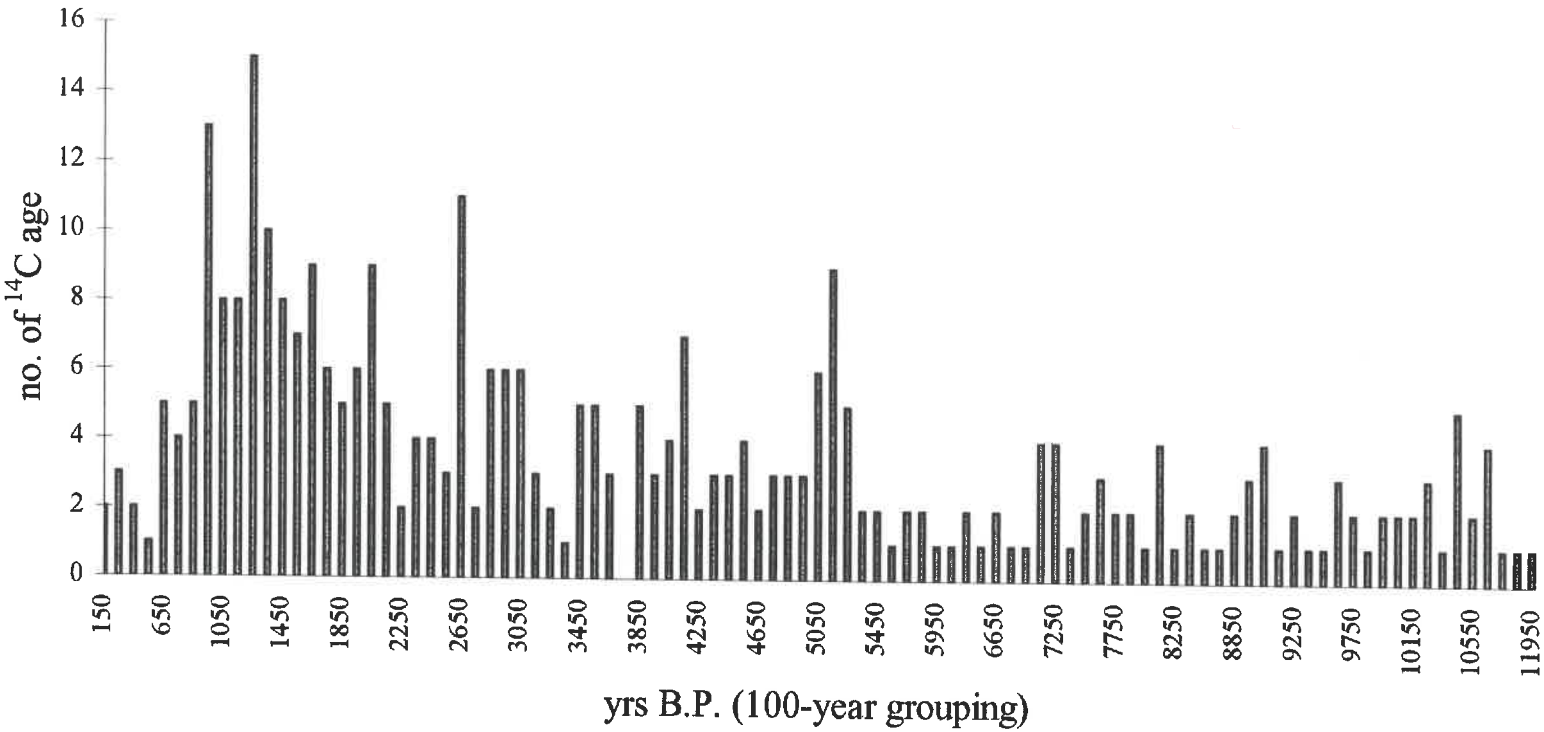

FIGURE 19. Frequency distribution of alluvial radiocarbon ages from the Holocene obtained in Nebraska and Kansas. (data from Johnson et al. 1996)

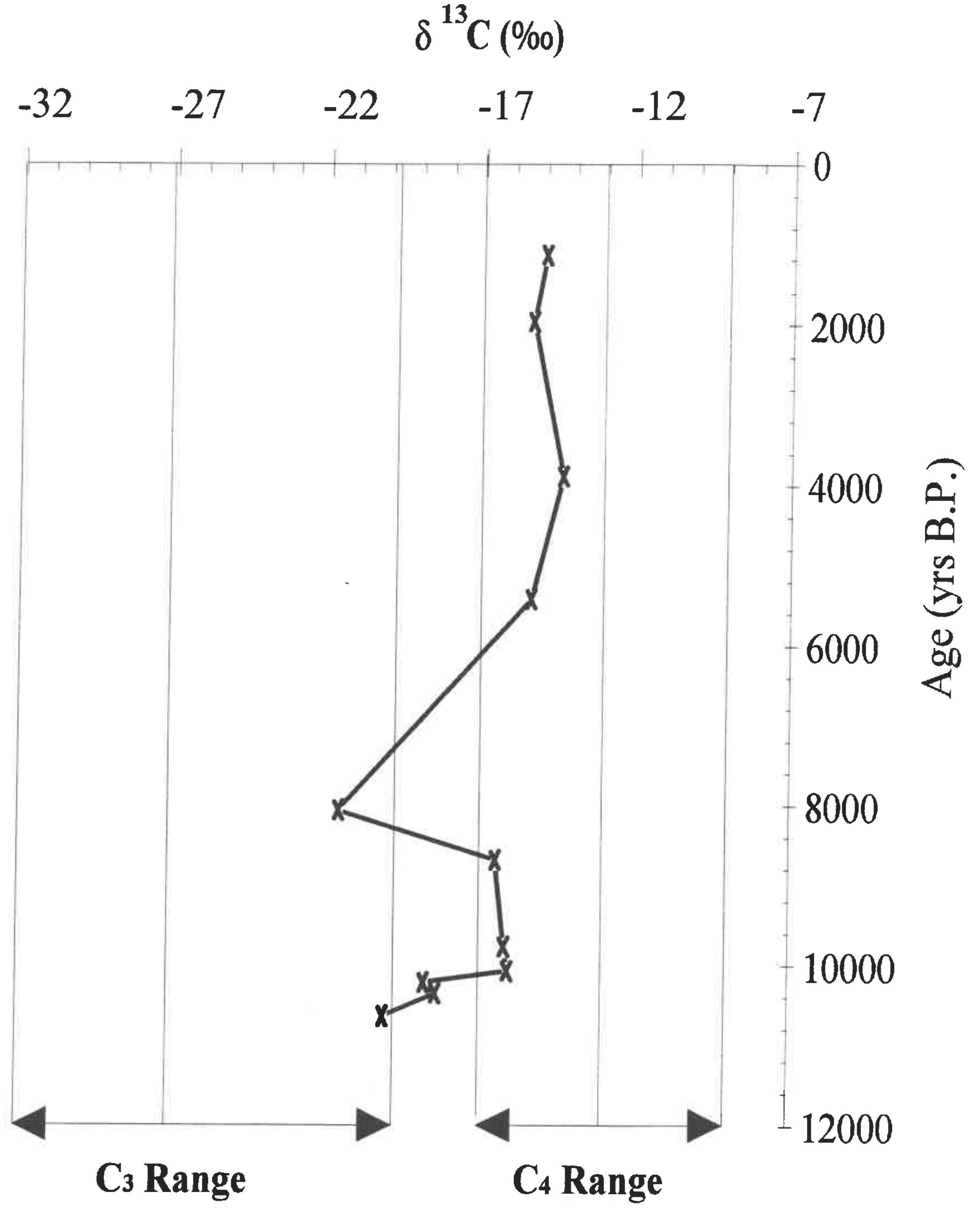

FIGURE 20. Composite stable carbon isotope values from the Sargent site, southwestern Nebraska for the past 11,000 years. Values are by-product of correcting radiocarbon ages for effects of fractionation. (Dort 1996)

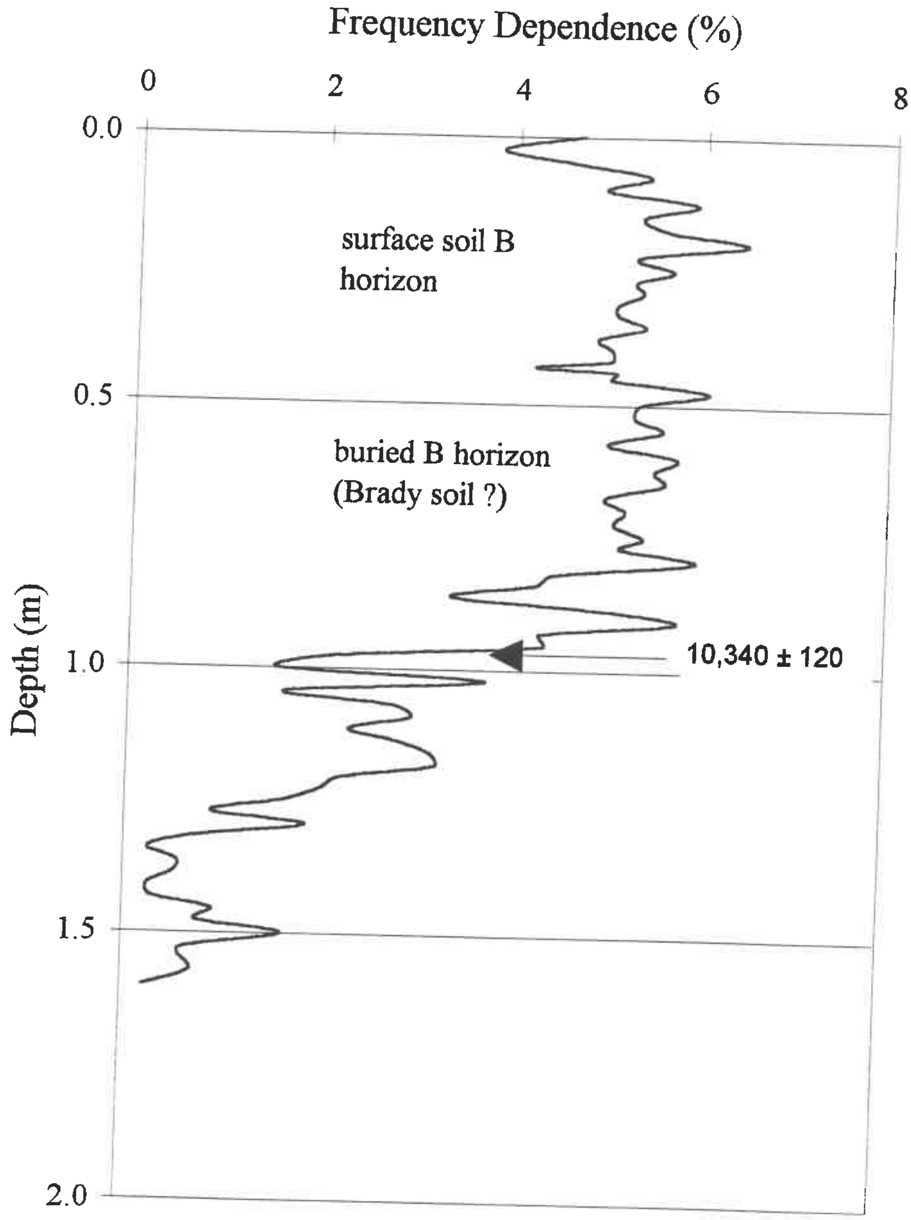

FIGURE 21. Magnetic frequency dependence from the DB site on Fort Leavenworth, northeastern Kansas. (Johnson 1996a)

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