地理學論叢 第32號(1998. 8), 1-24.

Climate History of the Central Great Plains During the Late-Wisconsinan Stage

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Abstract : Global circulation models (GCM) project enhanced warming and drying in the central Great Plains during the next few decades due to increased level of green house gases. Given the sensitivity of loess mantled landscapes, paleoenvironmental research has focused on the response of loess mantle and prairie vegetation through the use of climate proxies. Floral (pollen, phytolith, and macro-fossils) and faunal(snail) remains as well as $\delta^{13}C$ (stable isotope) values derived from the loess indicate that late-Wisconsinan climate was cooler with more effective moisture. Around 13-12 ka climate began to ameliorate. Although appreciable research has been conducted in the region these days, no complete paleo-climatic variations has yet reconstructed from the central Great Plains.

Key words : GCM, late-Wisconsinan, paleo-environment, loess, central Great Plains, climate proxies

요약 : 대기대순환 모델은 온실가스의 증가로 인하여 향후 미국 중부지방의 기온이 상승하고 건조해지는 방향으로 변화할 것으로 예측하고 있다. 뢰스가 피복하고 있는 지역특성으로 인해, 기후변화는 직접적인 식물학적 증거보다는 간접적인 기후지시자를 통해 이루어져 왔다. 연구결과에 따르면, 지난 위스콘신 빙하기 때에는 현재보다 훨씬 냉량하고, 상대 습도가 높게 나타나고 있다. 약 1만3천년-1만2천년 전부터 기후는 온난해져서 현재와 유사한 기후가 형성된 것으로 보인다. 이 연구는 지금까지 이루어진 고기후 연구 성과들을 정리하고 아직도 밝혀지지 않은 부분이 무엇인가를 정리하고자 하는 취지에서 이루어진 연구의 결과이다.

주요어 : 대기대순환모델, 후기 위스콘신 빙기, 고환경, 뢰스, 미중부평원, 기후지시자

Ⅰ. Introduction

This article is intended to introduce what is currently known about the paleoenvironmental history of the central Great Plains of the United States. 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 database.

Less is known about environmental conditions in the central Great Plains during the Late Pleistocene and Holocene than about many other regions of North America. This is probably because wide application of the more traditional investigative tools such as palynology and dendroclimatology is very difficult. Consequently, what is known has been and is being derived

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using some of the newer approaches to late-Quaternary environmental reconstruction such as stable isotope analysis, opal phytolith analysis, and environmental magnetism. Those sedimentological contexts being explored include loess, eolian sand sheets and dunes, alluvial fills, 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.

Ⅱ. Current Climatic Setting

The principal climatic features of this region are its continentality and increasing dryness towards the mountains. 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). This region is in the rainshadow of the Rocky Mountains: isohyets (equal precipitation) are approximately longitudinal, and mean annual precipitation decreases from about 750 mm at the southeastern margin to less than 380 mm in the western and northern parts. 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 (fig.1).

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 almost wholly prairie grassland, due to the subhumid-semiarid, markedly seasonal climate. The mean tropical Atlantic (mT) airflow

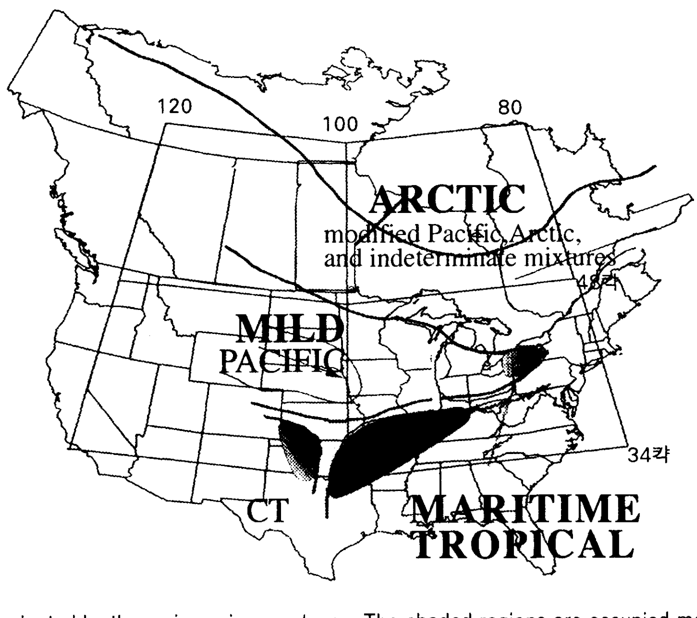

Figure 1. Regions dominated by the various air mass types. The shaded regions are occupied more than 50 percent of the time by the indicated airmass.

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.

The prairie crosses the region from north to south in three broad zones (fig.2)

In the west, the grama-buffalograss prairie consists of short grasses while the blue stem prairie with its tall grasses and many forbs prevails in the east. Between them lies the mixed prairie with tall, medium tall, and short grasses (K chler 1964; 1974). In the sand sheets of southeastern Colorado and westcentral Nebraska and central Kansas, edaphic conditions promote the existence of a sandsage-bluestem prairie. The sensitivity of prairie vegetation composition and boundaries to short-term climatic variation during the historical period is well documented for the region (Tomanek and Hulett 1970; K chler 1972). Similarly, long-term prairie expansion and contraction, presumably in response to climatic variation, is documented for the prehistoric time scale (e.g., Watts and Wright 1966; Gr ger 1973; 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. The magnitude of this is, however, certainly open to question (cf., Wedel 1961; Reeves 1973; Benedict 1979; 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 marked less rainfall than the remainder of Anglo-American 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. The grassland is characterized by large positive departures from

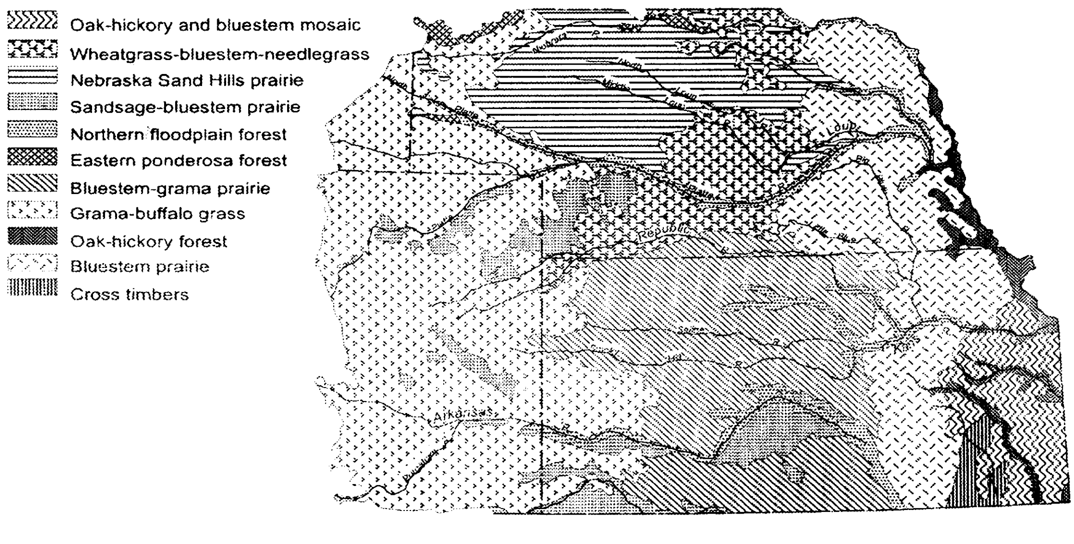

Figure 2. Potential vegetation map in the region (after Küchler,1964)

average temperature and by frequent hot winds during summer days.

Ⅲ. Physiographic Regions

The Great Plains physiographic region lies east of the Rocky Mountains and extends from southern Alberta and Saskatchewan nearly to the United States-Mexico border. 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 (Boelstorff 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 joined with the generally 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 bit of Ozark Plateau extends into Kansas. The streams of this region have carved the thick flint-bearing Mississippian limestones of local bedrock into modern 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, sluggish streams have planed the surface of this region almost flat.

West and north of the Cherokee Plain the topography consists of a series of parallel northeast-southwest trending cuestas. Cuesta topography is developed on a series of ridges having a sharp slope on the east side and a gently sloping on the west side.

Cuesta topography is developed on a series of relatively erosion-resistant limestone layers, exposed at the surface, which descend gently westward until they dip under the outcrop of thicker overlying shale. The shale that underlies the capping limestone, which is less resistant to weathering, erodes to form an abrupt east-facing escarpment.

To the west of the Osage Cuestas is a band of grass-covered Flint Hills that constitutes the preserve of large remnants of the Kansas tall-grass prairie. These are located at the eastern edge of huge expanse of grass-covered plains that extends continuously westward to the front ranges of the Rocky Mountains, northward into Canada, and southward into northern Texas. At their western margins of the Flint Hills dip gently under younger rocks which , in the north, slope gently westward under the Smoky Hills escarpment. In the south, these rocks dip below the McPherson-Wellington Lowlands. Extending from north of Salina southward to the Oklahoma border, the

McPherson-Wellington Lowlands mark the outcrop belt of thousands-foot-thick Wellington shale.

Along the southern border of Kansas, from Harper and Kingman counties to eastern Meade County, erosion has exposed Red Hills. The topography of the Red Hills is unique in Kansas. In some areas, erosion resistant dolomite cap the red-colored hills to form small buttes and mesas.

Extensive areas of grass-covered sand dunes lying south of the Arkansas River constitute the Great Bend Sand Prairie, which was formed over the last fifteen million years by the river's meandering deposition of sediment. A similar region exists south of the Cimarron River in the southwestern corner of Kansas. During the Late-Pleistocene and Holocene, strong winds picked up fine sediments from alluvial beds of the Arkansas River and piled in the dune area, which covered hundreds of square miles. North of the Great Bend Sand Prairie, cretaceous rocks are exposed over a large portion of Kansas.

These are the Smoky Hills named such because of their dark shales.

From Saskatchewan in the north to northern Texas in the south, lies the High Plains. Viewed from a broad perspective, the whole of the High Plains surface is upheld by several huge wedge-shaped alluvial fans of sediment derived from stream erosion of the eastern Rocky Mountains. These sediments are called the Ogallala Formation, and they represent tertiary stream deposition identical to that still occurring in the Arkansas River farther south. The Ogallala Formation is composed not only of river-borne sands and gravels but also windblown loess, volcanic ash beds, and diatomite deposits.

In the eastern part of central Great Plains is an area of rolling hills that, during the ice age about two million years ago, was invaded by two major advances of the great continental glaciers. The first major glaciation, called the Nebraskan, barely made it into Kansas. The second, the Kansan, overrode a huge area east of Blue River near Manhattan. Its southern margin lay at about the present location of the Kansas River, which was pushed south by the snout of glacier, whose slow-moving ice may have been as much as mile thick.

In a large portion of central Nebraska, the central Great Plains is underlain by extensive deposits of late-Wisconsinan to late-Holocene eolian sands known as the Sand Hills. The age and origin of this spectacular eolian features are still uncertain. Ahlbrandt et al. (1980) suggest that the dunes are late-Holocene features, possibly derived from older, unconsolidated sediment that mantled the Plains, and therefore may be genetically unrelated to the loess. In contrast, Wells (1983) regarded the Sand Hills as a coarse, upwind facies of a single late-Pleistocene sand-silt unit. Regardless of specific origins, the eolian deposits that are in large part responsible for giving much of the Great Plains its relatively flat topography were mostly derived from valley alluvium of streams such as the Niobrara, Platte, and Pecos rivers, and from outwash carried by these streams and others headed by glaciers during the Pleistocene time.

There are three dune fields in northeastern Colorado (Muhs 1985). The Greeley dune field, immediately north of the South Platte River, is the smallest; the Fort Morgan dune field lies south of the river; and 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. Northeast winds produced the Fort

Morgan and Wray dune fields, probably in late-Holocene time. The source and direction of movement of the Freely sand hills is uncertain (Muhs 1985).

Ⅳ. Paleoenvironment Reconstructed from the Proxies

1. Late Wisconsinan Stage

Due to its relative youth, the Late 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 include the Peoria loess, Brady soil and Bignell loess (fig.3).

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 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 have seen 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 do most marine records.

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 13C data from buried soils, as an independent variable, and a climatic element or complex of elements, expressed as a

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<tr><td>Time-stratigraphic units</td><td colspan="6">Rock-stratigraphic units</td></tr>
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<tr><td>Recent stage</td><td colspan="6">Eolian and fluvial deposits</td></tr>
<tr><td>Wisconsinan Stage</td><td>Bignell Formation</td><td>Fluvial deposits</td><td>Bignell Formation</td><td>Fluvial deposits</td><td>Bignell Formation</td><td>Fluvial deposits</td></tr>
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<tr><td colspan="2">David City Formation</td></tr>
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Figure 3. Late-Quaternary stratigraphic succession in Kansas(Bayne and O' Connor, 1968).

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.

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 information content of the fossil pollen assemblage (Delcourt and Delcourt, 1980). 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 limiting, several sites in the region have produced a picture of past environments.

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 United States 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 trees replaced *Picea* about 11ka, and prairie developed 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 suggest 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 Gruger (1973), a rather open vegetation with some 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. At Boney Spring, a mixed vegetation with some pine, willow, and sedge gave way to a spruce dominant forest at about 21,500 yr B.P. (King 1973). 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 (1987), 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 vegetation history. It is clear is that prairie vegetation existed sometime after 12,600 yrs 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 presence 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.

The closest vegetation type showing such a compositional mix can be found along the southern rim of the boreal forest on the Canadian Prairies. There, 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, however, were not 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.

Stable carbon isotopes

Regionally, 13C analyses have been applied to a variety of Quaternary soil fractions including soil organic matter, soil carbonate, and opal phytoliths (e.g., Fredlund and Tieszen, 1997; Nordt et al., 1994; Humphrey and Ferring, 1994; Kelly et al., 1993; Fredlund, 1993).

The natural difference in the stable carbon isotopic composition of C3 and C4 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, C4 plants discriminate less against isotopically heavier $^{13}CO2$ than do 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°/₀₀, with a mean of -13°/₀₀. Thus, C_3 and C_4 plant species have distinct, non-overlapping $\delta^{13}C$ values (Nordt *et al.,* 1994).

The stable isotope ratios for $^{12}C/^{13}C$ are measured by isotope ratio mass spectrometry, and the isotopic data are expressed as the difference, or delta value (δ), between the sample or standard times 1000.

$$13C = \frac{Rsample - Rstandard}{Rstandard} \times 1000$$

where R = $^{13}C/^{12}C$

The value for a carbon isotope in soil organics is defined as follows:

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

where $^{13}C_{C4}$ is the average of ^{13}C values of C_4 plants (-13‰); ($^{13}C_{C4}$) is the average of $\delta^{13}C$ values of C_3 plants (-27‰); and x is the proportion of carbon from C_3 plant sources.

The $\delta^{13}C$ value of soil organic matter or pedogenic carbonates formed largely from respired CO_2 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).

For the Gilman Canyon Formation, $\delta^{13}C$ values exhibit a good correlation with coincident phytolith data. $\delta^{13}C$ 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 (fig. 4).

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, indicating that vegetation and thus climate during Gilman Canyon time was similar to present warm, semiarid conditions in the central Great Plains (Johnson, 1993).

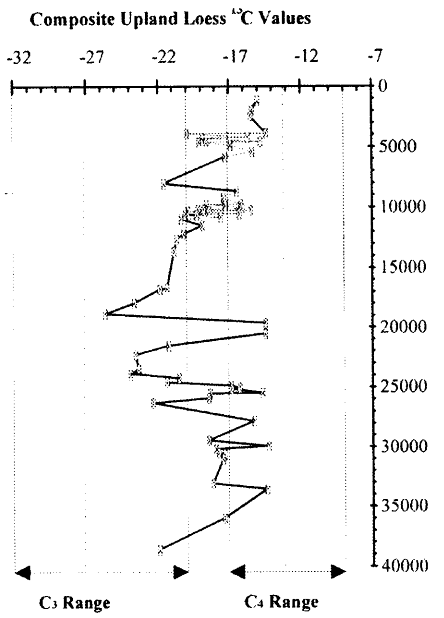

Figure 4. Composite stable carbon isotope values fromthe region for past 40,000 years. Most values are by-product of 14C dating. (Source: W.C. Johnson unpub. data)

Site specific factors should be borne in mind when interpreting $\delta^{13}C$ data from soil humates. For example, the 17,000 yrs 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°/••. During the Last Glacial Maximum, the dune temporarily stabilized and a soil formed. A $\delta^{13}C$ ratio of -11.9°/•• 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; Arbogast and Johnson, 1997).

Opal phytoliths

Although recent studies confirm that δ13C of soil organic matter in grassland soil samples accurately reflects the relative abundance of C3 and C4 grasses and temperature (Tieszen et al., 1997), δ13C analysis cannot distinguish between the relative contribution of C4 xeric short grasses and C4 mesic tall grasses to soil organic matter (Fredlund and Tieszen, 1997). Fredlund and Tieszen (1997) and Johnson and Bozarth (1996) suggested that short-cell phytolith morphotypes can distinguish among short C4 grasses and C3 grasses.

Growing plants absorb water containing dissolved silica through their roots. Microscopic amorphous silica bodies are subsequently produced by the precipitation of hydrated silicon dioxide ($SiO_2.nH_2O$) within the plant's cells, cell walls, and intercellular spaces. Silica bodies with characteristic shapes are called opal phytoliths.

On the Great Plains, two grass subfamilies commonly employ the C4 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 Chlorideae tribe of the Eragostoideae subfamily are the two most important of these grasses in the arid southwestern region of the Great Plains. The overall pattern, where festicoids (C_3) dominate the cool northcentral Great Plains, panicoids on the moist, warm eastern and southeastern margins, and chloridoids primarily in the western and southwestern Plains, is consistent with the pattern expected from general C_3 and C_4 adaptations of grasses.

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.

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 buried soil complexes indicated that the soil forming periods must have been warmer and drier than the periods of loess accumulation. The phytolith assemblages from the soils of the Gilman Canyon Formation are 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 the loess deposition and other proxy records.

The basic taxonomic system defines those opal phytolith types common to the short C_4 grasses (Chloridoideae subfamily), the tall C_4 grasses (Panicoideae subfamily), and those produced by primarily by the cool -season C_3 grasses (Pooideae subfamily). Based on the classification system used by Twiss *et al.* (1969), Diester-Haass *et al.* (1973) suggested that grass opal phytoliths could serve as indicators of continental climatic change. Diester-Haass *et al.* (1976) calculated humidity-aridity tendencies using the following formula:

$$Tp = \frac{chloridoid}{chloridoid + panicoid} \times 100$$

where Tp indicates the phytolith index. Since chloridoid type is characteristic for an arid grass assemblage, panicoid type for a humid assemblage, higher values indicate increased aridity, lower values indicate increased humidity. Ongoing opal phytolith analysis of the Peoria loess is producing a climatic signal consistent with that of the carbon isotope data (Johnson *et al.,* 1993). 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.

Snail assemblages

In central and western parts of Kansas where Late Pleistocene deposits are less severely affected by weathering, assemblages of fossil mollusks greatly exceed in variety of species, and in population density, the local living molluscan fauna (Leonard, 1952). During the Peoria time, approximately 30 species of landsnails were distributed on the interfluvial uplands of the Great Plains (Wells and Stewart, 1987).

Leonard (1952) recognized two main faunal zones and one transitional zone within Peoria loess of Kansas. A basal zone is devoid of fossil mollusks, but it is correlative with Gilman Canyon Formation. The basal zone silts were deposited so slowly that the substantial weathering took place. The depositional rates gradually accelerated with time until rate of accumulation outstripped rate of weathering of the basal zone.

Ecological conditions during the Peorian deposition were relatively favorable to terrestrial gastropods, since the fossil faunas were more varied than are the modern faunas in the same area. A reasonable amount of rainfall and a floral cover at least as dense as that prevailing now has been inferred to have existed over the area in Kansas in Peorian time, since terrestrial gastropod are active and can produce only during intervals when the soil and overlying organic matter are moist.

Rousseau and Kukla (1994) counted mollusk assemblages within the Eustis ash pit, where the same section and stratigraphic marker has been employed for the detailed rock magnetic measurements for this study. Based on land snail assemblages, three mollusk zones have been distinguished. The first zone (18 to 14 m in depth) is characterized by small number of species and the lack of Discus or Columella alticola. This points to a moderately cool and relatively moist climate and to the presence of grassland. The second zone (14 to 10 meter) is distinguished by the presence of Discus and the first appearance of the Cordilleran-boreal species Columella alticola which indicates some arboreal vegetation was present. The third zone (10 to 1 meter) is characterized by abundance of recognized species and seems to correspond to an interval of extreme cold, with low and gradually decreasing annual precipitation concentrated in a short growing season.

2. 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 yrs 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 (Bard and Broecker, 1992).

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 that 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.

Climatic Proxies

Evidences of climate change during late Pleistocene/Holocene are derived from the limited number of fossil pollen and macrofossils and d13C data. Such proxies contain a climatic signal.

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, Gruger (1973) documented spruce forest from 23,000 to 15,000 yrs 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 yrs 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) have suggested that such 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 reintroduction 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, 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).

Stable carbon isotopes

Temporal changes in $\delta^{13}C$ data derived from carbon contained within soil and sediment 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, the climate 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.

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 yrs B.P. As this decline occurred, deciduous tree species increased until about 9000 yrs B.P. From a site in central Texas, Nordt et al. (1994) interpreted the time between 11,000 and 8000 yrs B.P. as transitional between late-Pleistocene conditions and warmer and drier Holocene conditions based on a slight increase in the abundance of C4 plant biomass using stable carbon isotopic data.

3. Holocene

The driving mechanism behind the region s environmental change was disintegration of the Laurentide ice sheet (Andrews, 1987), which promoted more arid, zonal atmospheric flow (Knox, 1983). At Muscotah Marsh, the combined effects of increased solar radiation (Kutzbach, 1985, 1987) and increased zonal flow resulted in the complete displacement of forest by grassland till 9000 yrs B.P. (Gruger, 1973).

As the Laurentide ice sheet continued to waste during the early Holocene, the steep north-south temperature gradient which had been present during the late-Wisconsin continue to weaken, promoting further zonal flow. These factors triggered the generally warm and dry conditions of the Altithermal that prevailed in the central North America from about 8000 to 5000 yrs B.P. (Knox, 1988; COHMAP Members, 1988).

Climatic Proxies

Some of the climatically-sensitive parameters that have recently been examined in the central Great Plains include fossil pollen, opal phytoliths, and stable carbon isotopes. Recent archaeological investigation in DB site and Fort Riley yielded climatic proxy records including stable carbon isotopes and opal phytoliths during the Holocene (Johnson and Park, 1997b).

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, 1987).

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 yrs 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 yrs 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 with no Altithermal signal.

Two sites in Kansas provide palynological information for the Holocene: Muscotah Marsh (Gruger, 1973) and Cheyenne Bottoms (Fredlund, 1995). The Holocene portion of the record at Muscotah Marsh in north-central Kansas contains unconformities and lack 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) also divided the Holocene into four microzones based on changes in the local pollen signal. The latest Pleistocene-earliest Holocene zone (>9,690 yrs 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 yrs 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 yrs 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 yrs B.P., peaking at 7,200 yrs 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 yrs B.P. and culminated 6,000-4,500 yrs B.P., probably because of warmer, drier conditions that reduced vegetation cover (Holliday, 1989). In the northern Great Plains, the atmospheric anomalies that produce drought today were more frequent and persistent before AD 1200 (Laird et al., 1996).

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 yrs B.P., except for a brief drying period between about 12,000 and 11,000 yrs 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 yrs B.P. (Holliday 1985, 1989; Pierce 1987). $\delta^{13}C$ values derived from buried soils 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 18O values from these same carbonates did not indicate a significant temperature change.

Limited δ13C data from the Sargent site, an upland loess exposure in southwestern Nebraska, suggest a gradual increase in dryness through the Holocene; this is interpreted as a shift in the abundance of C4 species from slightly under 50% during the late Pleistocene to 80 - 90% in the middle Holocene. δ13C 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 C4 plants from about 12,000 yrs B.P. through the Holocene, but these data are relatively noisy due to the edaphic conditions encountered on bottomlands.

Tree rings

Variations in tree-ring widths from one year to the next have long been recognized as an important source of paleoclimatic 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.

Ⅴ. Problems in the records

A substantial body of knowledge exists concerning the paleoenvironmental history of the central Great Plains. This body of knowledge is difficult to summarize because it is unevenly distributed both geographically and chronologically, it is derived from many different types of proxy data using many different types of methods. To extract the paleoclimatic signal from proxy data, the record must first be calibrated. calibration 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 natural phenomenon today. Dendroclimatic studies, for example, have benefitted from a wealth of research into climate-tree growth relationships, which have enabled dendroclimatic models to be based on sound ecological principles (e.g., Fritts 1983). Significant advances have also been made in opal phytolith studies by improvements in our understanding of the relationships between modern climate and modern phytolith data (e.g., Twiss 1987; Fredlund 1993).

However, 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 1987). One must therefore 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.

Chronological control

A major problem in paleoenvironmental studies in the region is that studies tend to produce paleoenvironmental data which lack the chronological control. Chronological control for paleoenvironmental data must exist, or it is almost impossible to relate data from place to place, and the data cannot contribute to regional understanding. Data relating to a particular stratigraphic unit may be related by correlation to other data, but in general, paleoenvironmental data must be associated absolute age for that data to be used.

Tree rings can be accurately dated to an individual year, and may provide continuous records up to several thousand years in duration. With a minimum sampling interval of one year, they provide high frequency (short-term) paleoclimatic information. The central Great Plains appear not to be an ideal place for the long-term preservation of trees. Although Weakley (1962) and Fritts (1983) have demonstrated the potential of dendroclimatology, these reconstructions will be improved only when more tree ring sites are available.

For the most part, age control on late Pleistocene deposits derives from radiocarbon ages of humates extracted from the Gilman Canyon Formation and Brady soil, although wood and charcoal preserved in Peoria loess have provided a few assays. In late Quaternary deposits of the central Great Plains, however, materials for radiocarbon dating are limited. The late Pleistocene vegetation cover of the region provided only scattered wood and charcoal in loess and alluvium; the grass cover of the Holocene has furnished even fewer datable materials. For radiocarbon control in this region, researchers frequently use humates preserved in buried soils. Despite general acceptance of radiocarbon dating by Quaternary scientists, there is considerable debate about the accuracy of ages derived from buried soils.

Thermoluminescence dating (TL) has been employed in several studies (e.g., Feng et al. 1994; Maat and Johnson 1996). TL dating provides an age estimate for the time of burial of mineral grain. TL age estimates from all samples from Bignell Hill and Eustis ash pit were internally consistent. A combination of TL and 14C dating provides a more comprehensive evaluation of ages due to the particular advantages each technique possesses and has the potential to improve the temporal control on loess deposition. TL dating is, however, still somewhat experimental and with uncertain maximum age limits.

Lack of modern analog data on ecosystems

Unfortunately, even data of known age often are meaningless because they lack relation to some modern analog conditions. Modern analog means the composition of natural vegetation as it existed in presettlement times, that is, prior to disturbance by EuroAmerican settlers, by sampling old-growth or virgin stands of vegetation or by using early land surveys that recorded trees on a systematic grid before settlement of land (Delcourt and Delcourt 1991). Moreover, terrestrial ecological systems vary continuously across the landscape in response to local conditions such as soil, slope steepness and aspect, ground-water level, etc. This continuous variation renders interpretation of paleoecological data difficult and often quite site-specific. For example, stable carbon isotopic data from Wilson Ridge in the Great Bend Sand Prairie suggest a C4-dominated environment during the late Wisconsin glacial maximum, which is contradictory with well-known late Wisconsin mesic climatic conditions.

Quantitative paleoclimatic data

Although there is considerable data on paleoenvironment from the central Great Plains, very little can be interpreted quantitatively, that is, in terms of degrees of temperature or inches of precipitation. Quantitative paleoclimatology seems to be conceptually quite difficult, with the few attempts thus far based on tree rings (e.g., Weakley 1962), pollen (e.g., Webb et al. 1993), phytolith morphology (Johnson et al. 1993), and fractionation of stable carbon isotope (e.g., Nordt et al. 1994; Fredlund 1993; Johnson et al. 1993). However, mathematical techniques known as "transfer functions" hold the promise of allowing quantitative reconstruction of climate by relating quantitative data on modern environments in which taxa now live to paleoenvironmental data. The use of such transfer functions requires systematic modern baseline data, both for abundance of the taxa, and for modern climate, and they therefore require data beyond the area of impact of almost any conceivable project.

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