VARIATION IN OPAL PHYTOLITH ASSEMBLAGES AS AN INDICATOR OF LATE-QUATERNARY ENVIRONMENTAL CHANGE ON FORT RILEY, KANSAS

by

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Introduction

Environmental Change

The sensitivity of grassland community composition and its boundaries to short-term climatic variation during the historical period is well documented for the central Great Plains (Tomanek and Hulett, 1970; Küchler, 1972). Similarly, long-term prairie expansion and contraction, in response to climatic variation, is documented for the prehistoric time scale (e.g., Watts and Wright, 1966; Grüger, 1973; Bradbury, 1980; Webb et al., 1983). Despite the many published studies little is known, however, about the environmental conditions prevailing in the central Great Plains during the late Pleistocene and Holocene. Because the region has few bogs and natural lakes, there is a paucity of good sites for preservation of fossil pollen and botanical macrofossils. Therefore, climatic reconstruction in this region has traditionally relied heavily on the synthesis of other proxy sources such as vertebrate fauna, snails, and occasional botanical macrofossils. As a result, interpretations of late-Pleistocene vegetation of the region range, for example, from continuous taiga-like forest (e.g., Wells and Stewart, 1987) to grassland or steppe (e.g., Graham, 1987)

The quasi-continuously deposited loess of the central Great Plains represents some of the thickest and most complete loess deposits in North America and provides a largely untapped potential for reconstructing past climates. Ongoing research by this investigator and colleagues on magnetic records from these deposits indicates a tremendous potential for environmental reconstruction. For example, lower magnetic susceptibility of the loess associates with times of rapid accumulation, i.e., the cooler glacial intervals, whereas higher susceptibility associates with the intercalated paleosols and weathering zones, which represent times of landscape stability and/or reduced accumulation rates. In addition to correlation with long-term, regional cyclic climatic changes, the relatively rapid accumulation rates associated with loess deposits also permit the resolution of short-term changes in climate (e.g., Zhou et al., 1994; Heller et al., 1993; An et al., 1991, 1993).

Recent research using nonmagnetic parameters has also been particularly rewarding. Two approaches that are now being employed include stable isotope ratio analysis (SIRA), especially that of carbon, and biogenic opal analysis, namely that of phytoliths; these techniques have been very effective in yielding unique environmental information. The coincidental use of these three proxies of past environments provides a more comprehensive picture of past environments than use of only one parameter. Magnetic information, e.g., susceptibility, indicates times of weathering and soil development in the stratigraphic sequence; SIRA of carbon provides an overall impression of the type of climate to which the plants have adapted; and opal phytolith morphology permits identification of specific groups of plants.

Fort Riley

Geomorphological and geoarchaeological research on Fort Riley, Kansas for the U.S. Army Construction Engineering Research Laboratory (CERL) is being conducted to assist in the development of a dynamic paleoenvironmental model of late-Pleistocene and Holocene landscape evolution. Research began with an overview study by D.L. Johnson (1992). On-site work by D.L. Johnson, an assistant and the contractor in June, 1993 involved further reconnaissance, subsurface exploration with a vehicular-mounted coring rig, and documentation and sampling of natural exposures. Eighteen localities, numbered 21-38, were studied, with 16 yielding sediment and soil cores. These sites represented various landscape positions within Fort Riley. Of the sites which were cored, all except 21, 26, and 33 extended to bedrock or to the residual soil developed on the bedrock. Laboratory analyses were conducted on samples from cores 21, 24, and 25 by D.L. Johnson (1994) and subcontractors. Analyses included particle size determination (pipet, hydrometer, sieve, and elutriation), various wet chemical analyses (pH, cation exchange, organic matter, phosphorous, elemental ppm, base saturation), and ^{14}C dating. Additional study has described the distribution of the loess mantle on the upland and refined the stratigraphy at the Sumner Hill locality, i.e., core site 21 (D.L. Johnson, 1996).

The first phase of the paleoenvironmental research was a test of the potential for SIRA (carbon) and opal phytolith analysis to provide a proxy time series of climate for the reservation (W.C. Johnson et al., 1994). Study was conducted on a core extracted from the Sumner Hill locality. Isotopic analysis of the carbon ($\delta^{13}C$) provided a time series that compares well with a composite of regional carbon isotope data, whereas the opal phytolith record was largely uninterpretable below about 3.5m depth, i.e., for the lower 8.5m of the core. On the basis of a recently derived chronostratigraphy of the site (D.L. Johnson, 1996; this study), the viable portion of the phytolith record does, however, include the time interval related to cultural occupation, i.e., the last 13,000 years or so.

A second phase of environmental reconstruction was designed to consider the record from several sites distributed over the reservation in upland and lowland environments. The research design was constructed as to make coincident use of magnetic analysis, SIRA, and analysis of biogenic opal (opal phytoliths). This report presents the context, results and discussion of the opal phytolith analyses conducted.

Regional Late-Quaternary Loess Stratigraphy

An articulation of the regional loess stratigraphy is necessary for the appreciation of the late-Quaternary record preserved at Fort Riley. Considerable research has been conducted on these deposits in recent years, with an emphasis on development of the chronostratigraphy. Research results from Fort Riley are, however, making a major contribution to the data base and to the understanding of paleoenvironmental conditions for the central Great Plains.

A very good empirical relationship exists between cold stages in the marine oxygen isotope record and documented times of glaciation in the United States (Richmond and Fullerton, 1986b). Given the close correspondence between the glacial record and marine isotope record and the fragmentary nature of the former, it follows that the nearly continuous loessal record should be an excellent terrestrial cognate of the marine isotope record. Therefore, the climatic and chronological record assembled for the marine sequence should match the loessal record well. The generally accepted model relating climate to the loessal record indicates that periods of stability and pedogenesis are usually associated with warm interglacials, and periods of significant loess accumulation coincide with the colder glacial times (Kukla 1977, 1987). Morphologic and isotopic analyses of plant opal phytoliths from loess exposed at the Eustis ash pit in southwestern Nebraska support the model and the relationship with the marine isotope record for the Illinoian, Sangamon and early middle Wisconsin stages (Fredlund et al., 1985; Fredlund, 1993).

Pre-Illinoian Stages

Little is known of the pre-Illinoian loesses because far fewer exposures exist than of the Loveland (Illinoian) loess and certainly the Peoria (late-Wisconsinan) loess. Pedogenesis has been recognized in these early loesses, however. Zones of carbonate enrichment, occurring at about 410-360, 330-290, and 250-200 ka, within the Barton County sanitary landfill exposure were interpreted to be pedogenic in origin (Feng et al., 1994). These carbonate zones are likely analogous to the soils observed in the pre-Loveland loesses at the Eustis ash pit, Nebraska (Fredlund et al., 1985) and elsewhere in the region (Schultz and Martin, 1970; Frye and Leonard, 1951). The zones are temporally equivalent to the nonglacial, or warm marine isotope stages (Feng et al., 1994).

Fossil pollen evidence indicates that the grasslands of the Great Plains expanded during the interglacials and contracted, or perhaps disappeared, during the glacial periods (Kapp, 1965, 1970; Fredlund and Jaumann, 1987). This suggests that the carbonate enrichment was a product of grassland pedogenesis, not unlike that of today. Fredlund and others (1985) extracted the grass opal phytoliths contained within these multiple soil zones at the Eustis ash pit in southcentral Nebraska and found that the soil-forming periods were warmer and periods of dust accumulation cooler.

Fort Riley. No pre-Illinoian Quaternary sediments have yet been recognized on Fort Riley as a result of this series of CERL-funded studies or are reported in the literature. It appears that major entrenchment and denudation preceded the Illinoian because exposures and cores to date have exposed only what is presumed to be Illinoian-age material overlying bedrock. For example, the

Illinoian Sangamon soil, loess, and alluvium rest upon a strath, or bedrock-defended terrace at the Sumner Hill locality; conversely, middle Pleistocene loess rests on an erosion surface developed on limestone at the Bala Cemetery site.

Illinoian Stage

Loveland loess. The Loveland loess is the most widespread pre-Wisconsinan loess in the Midcontinent. Several investigators (e.g., Reed and Dreeszen, 1965; Ruhe, 1969; Willman and Frye, 1970; Ruhe and Olson, 1980) have described it throughout the Missouri, Mississippi and Ohio River basins. Further, it has been recognized south into Mississippi and Arkansas (McCraw and Autin, 1989). The Loveland has been far less studied (e.g., absolute chronology, geometry, mineralogical composition) than the Wisconsinan loesses, namely the Peoria.

The Loveland may be described as a yellowish-brown or reddish-brown eolian silt. Red hues increase toward the top of the formation due to development of the Sangamon soil within the uppermost Loveland. The thickest accumulations occur in the northcentral part of the state: recorded thicknesses approach 15 m. A thinning in the loess occurs both southward and westward such that the distribution becomes discontinuous to the southwest. In Kansas, the Loveland is typically less than 10 m thick, but produces a very distinctive mark on the landscape via its variation in stratigraphy. It occurs on uplands and valley side slopes. As a result, the Loveland and its capping Sangamon soil are well expressed in exposures, particularly freshly cultivated fields.

The absolute age of Loveland loess in Kansas is largely uncertain, but recent work at sections exposed in a Geary County quarry in northeastern Kansas, the Barton and Pratt County sanitary landfills of central Kansas, and the Eustis ash pit of southwestern Nebraska provided the first absolute-age information on the Loveland beyond that carried out at the paratype section. Oviatt and others (1988) reported TL ages of 136 ka and 130 ka for the upper part of the presumed Loveland loess exposed in an abandoned quarry near the town of Milford, immediately west of Fort Riley. TL age data from this and others sites in Kansas indicates that the Loveland loess began accumulating sometime before 130 ka (Feng et al., 1994; Johnson and Muhs, 1996). Recently, Maat and Johnson (1996) derived a TL age of approximately 160 ka immediately below the Sangamon soil in Loveland loess at the Eustis ash pit. Dating at the Loveland paratype section was the first attempt to employ the TL technique on loess in the Midcontinent, and results were consistent with the data obtained from loess at other localities in the North American Continent, including those in central Kansas. Four TL ages derived from the Loveland loess indicate the sediment was deposited approximately 140,000 ka. (Forman, 1990b).

Sangamon soil. This paleosol is strongly developed and occurs throughout the Midcontinent beneath deposits of the Wisconsinan glaciation and within deposits of the Illinoian glaciation or older deposits. The Sangamon soil has been recognized in Indiana (Hall, 1973; Ruhe et al., 1974; Ruhe and Olson, 1980), Illinois (Bushue et al., 1974; Follmer, 1979) where the type section is located (Follmer, 1978), Iowa (Simonson, 1941; Ruhe, 1956, 1969), Nebraska (Schultz and Stout, 1945; Thorpe et al., 1951) and Kansas (Frye and Leonard, 1952). In Kansas, the Sangamon soil is well expressed, occurring throughout the state. Although the soil has received considerable attention in northeastern Kansas (Frye and Leonard, 1949, 1952; Tien, 1968; Caspall, 1970; Bayne et al., 1971; Schaetzl, 1986), it has been recognized at many localities in the state (Bayne and O'Connor, 1968) and recently studied in central Kansas (Feng et al., 1994). Historically, it has been referred to as a "soil in the Sanborn formation" (Hibbard et al., 1944), the Loveland soil (Frye and Fent, 1947), and the Sangamon soil (Frye and Leonard, 1951). The color of the soil ranges from a vivid to pale reddish-brown, with a loss in color occurring westward. Regionally, the soil character varies according to parent material, local drainage and climate which prevailed at the time of pedogenesis. The soil occasionally contains sufficient clay to create a subtle bench on cultivated slopes. Schaetzl (1986) noted that the soil appears to have been a very strongly developed Ultisol or Mollisol.

The Sangamon soil was first used in a time-stratigraphic context to differentiate deposits of the Illionian and Wisconsinan glacial stages (Leverett, 1899). An appreciable time span for regional landscape stability and soil formation are suggested by oxidation, apparent deep leaching, and high clay accumulation. A major problem associated with the Sangamon soil is its diachronous upper and lower boundaries (Follmer, 1978, 1982, 1983). To further confuse the time element, the lower 1-2 m (3.3-6.6 ft) of the early Wisconsinan loess is typically weathered and forms a pedological continuum with the underlying Sangamon soil (Follmer, 1983), and early investigators mistakenly included the former in the Sangamon profile. The Sangamon should be considered a *pedocomplex* rather than a single soil which developed under a unique environmental condition (Schultz and Tanner, 1957; Fredlund et al., 1985; Morrison, 1987). It apparently represents two or more paleosols welded together to form a complex that reflects significant spatial and temporal variation in environmental conditions and an appreciable time span. Although laboratory data from exposures in central Kansas indicate the Sangamon soil was strongly weathered chemically, presumably under a warm, moist climate (Feng, 1991), recent data from the Eustis ash pit in south-central Nebraska indicate that the Sangamon soil is not very mature geochemically (Muhs and Johnson, 1996).

Because of apparent time transgressiveness, the age of the Sangamon soil is not precisely known. Follmer (1983) reported a radiocarbon age of 41,700$\pm$1100 yr B.P. on plant material from the top of the Sangamon in its type area in Illinois. Forman (1990a) reported TL ages of 140$\pm$20 and 70$\pm$10 ka from loess below the Sangamon soil at two separate sites in Iowa and Illinois, and concluded the Sangamon soil is diachronous and may consist of multiple soils. Feng (1991) and Feng and others (1994) reported a TL age of about 70 ka in the lowermost part of the Sangamon soil exposed in central Kansas and associate it with marine isotope stage 3. Although Richmond and

Fullerton (1986b) assigned Sangamon time to 132-122 ka (isotope substage 5e), they acknowledged reported ages (relative and absolute) ranging from early Illinoian to middle Wisconsin. Basal ages on the overlying Gilman Canyon Formation from numerous locations in Kansas and Nebraska (Johnson, 1993a, b) provide a minimum age of about 45 ka for the Sangamon soil. Also, Forman (1990b) and Forman and others (1992) obtained TL and radiocarbon ages of 35-30 ka within the loess overlying the Sangamon soil at the Loveland paratype section in Iowa and the Pleasant Grove School section in Illinois.

Post-Sangamon time was one of extensive landscape instability including upland erosion, as evidenced by the partial or complete removal of the Sangamon soil. In a quarry near Woodruff in Phillips County, Kansas, the Loveland and Sangamon have been removed and the top of the Ogallala eroded. Similarly, the same units were apparently stripped and channels cut into the Smoky Hill Chalk prior to deposition of the Gilman Canyon Formation. Consequently, erosional truncation of the soil may be in part responsible for the apparent diachronous character of the soil. Deposition occurred at some locations: for example, a sandy zone overlying the Sangamon soil in the Phillips County sanitary landfill suggests a dry and windy transition to the Gilman Canyon Formation above. A similar unit has been observed by the author at the Eustis ash pit.

Although relatively little is known of the climate prevailing during Sangamon time, some recent research provides a first indication. In a regional examination of the stratigraphy representing Sangamon time, Dremanis (1992) noted that for the midwestern and eastern United States (the southeastern margin of the Laurentide ice sheet of North America) temperatures of the Sangamon climatic optimum were several degrees warmer and precipitation was less than at present. This time of maximum temperature and minimum precipitation, likely occurring during oxygen isotope stage 5e (c. 120ka), was followed by gradual cooling characterized by oscillations between cool and cold climates. Through global circulation model experiments, Harrison and others (1995) simulated regional climate and associated biome changes in North America for the two extremes in orbital parameters during isotope stage 5e. Maximum solar insolation, at about 125-126ka, produced midcontinental summer temperatures about 8°C warmer than today and a corresponding expansion of the warm season grasses. Conversely, minimum solar insolation, at about 115ka, depressed the summer temperatures 5°C below those of today, and expanded the cool season grasses and forests. Pedologic data obtained by Muhs and others (1996) support the notion of a climatic regime warmer and/or drier than at present.

Fort Riley. Thus far, investigations on the reservation have indicated that Illinoian age sediments are ubiquitous but relatively thin, and expressed primarily as the Sangamon soil developed within them. Best known expression of the Sangamon soil occurred at the Sumner Hill locality in a thin layer of Loveland loess overlying alluvium.

Wisconsinan Stage

Stratigraphy associated with the Wisconsinan glacial period in the central Great Plains consists of two loess units and associated soils: Gilman Canyon Formation and Peoria loess.

Gilman Canyon Formation. The Gilman Canyon Formation, first recognized in Nebraska (Reed and Dreeszen, 1965), is a middle to early-late Wisconsinan (cf. Farmdalian) loess. Equivalents of the formation have been recognized elsewhere: the Loveland loess is buried by the Roxana silt from Minnesota and Wisconsin to Arkansas and by the Pisgah Formation in western Iowa (Bettis, 1990). The Gilman Canyon of Nebraska and Kansas is typically dark in color, silty, leached of calcium carbonate, and heavily enriched in organic carbon via pedogenesis (melanization). As noted above, the formation was once considered to be the attenuated A horizon of the Sangamon soil (Thorpe et al., 1951; Reed and Dreeszen, 1965).

Reed and Dreeszen (1965) provide limited textural data and description of the Gilman Canyon Formation at the type section. Their description within the columnar section at the Buzzard's Roost exposures states (p. 62): "Upper 12 inches [31 cm] is medium dark gray, slightly humic, silt; middle 1 foot 1 inch [33 cm] is light brownish-gray silt; basal 3 feet 8 inches [1.12 m] is dark brownish-gray, humic, soil-like silt; entire thickness is noncalcareous . . . 5 feet 9 inches [1.75 m]." Although all of these attributes described at the type section appear representative of the formation as observed in Nebraska and Kansas, the bimodal distribution of humus is curious: this suggests the existence of two periods of relative stability, or low accumulation rates, and an intervening period of accelerated accumulation rates. Consequently, the Gilman Canyon Formation often appears as one or more cumulic A horizons that are developed within a variably to noncalcareous loess, usually no more than 1.2 m thick. In a section revealing an expanded valley phase of the Gilman Canyon Formation, May and Souders (1988) recognized three distinct organic zones, each of which may represent a separate episode of pedogenesis. Two such zones have been recently observed by the author at the Eustis ash pit in south-central Nebraska. If two or more distinct periods of soil formation did indeed occur regionally, they are obscured at many localities, likely due to bioturbation. Overall, the formation reflects a sufficiently slow rate of loess fall (<.08mm/yr) such that pedogenesis was operating more or less continuously, but with a decreased intensity at one or more times.

As expressed, the Gilman Canyon Formation is frequently overlain by .9-1.5 m of leached loess which is considered to be basal Peoria Formation. Correlative with the Gilman Canyon and overlying leached loess zone is the *Citellus* zone (a ground squirrel now recognized as the genus *Spermophilus*) of Nebraska (Condra and Reed, 1950). The leached zone is transitional between the well developed A horizon(s) in the Gilman Canyon and the calcareous Peoria loess above, and probably reflects a sufficiently slow accumulation of Peoria loess such that pedogenesis could keep pace only partially. A.B Leonard (1951, 1952) supported the contention that the leached, or basal zone was slowly accumulating, early Peoria loess experiencing pedogenesis through inference that gastropods were originally present, but subsequently destroyed during weathering of the loess. Above the leached zone, the rate of accumulation of Peoria loess was sufficiently rapid (c. 0.6mm/yr) as to preclude any soil development.

Radiocarbon ages from the Gilman Canyon Formation range from approximately 40 ka at the base to 20 ka at the top (May and Souders, 1988; Johnson et al., 1993a). The basal age of 40 ka agrees well with the time set by Richmond and Fullerton (1986a) for the beginning of the late Wisconsin. While Nebraska has many dated locations forming an arcuate pattern around the eastern and southern sides of the Sand Hills, data from Kansas are relatively limited. The ages in Kansas do show, however, good agreement from the south-central to the north-central part of the state and with those from Nebraska.

Given the radiocarbon time control and stratigraphic information currently available for the Gilman Canyon Formation within Kansas and Nebraska, it is clear that the associated soil(s) is a *geosol*, i.e. a laterally traceable, mappable, pedostratigraphic unit with a consistent time-stratigraphic position (Morrison, 1965; North American Commission on Stratigraphic Nomenclature, 1983, p. 865). The entire formation may be considered a geosol, but, because of the possibility for the existence of two or more identifiable cumulic A horizons merged or welded together, it may ultimately be considered a *composite geosol*.

Limited paleoenvironmental data are emerging for the Gilman Canyon Formation. $\delta^{13}C$ values are a potential source of proxy data for vegetation type and hence climate (Krishnamurthy et al., 1982). When determinations are derived from the organic fractions of the soil, they reflect inputs by the plants, particularly grasses, growing on those surfaces. C_3 (cool-season) species have an average $\delta^{13}C$ composition of -27 ‰ and C_4 (warm-season/arid) species -13 ‰, relative to the PDB standard (Deines, 1980). Terrestrial plant ecology of the Gilman Canyon Formation appears to have been characterized by primarily C_4-type grasses, or a relatively warm, possibly dry climate. From plant opal phytolith morphology (Fredlund et al., 1985; Fredlund and Jaumann, 1987) and isotope data (Fredlund, 1993), it is evident that there existed a panicoid-dominated grassland, i.e., one of moist, temperate-adapted tall grasses. These data are not inconsistent with the $\delta^{13}C$ values, since panicoid grasses are C_4 types. Further, some of the C_3-level values derived, specifically those from La Sena and Lime Creek sites, are reflecting former peaty or otherwise local, wet valley bottom environments (D.W. May, pers. comm.) which are characterized by C_3 plants meso- or hygrophytic in habit.

Interpretation of a fossil pollen assemblage from a core extracted from Cheyenne Bottoms, a large marsh in central Kansas, indicates mesic conditions in the marsh and an upland vegetation of grass and sage with scattered trees in the valley and along escarpments during the period from approximately 30 to 25 ka (Fredlund, 1991). The Farmdalian-Woodfordian transition, approximately 25-24 ka, was characterized by increased aridity. The Muscotah Marsh fossil pollen record of northeastern Kansas reflects a mosaic of deciduous forest and prairie for the late Pleistocene (Grüger, 1973; Fredlund and Jaumann, 1987). Regionally, the Farmdalian grasslands were apparently found as far east as Iowa (Baker and Waln, 1985) and north to the Sand Hills region of Nebraska (Fredlund and Jaumann, 1987).

Peoria loess. Leverett (1899) first proposed the name Peoria for an interglacial period between the Iowan and Wisconsin glacial stages. When Alden and Leighton (1917) demonstrated the Peoria was younger than Iowan, usage shifted to that of a loess, rather than a weathering interval. Within the Midcontinent, several names have been used for post-Farmdalian loess. Ruhe (1983) prefers to use the term "late Wisconsin loess" because of the uncertainties in stratigraphic equivalency from one region to another. The Peoria Formation is typically an eolian, calcareous, massive, light yellowish-brown silt that typically overlies the Loveland Formation or an approximate equivalent of the Gilman Canyon Formation.

Ruhe (1983) notes 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 latter feature is unresolved and results in correlation problems. Ruhe (1969) realized a decrease in the age of the soil under the loess from 24,500 years B.P. near the Missouri River to about 19,000 years B.P. eastward across southwestern Iowa. A decrease from 25,000 to 21,000 years 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 years B.P. in Illinois (McKay, 1979b) to 14,000 years B.P. in central Iowa (Ruhe, 1969).

In Kansas, the Peoria is a reddish, yellowish, or tan buff color, homogeneous, massive, locally fossiliferous, variably calcareous, and ranges from coarse silt and very fine sand to medium to fine silt and clay (Frye and Leonard, 1952). Thicknesses vary from in excess of 30 m adjacent to the Missouri River valley to 0.6 m in discontinuous patches. Any accumulation less than 0.6 m is presumed unrecognizable in the field because it has become incorporated into the existing surface soil. Peoria loess typically rests conformably upon the Gilman Canyon Formation.

Despite the amount of attention given Peoria loess in Kansas, the source of the silt is not completely certain. Upon a review of the available data, Welch and Hale (1987) conclude 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. Research on trace element concentrations in loess (Johnson and Muhs, 1996) indicates, however, that the Platte River valley was the primary source, with secondary inputs from the major river valleys to the south (e.g., Republican, Smoky Hill, Solomon, Arkansas).

Although readily visible stratigraphic breaks such as the Jules soil recognized in Illinois (Frye and Willman, 1973; Frye et al., 1974; Ruhe, 1976; McKay, 1979a, b) and the soil zones in Iowa (Daniels et al., 1960; Ruhe et al., 1971) have not yet been identified in Kansas and adjacent Nebraska, evidence of one or more stable or vegetated surfaces is common. The only indication 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, 1991). The most common line of evidence for a discontinuity(ies) in Peoria loess deposition is that of plant remains, usually outcropping as lenses. Many of the age determinations were made from *Picea* remains, indicating a cool, moist environment. Although radiocarbon data document the burial of vegetative material throughout the Woodfordian, two temporal clusters or modes of ages appear from the limited data: one 18-17 ka and another 14-13 ka. The former time interval represents the last glacial maximum and the latter the time of major deglaciation (Ruddiman, 1987). Interpreting ice core data from Greenland, Paterson and Hammer (1987) record a dramatic decrease in atmospheric dust content from about 13,000; this period of reduced atmospheric dust may relate to the time of relative surface stability and tree establishment. Regional geomorphic data also support the existence of a hiatus at this time. May (1989), identifies deposition of the Todd Valley Formation in the South Loup River of central Nebraska at about 14 ka, which is subsequently buried by loess. Further, Martin (1990) identifies entrenchment in the Republican River of south-central Nebraska at about 13 ka, after which valleys were filled with late Peoria Loess.

Fort Riley. The Gilman Canyon Formation sediments and soil were present at all sites sampled for magnetic and isotopic analyses, indicating that the formation has both upland (loess) and valley (alluvium) phases preserved at Fort Riley, as elsewhere in the region. Radiocarbon ages on the formation range from about 19ka to over 23ka (table 1); these relatively young ages represent the most recent period of Gilman Canyon time pedogenesis; the relatively unaltered loess and alluvium was not sampled for dating.

Coring and backhoe trenching indicates that Peoria loess is typically relatively thin or undetectable. Only one of the magnetic and isotopic study localities exhibited clearly identifiable unaltered Peoria loess below the Brady soil.

Holocene Series

The beginning of the Holocene, about 10 ka (Hopkins, 1975), was a time of dramatic environmental change and attendant stratigraphic discontinuities. This boundary is generally considered only geochronometric, i.e., without specific stratigraphic reference, although a stratotype in Sweden has been proposed for the boundary (Mörner, 1976) and has a reported age of 10,000±250 years B.P. (Fairbridge, 1983). Watson and Wright (1980) contended that major climatic and environmental change at 10 ka may be documented only on a local scale, i.e., all changes recorded in the stratigraphic record are diachronous. This notion now seems to be faulty on the regional and subcontinental scale in that research of the last decade has documented major pedogenesis at 10 ka in both alluvial and eolian/upland settings. This is the first major geosol to occur in the stratigraphic record of the region since the Gilman Canyon soil 10,000 years earlier.

Brady soil. The Brady soil was first named and described by Schultz and Stout (1948) at the Bignell Hill type locality, an eolian 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, 1949, 1951). It is regionally extensive only in the northwestern and west central parts of Kansas, and 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 or may have developed in Peoria loess subsequent to the erosion of the Brady soil and Bignell loess. 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.

The age of the Brady soil has been uncertain, even at the type locality. Dreeszen (1970, p.19) reported an age of 9160±250 (W-234) obtained in 1954 and another in 1965 of 9750±300 (W-1676), both from the type section but very likely contaminated by modern plant roots. Subsequently, Lutenegger (1985) reported an age of 8080±180 years B.P. but provided no specifics other than that the source was the A horizon of the Brady soil at the type section. Better age control for the type section has since been secured by this investigator: ages of 9,240±110 (Tx-7425) and 10,670±130 (Tx-7358) years B.P. were obtained on the upper and lower 5 cm, respectively, of the Brady A horizon.

The Brady soil has been recently dated at localities in Nebraska and Kansas. Souders and Kuzila (1990) obtained a radiocarbon age of 10,130±140 years B.P. on the Brady soil occurring within the Republican River valley of south-central Nebraska. Sites along Harlan County Lake upstream from Naponee have yielded a number of ages, ranging from 10,550±160 to 9,020±95 years B.P., on exposures of the Brady soil (Cornwell, 1987; Johnson, 1989; Martin, 1990; Martin and Johnson, 1995). Two radiocarbon ages of 9820±110 (TX-7045) and 10,550±150 (TX-7046) years B.P. have been derived from the upper and lower 5 cm, respectively, of the Brady A horizon exposed in Barton County, central Kansas (Feng, 1991).

Although it appears Brady pedogenesis occurred from about 10,500 to as recently as 8,500 years B.P., greater refinement of the Brady soil chronology is necessary, but present data clearly indicate it was a product of a major period of landscape stability at a time when widespread climatic shifts were occurring at the end of the Wisconsin. This was the first significant period of soil development since Gilman Canyon time, and represents the climate of the early Holocene. There is an isochronous alluvial soil found throughout the region which is particularly well expressed within the Kansas River basin (Johnson and Martin, 1987; Johnson and Logan, 1990). The two ages of 8,274±500 (C-108a) and 9,880±670 (C-471) years 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 on the Brady soil. The soil, occurring in both eolian and alluvial contexts, appears to qualify, based upon present radiocarbon data, as a *geosol*, like the Gilman Canyon Formation soil. A typical exposure of the Brady soil would be that in Phillips County, located in west-facing roadcuts in the SW4, SW4, Sec.24, T4S, R19W. The locality was recognized by A.R. Leonard (1952, p.42-3) and revisited by Johnson (1993b). It is the east face of a road cut about .8 km north of Speed, Kansas in which the Peoria loess, Brady soil, and Bignell Loess are visible. In the late 1940s and early 1950s the Loveland loess, Sangamon soil, and Gilman Canyon Formation were also exposed in the roadcut; they can yet be distinguished in a poor quality exposure around on the north face at the end of the roadcut. A profile within the road cut was excavated and sampled for radiocarbon dating in the uppermost and lowermost 5 cm of the A horizon: ages of 8,850±140 (Tx-6626) and 10,050±160 (Tx-6627) years B.P. were obtained, respectively (Johnson, 1993b). In sum, age data indicate a soil forming interval lasting 1,500-2,000 years.

Development of the Brady soil correlates well with indicators of regional climatic change. The fossil pollen record at Muscotah Marsh of northeastern Kansas indicates that spruce had essentially disappeared from the region by about 10,500 years B.P. As this decline occurred, deciduous tree species increased until about 9,000 years B.P., the time at which grassland expansion began (Grüger, 1973). On a hemispheric scale, the abrupt decrease in atmospheric dust noted in the Greenland ice core at 10,750 years B.P. (Paterson and Hammer, 1987) reflects decreased loess deposition and possibly Brady-age pedogenesis associated with relative terrestrial stability. Further, ^{18}O levels within the same core suggest rapid warming about 10,750 years B.P., with the characteristic Holocene temperature regime being established about 9,000 years B.P.

Bignell loess. The Bignell loess was first described and named at the type locality in a bluff exposure on the south side of the Platte River valley southeast of North Platte, Nebraska (Schultz and Stout, 1945). It is typically a gray or yellow-tan, massive silt, calcareous and seldom more than 1.5 m (5 ft) thick. Although it is often somewhat less compact and more friable than the underlying Peoria loess, no certain identification can be made without the presence of the Brady soil. The Bignell loess does not form a continuous mantle on the Peoria; instead, it occurs as discontinuous deposits which are most prevalent and thickest adjacent to modern-day valleys, particularly the south side, and often within depressions on the Peoria surface. Feng (1991) speculates that the Bignell loess of central Kansas is relatively well weathered because it was derived from a preweathered source, the Brady soil surface, perhaps eolian and alluvial phases alike. This is consistent with the earlier interpretation derived in Nebraska that Bignell loess is at least partially comprised of re-worked Peoria loess (Condra et al., 1947, p. 33).

It appears from the radiocarbon ages obtained at the type section in Nebraska and the Speed roadcut in northwestern Kansas that the Bignell loess can be no older than 8,000 to 9,000 years B.P. Snails collected by A.B. Leonard from the lower part of the Bignell in Doniphan County, northeastern Kansas, produced ages of 12,500±400 (W-231) and 12,700±300 (W-233) years B.P. (Frye and Leonard, 1965). Because the shell material had absorbed an indeterminate amount of dead carbonate, Frye and others (1968) proposed an averaged age of approximately 11,000 years. Based upon the age data available for the Brady, the soil humate-derived ages are probably closest to reality.

A pronounced feature of the Holocene climate of the Great Plains was an extended warm, dry period (Wright, 1970; Benedict and Olson, 1978; Barry, 1983), identified as the Altithermal (Antevs, 1955) or, less commonly, as the Hypsithermal (Deevey and Flint, 1957). This dictates that the Bignell loess was a warm-climate loess, unlike the cold-climate loess of the Woodfordian. Reconstruction of the general circulation patterns for North America indicates that from the last glacial maximum about 18 to 15 ka there was no detectable change in atmospheric circulation: the westerly jet was split by the Laurentide ice sheet into a north and south flow around a strong glacial anticyclone (Kutzbach, 1985, 1987; COHMAP Members, 1988). By 9 ka, the ice had wasted appreciably, the jet was no longer split, orbital parameters were favoring increased temperatures, and zonal flow was dominating (Kutzbach, 1981,1985, 1987). Model results produced mean summer temperatures 2° to 4° C higher (COHMAP Members, 1988) and annual precipitation up to 25 % less than at present in the region (Bartlein et al., 1984; Kutzbach, 1987).

Because of the increasing zonal flow and aridity of the Altithermal, species of the tall grass community migrated eastward to the present areas of mixed deciduous-prairie vegetation, i.e., the prairie-forest ecotone shifted eastward (Van Zant, 1979; Semken, 1983; Webb et al., 1983). The fossil pollen record from Muscotah Marsh provides a disrupted but interpretable Holocene signal, indicating a middle Holocene prairie expansion (Grüger, 1973). Fossil pollen data from Cheyenne Bottoms suggest consistently lower water levels in the marsh during the middle Holocene (Fredlund, 1991, 1995). Molluscan fauna from the Bignell Loess of Kansas suggest that climate was somewhat drier than during Peoria time (Frye and Leonard, 1951). After a period of soil formation near the end of the Pleistocene, pedogenesis is not recognized until about 5,800 years B.P. in the sand sheet of the Great Bend Prairie, central Kansas (Johnson, 1991). Therefore, based upon various climatic proxies and a limited number of radiocarbon ages, it appears the Bignell loess was deposited, for the most part, from the end of Brady pedogenesis at about 8,500 years B.P. to about 5,500 years B.P.

Fort Riley. The Brady soil and Bignell loess are prominent elements of the late-Quaternary stratigraphy on the reservation. Together with the surface soil, they typically comprise the approximately upper 2m of most sites documented (D.L. Johnson, 1996), including many of the sites analyzed for this report. Therefore, it appears that appreciable loess was deposited during the Holocene. This is not surprising given the proximal juncture of two large stream systems, which formed a significant loess source (dust from a relatively wide valley floor), and the situation of the sites on the north side of the river valley with southerly winds.

Paleoclimatic History during the Wisconsinan Glacial Stade

As the most recent glacial episode, the Wisconsin has the greatest chronostratigraphic resolution. However, existing knowledge of the climatic environment for this time interval is relatively limited and inconsistent for the central Great Plains. To date, either pollen records from peripheral areas or synthesis of various types of proxy data have been used to reconstruct climate history of the region.

The insolation record exhibits two relatively warm peaks (c. 50 and 30 ka) during marine isotope stage 3. Each peak is followed by a relatively minor and gradual decrease, culminating in the decrease to the glacial maximum (c. 18ka). In contrast to the gradual insolation changes, most paleoclimatic records from this interval indicate rapid alternations of warm and cold events, which in frequency and timing appear to be unrelated to the Milankovitch forcing (Curry et al., 1992). Dansgaard and others (1985) showed that rapid and extreme fluctuations in stable isotopes (signifying air-temperature differences of 5°C) from two sites in Greenland appear to correlate over the past 50,000 years . These fluctuations are in phase with changes in CO_2 and dust content. Rapid and substantial temperature fluctuations recorded in ice core segments during stage 3 also correspond with oscillations in the North Atlantic marine sediment record of species abundance and ice rafting (Heinrich, 1988). Using accelerator ^{14}C ages on a planktonic polar species, Broecker and others (1988) identified four rapid climatic oscillation between 40 and 22 ka in the North Atlantic.

The regional upland vegetation, as inferred from the pollen record from Cheyenne Bottom in Kansas (Fredlund, 1995), appears to have been nearly treeless throughout the Farmdalian (ca. 30-24ka). The pollen record suggests that, although regional tree and shrub populations were higher and more diverse than in Holocene, they were a secondary component of the overall vegetative structure. Grassland-sage steppe dominated the regional uplands surrounding the Cheyenne Bottoms basin throughout the Farmdalian period. The rise in Cheno-Am pollen percentages and an influx of sand beginning at about 25 ka probably mark the rapid onset of a cycle of aridity. Immediately after the onset of aridity, the most noticeable changes are declining percentages in the Cheno-Am types and rising percentages of *Pinus*. The increase in *Picea* and other arboreal pollen may signal a climatic shift toward cooler climatic conditions at ca. 24 ka. Unfortunately, the Woodfordian substage of the Wisconsin is missing from the Cheyenne Bottoms record.

Limiting ^{14}C dates in North America indicate that glaciation commenced about 25-27 ka and thus allow less than 10,000 years for ice buildup prior to 18 ka (Andrews, 1987). The structure of deglaciation 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 data supporting the smooth deglaciation model are maps of Laurentide ice area based on ^{14}C-dated glacial deposits (Andrews, 1987). 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 America 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 warming at 10 ka and about 13 ka; these warming 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 in centered on 13.7 and 12.3 ka. (Ruddiman, 1987).

The Younger Dryas (e.g., Osborn et al., 1995: Bard et al., 1993; Peteet et al., 1992) deglaciation model is suggested by the strong signal of 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.

By the middle Holocene, drying had reached a maximum according to most studies. Northwest Texas was experiencing conditions of maximum temperatures, minimum precipitation, and eolian activity between 6000 and 4500 yrs B.P. (Holliday et al., 1983; Holliday, 1985; 1989; Johnson, 1987; Pierce, 1987). This episode coincides with $\delta^{13}C$ values from soil organic matter from the same area revealing a shift from -23‰ in the early Holocene to -15‰ in the middle Holocene (Haas et al., 1986). These results were interpreted to represent a shift from cool-season C_3 grasses to warm-season C_4 grasses. Based on enriched $\delta^{13}C$ values in soil carbonate from northwest Texas, Humphrey and Ferring (1994) also show a middle Holocene xeric episode, although the $\delta^{18}O$ values from these same carbonates do not indicate a significant temperature change.

A noticeable shift back to cooler and/or wetter conditions was detected in many areas shortly after 5000 yr B.P. The Great Bend Sand Prairie transformed to conditions much like present (Arbogast, 1995). According to Humphrey and Ferring (1994), the return to mesic conditions after 5000 yrs B.P. was interrupted in north-central Texas by a brief warming and drying episode between 2000 and 1000 yrs B.P. Based on depositional environments, they concluded that cooler and wetter conditions returned after 1000 yrs B.P.

Study Localities

Sites were selected for opal phytolith study on the basis of spatial coverage of the reservation, landscape position, and absolute age control resulting from studies by D.L. Johnson (1994,1996). Sumner Hill (core site 21; D.L. Johnson, 1994) and Bala cemetery (core site 19; D.L. Johnson, 1994) were the upland sites investigated. The Sumner Hill site is located along the bluff line above Camp Funston and consists of approximately 12m of Quaternary sediment over a strath cut into the Permian Wreford Limestone. A basal gravelly alluvium is overlain by over 11m of loess, i.e., the Loveland loess and Sangamon soil up through Bignell loess capped by the surface soil. Due to the analytical constraints imposed by the sampling of cores, a series of six large trenches were excavated in the valley wall slope to form a stair-step exposure of the sediments in order that the entire sequence could be clearly viewed and thoroughly and accurately sampled. This site was investigated in the first phase of opal phytolith investigation (Johnson et al., 1994), but is included here to provide a comprehensive perspective. The other upland sample site is located adjacent to the Bala cemetery, a location nearly 40km northwest of the Kansas River valley and about 10km east of the Republican River valley. The site consists of approximately 2.75m of loess overlying the Permian Winfield Limestone, a much thinner loess mantle than at the former site located adjacent to the river valley.

The Pump House Canyon site consists of a loess deposit situated on a strath located within a small tributary valley to the Kansas River valley. Sampling was conducted in a 3.3m deep backhoe trench. Although the trenching did not perforate the loess layer, its total thickness is likely less than 5m because the upper part of the Gilman Canyon Formation was exposed. The Manhattan Airport site (core site 33), located approximately 1.5km northwest of the airport, consists of clay-rich late-Pleistocene and Holocene alluvium. The site was backhoe trenched to expose the upper 3.8m of fill, which included alluvial phases of the Sangamon soil (?), Gilman Canyon Formation, and Brady soil, which was overlain by Holocene alluvium.

Opal Phytolith Analysis

Due to a lack of old-growth trees suitable for dendroclimatology and bogs and natural lake for palynology and macrofossil analysis, the central Great Plains offers little in the way of climatic proxy sources. Opal phytolith analysis, however, has the potential to offer paleoenvironmental and paleoclimatological information that will provide a relatively detailed picture of the past.

The objective of this research was to determine the feasibility of recovering fossil opal phytoliths (siliceous plant cells) from sediment and soil samples collected at the study sites in order to reconstruct the vegetative history for the reservation. Phytoliths are the most common biosilicate in upland loess deposits and are the most useful for environmental reconstructions. Sponge spicules, another form of biogenic silica, may also be present in upland deposits but are much less common and of limited use when compared to the potential of phytoliths.

Grass opal phytoliths are the best studied and can be separated into morphologic categories related to the plant photosynthetic pathways and the major subfamilies of grasses. Twiss and his students (Twiss, 1980, 1983, 1987; Twiss et al., 1969; Kurmann, 1981, 1985) were the first to recognize the correlation between grass photosynthetic groups (adaptations) and phytolith morphology, i.e., the major subfamilies of grasses correspond to three morphologic classes of phytoliths.

Most Poaceae (grasses) employ the C_3 pathway (Calvin) for the fixation of CO_2 in the photosynthetic process. Commonly, these grasses belong to the Pooideae (festicoid) subfamily. Grasses included in this subfamily include the bromes (*Bromus* spp.), fescues (*Festuca* spp.), needlegrass (*Stipa* spp.), wheatgrass (*Agropyron* spp.), bluegrasses (*Poa* spp.), and many cereals such as rye (*Secale cereale*), oats (*Avena* spp.), barley (*Hordeum vulgare*), and wheat (*Thriticum aestivum*). The C_3 grasses are widespread but are best adapted to the higher (cooler) latitudes and altitudes.

Conversely, the C_4 grasses, employing the Hatch-Slack CO_2 photosynthetic pathway, are most successful in the lower (warmer) latitudes and altitudes. This system is better adapted to high temperatures and low moisture conditions. In the Great Plains, two groups (subfamilies) of grasses typically utilize the C_4 pathway, the Chloridoideae and Panicoideae subfamilies. Examples of grasses within the Chloridoideae subfamily are the three-awns (*Aristida* spp.), gramas (*Bouteloua* spp.), buffalo grass (*Buchloe dactyloides*), saltgrasses (*Distichlis* spp.), sandreed grass (*Calomovilfa* spp.), lovegrasses (*Eragrostis* spp.), muhly grasses (*Muhlenbergia* spp.), and dropseed grasses (*Sporabolus* spp.). Although the Panicoids are well adapted to high temperatures, they require more moisture than the Chloridoids, and are consequently better adapted to the eastern Great Plains, e.g., eastern Kansas. Included in the Panicoids are the bluestems (*Andropogon* spp.), panicums (*Panicum* spp.), indian grass (*Sorgastrum nutans*), gama grass

(*Tripsacum dactyloides*), and the cereal grasses corn (*Zea mays*) and sorghum (*Sorghum halepense*).

Loess deposits of the central Great Plains have been found to contain large amounts of grass opal phytoliths, which produce an interpretable climatic record (e.g., Fredlund et al., 1985; Bozarth, 1991b, 1992b; Fredlund, 1993; Johnson, 1993a; Johnson et al., 1993b). Further, opal phytolith data from the Gilman Canyon Formation at the Eustis ash pit indicate the dominance of C_4 grasses throughout most of the formation, and appear to correlate nicely with the $\delta^{13}C$ values. Opal phytoliths, unlike $\delta^{13}C$ values, can indicate the presence of an arboreal component (e.g., Rovner, 1971; Geis, 1973; Wilding et al., 1977) and differentiate to some extent between deciduous and coniferous trees (Bozarth, 1992a, 1993c).

Opal phytoliths are generally well preserved in most sediment and can be isolated from sediment samples and analyzed to reconstruct the paleoenvironment for a particular area. This has been successful on a number of sediment types, including loessal sites in China (Lu et al., 1991), Nebraska (e.g., Fredlund et al., 1985; Bozarth, 1991b, 1992b; Johnson et al., 1993a; Fredlund, 1993) and the Southern High Plains (Bozarth, 1995), as well as alluvium in Kansas (Kurmann, 1981, 1985; Bozarth, 1986) and the Southern High Plains (Bozarth, 1995), and swamp and upland sediment in Panama (Piperno, 1988).

Formation and Stability

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 characteristic shapes are called opal phytoliths. Phytolith is derived from the Greek words *phyton*, meaning plant, and *lithos*, meaning stone. Opal is the common name for hydrated silicon dioxide. Opaline bodies formed in plants without specific shapes are simply plant opal. Phytoliths form in most plants and are produced in many shapes and sizes. Many phytolith types are specific to particular groups of plants. Phytoliths are largely a "decay in place" fossil (Rovner, 1975) and represent the vegetation of a site at the time of deposition (Piperno, 1988).

The dissolution and stability of phytoliths in soil is not fully understood. Laboratory experiments demonstrate, however, that the solubility of silica is a function of temperature, particle size, pH, and the presence of a disrupted surface layer. Studies show that the solubility of amorphous silica increases linearly with temperature from 0°C. Particle size is another factor affecting stability as opal dissolution is greater with a decrease in size (Wilding et al., 1977, 1979). Pease (1967) experimentally determined that there appears to be a slight increase in phytolith solubility in the range of 5.0 to 8.5, an added increase between pH 8.5 and pH 9.0, and a large increase beginning at pH 9.0.

Opal stability is also a function of the presence of certain metallic ions and sesquioxides. The adsorption of Al and Fe ions onto the surface of opal will decrease silica dissolution due to the formation of relatively insoluble silicate coatings. The presence of sesquioxides may increase dissolution of phytoliths due to the adsorption of monosilicic acid (Wilding et al., 1977).

Morphology and Taxonomy

Monocotyledons, particularly the Poaceae, produce a wide variety of morphologically distinctive opal phytolith forms. The most taxonomically useful types of grass phytoliths are silicified short cells. Several types of trapezoidal circular, rectangular, and elliptical short cells are diagnostic of the Pooideae (Brown, 1984; Twiss, 1987; Bozarth, 1992b), a C_3 grass subfamily adapted to cool temperatures (Twiss, 1987). Saddle-shaped bodies occur most commonly in the Chloridoideae, a C_4 grass subfamily (Brown, 1984; Twiss, 1987; Mulholland and Rapp, 1992) that flourishes in areas with warm temperatures and low available soil moisture. Saddle-shaped phytoliths are similar in appearance to double-edged battle axes formed by two opposite convex edges and two opposite concave edges. However, a few saddle-shaped phytoliths have only one concave side (Brown, 1984).

Bilobate and cross-shaped phytoliths are formed in the Panicoideae, the C_4 grass subfamily (Brown, 1984; Twiss, 1987; Mulholland and Rapp, 1992) that thrives in warm temperatures and high available soil moisture (Twiss, 1987). Bilobates with indented, concave, or pointed lobes are formed only in grasses in the Panicoid subfamily. Bilobates with raised lobe edges and round or flat ends which are symmetrical in side view are also formed only in Panicoids (Bozarth, 1992b).

Bilobate phytoliths with raised lobe edges and round ends are also formed in *Aristida* (needlegrass, wiregrass), a genus in the Chloridoid subfamily (Gould and Shaw, 1968). However, bilobates formed in *Aristida* differ from Panicoid bilobates in that the raised edges on the top (the longer part) slope down at the ends. In addition, they are asymmetrical in side view as the top is more concave than the bottom (Bozarth, 1992b). *Stipa*, a genus in the Pooid subfamily (Gould and Shaw, 1968), also produces bilobates (Bozarth, 1992b). These bilobates differ from those produced in Panicoids and *Aristida* by not having raised lobe edges. Many have a small lobe on one side in the middle. Unlike most Pooids, *Stipa* species grow in dry areas (Pohl, 1968).

There are several other types of phytoliths produced in grass in addition to short cells. Long cells are relatively large, elongate bodies with smooth or wavy edges. Bulliform cells are large keystone shaped-cells. Dendriforms are cylindrical rods of varying length that have protrusions or spines radiating from a central core. Asteriforms are roughly spherical spiky phytoliths. Trichomes are silicified prickly-hairs composed of two parts, an outer sheath and an inner core. The outer sheath dissolves soon after being deposited on the soil, while the inner core remains well preserved. The silicified stomata are taxonomically useful at various levels but are typically not well preserved. The other types are not specific to any particular subfamily but are preserved in most sediment. Piperno (1988) reported that dendriforms and asteriforms are apparently formed only in grass floral bracts.

Non-grass monocots also produce numerous taxonomically valuable phytoliths. *Cyperus* (sedge) produce distinctive phytoliths in the form of cone shaped-bodies with round wavy margins. These phytoliths are occur both singly and in multiples. Truncated cones with multiple peaks and round wavy bases are formed in *Scripus pallidus* (bulrush). Both of these phytolith types appear to be diagnostic of the genera that produce them (Bozarth, 1993).

Several types of phytoliths are produced in woody dicotyledons (deciduous shrubs and trees) and herbaceous dicotyledons (forbs and weeds). The two most common types of diagnostic dicot phytoliths are flat polyhedrons with 5-8 sides and anticlinal cells (Rovner, 1971; Wilding and Drees, 1971; Geis, 1973; Wilding et al., 1977; Bozarth, 1992a). Anticlinal cells have wavy, undulating walls with the appearance of jigsaw-puzzle pieces. Most of these polyhedral and anticlinal phytoliths consist only of silicified cell walls and are not well preserved in sediment (Wilding and Drees, 1974; Bozarth, 1992a). Other phytolith types formed only in dicots include branched elements with spiral thickening and honeycomb-shaped assemblages (Geis, 1973; Wilding and Drees, 1973, 1974; Bozarth, 1992a).

Several species of arboreal dicots produce opal spheres that range in size from 1 to 50 micrometers (Wilding and Drees, 1973, 1974). Opal spheres are also produced in conifers (Klein and Geis, 1978), but are much smaller (3 to 8 micrometers). Opaque opal spheres have been extracted from the A horizon of several forested soils in Ohio demonstrating that they are well preserved (Wilding and Drees, 1973, 1974).

Spiny spheres are formed in neotropical palms (Piperno, 1988) but have not been reported in temperate vegetation. However, the association of spiny spheres with deciduous tree phytoliths in a loessal site in Nebraska (Bozarth, 1992b) suggests that they are also formed in this, or an associated, group of plants.

Wilding and Drees (1973) reported opaque bladed forms (which appear to be opaque platelets), in white oak (*Quercus alba*). Similar particles were observed in isolates from a soil formed under deciduous forest.

Several families and genera of dicots produce phytoliths unique to those taxa. Opaque platelets with systematic perforations and certain types of segmented hairs are diagnostic of Asteraceae (the sunflower family). Platelets with irregular edges and echinate (spiny) sculpturing on one side are formed in the fruit of hackberry (*Celtis occidentalis*) and appear to be unique to that genus. These types are well preserved in sediment. Flat polyhedrons with 5-8 sides that are filled with coarse verrucae (bumps) appear to be unique to Ulmaceae (the elm family) (Bozarth, 1985, 1987c, 1992a).

Certain types of stalked verrucate phytoliths are specific to hackberry, mulberry (*Morus*), false nettle (*Boehmaria*), or nettle (*Urtica*). Elongate verrucate phytoliths with one or both ends tapering to a point are unique to *Pilea* (Bozarth, 1992a). Phytoliths with deeply scalloped surfaces of contiguous concavities are unique to *Cucurbita* (Bozarth, 1987a).

Several types of phytoliths are produced in the Pinaceae (pine family). Silicified, irregularly-shaped, polyhedral cells are the most common taxonomically useful Pinaceae phytolith. This type of phytolith is produced in *Picea rubens* (red spruce), *P. mariana* (black spruce), *P. glauca* (white spruce), *P. engelmannii* (Engleman spruce), and *Pinus banksiana* (jack pine) (Norgren, 1973; Klein and Geis, 1978; Bozarth, 1988, 1993c). Blockly polyhedra with smooth surfaces and at least eight non-parallel sides are characteristic but not diagnostic of Pinaceae, because they are also produced, although relatively infrequently, in grasses (Bozarth, 1993c).

In contrast to smooth polyhedrons, polyhedrons with bordered pit impressions on the surface are unique to the Pinaceae. This type of phytolith is abundant in *Pinus* (pine), *Picea* (spruce), Douglas-fir (*Pseudotsuga*), and less commonly in *Larix* (larch), *Tsuga* (hemlock), and *Abies* (fir) (Klein and Geis, 1978). *Pseudotsuga menziesii* (Douglas-fir) needles produce distinctive, branched, silicified particles (Brydon et al., 1963). This same type of phytolith was also reported in Douglas-fir by Garber (1966) as irregular shapes with spiny processes and by Norgren (1973) as amoeboid bodies with tapering, conical protrusions. Thin plates with wavy margins on all four sides are formed in needles of *Picea glauca* (white spruce) and appear to be unique to that species. Phytoliths with spiny irregular bodies are commonly formed in needles of *Pinus banksiana* (jack pine) and appear to be diagnostic of that species (Bozarth, 1993c).

Methodology

Sampling was done either from a core extracted with a trailer-mounted Giddings drilling machine or from an exposure created by backhoe trenching. In both cases, samples were collected every 10cm using a trowel cleaned thoroughly between samples. Samples were placed in sterile plastic bags for storage until extraction. Phytoliths were isolated from 5-gram subsamples using a procedure based on heavy-liquid (zinc bromide) flotation and centrifugation (Bozarth, 1991a). This procedure consists of five basic steps: 1) removal of carbonates with dilute hydrochloric acid; 2) removal of colloidal organics, clays, and very fine silts by deflocculation with sodium pyrophosphate, centrifugation, and decantation through a 7-micron filter; 3) oxidation of sample to remove organics; 4) heavy-liquid flotation of phytoliths from the heavier clastic mineral fraction using zinc bromide concentrated to a specific gravity of 2.3; 5) washing and dehydration of phytoliths with butanol; and 6) dry storage in 1-dram glass vials.

A representative portion of each phytolith isolate was mounted on a microscope slide in immersion oil under a 22x40 mm cover glass and sealed with clear nail lacquer. Each isolate was then studied at 400x with a research-grade Zeiss microscope. Each sample slide was first examined to determine the quality of preservation of the phytoliths. At least 200 phytoliths, were counted in all of the samples with adequate preservation. A complete slide was scanned and all phytoliths classified in those samples with poor preservation.

Estimates of phytolith concentration were made using an indirect method reported by Piperno (1988). A known number of exotic spores (in this case *Lycopodium*) were added to each sample after the oxidation stage. The concentration of phytoliths (per gram) was computed as follows:

Phytolith conc. = no. of phytoliths counted x (total no. exotics added / no. exotics counted) / 5

Concentration permits an evaluation of the phytolith production, preservation, and sedimentation rate for a given sample interval.

Phytoliths were classified according to a convention that has been developed and used by other reports and publications. An extensive reference collection of plants native to the Great Plains has been developed in the palynology laboratory through field collection, research plots, solicited samples, and specimens supplied by the University of Kansas Herbarium. The phytolith reference collection consists of phytoliths extracted from complete or representative aerial portions of the following: 1) 25 species of 20 genera of 11 tribes of 6 subfamilies of the Poaceae (grass); 2) 11 species of 4 genera of 4 non-grass monocot families; 3) 65 species of 62 genera of 11 families of herbaceous dicots; 4) 20 species of 18 genera of 13 families of woody (mostly arboreal) dicots; 5) 14 species of 7 genera of 5 families of gymnosperms; and 6) 2 species of Equisetum. These reference materials include all the dominant species in the study area as reported by Küchler (1974).

An unknown group consists primarily of phytoliths too poorly preserved to be classified any other way. There were a few other phytoliths included under this heading that were stuck under the cover glass and could not be rotated for three-dimensional viewing, thereby precluding positive taxonomic classification.

The biosilicate data are presented in computer-generated, percentage diagrams using Tiliagraph software (Grimm, 1992) designed for depicting fossil pollen data. For each site analyzed, two diagrams were constructed, one for all of the data and another for only the short cells diagnostic of grass subfamilies and the arboreal-type phytoliths. The latter diagram presents an unobscured perspective on the more diagnostic microfossils. Biosilicate data for each site have been subdivided on the diagrams into zones. Zonation is a common tool in botanical microfossil analysis, particularly palynology (fossil pollen analysis), and is rapidly being adopted in opal phytolith studies. This approach is generally used to identify periods which are similar with regard to the assemblages of the various microfossil types. Zonation has been used in this study to differentiate periods of concentration and preservation as well.

The phytolith data are also represented in this study as an aridity index, or measure of the moisture stress. The index may be computed using two or more of the three categories of phytoliths. Diester-Haass and others (1973) presented a climatic index consisting of the ratio of chloridoid phytoliths to the total number of chloridoid and panicoid phytoliths, based on a total count of 100, i.e., the higher the index, the more arid the climate and vise versa. If the pooid phytolith data are also used, an index of the ratio of chloridoid types to the sum of the three classes can be computed as an aridity index (Twiss, 1987). Diekmeyer (1994) created a third variation on the index consisting of the ratio of the chloridoid-panicoid sum to the sum of all three classes. The latter index has been adopted for use in this study because it appears to have the best intuitive basis.

Results

Sites were selected to represent the loess-mantled upland, both proximal and distal to the Kansas River valley, loess and colluvium-mantled straths on the Kansas River valley wall, and late-Pleistocene terraces in the Kansas River valley. These landscape situations are diverse and represent different depositional and environmental regimes. All sites have been ^{14}C dated to establish the chronostratigraphy (table 1), but two of the sites have been sampled for additional age control; the samples are currently pending.

Sumner Hill

Results from the extraction and analysis were unanticipated, based upon extensive experience in the extraction and analysis of phytoliths from loess deposits elsewhere in the region. Both concentration and preservation of phytoliths were poor, except for the upper part of the core. From 0 to 40 cm phytolith concentration and preservation were good, and from 40 to 60 cm concentrations and preservation were adequate for interpretation. From 60 to 350cm, phytolith preservation and concentration were marginally adequate for analysis. Extraction procedures were repeatedly verified and samples rerun in order to check the concentrations. The effectiveness of the extraction procedure was attested to by good recovery of *Lycopodium* spike grains. Poor preservation typically associates with low concentrations. Further, phytoliths recovered frequently exhibited deep solutional pitting.

The degree of phytolith preservation correlates well with the pH stratigraphy of the core: Within the upper part of the core where preservation is good, pH values are about 5.9 to 7, but below that pH rises to a precarious level for the preservation of phytoliths (Pease, 1967). Other locations where phytolith analyses have been conducted on loess, pH, where measured, has been generally between 7 and 8. Although high pH levels may not be exclusively responsible for the poor preservation, they certainly seem to be a contributing factor given the close correlation between pH and preservation.

All isolates were studied microscopically, but, because of concentration and preservation, every sample was quantitatively analyzed in the upper 350 cm and every other sample below that level. Results are presented in two formats: relative frequencies of all phytolith types encountered (fig.

2) and relative frequencies of grass short cells alone (fig. 3). The latter presentation was produced so that the presence of the other phytolith types would not obscure the interpretation of the phytoliths diagnostic of the various grass subfamilies.

The percentage diagrams (figs. 2, 3) have been subdivided into three zones: A, B, and C. Zone C represents the bulk of the core in which phytolith data are not interpretable, i.e., Gilman Canyon and earlier material. The break between Zones B and C coincides with the sedimentary, pedogenic, and isotopic break at about 358 cm. With such small short cell sums, the percentage bars are inflated and largely meaningless. Zone B, extending from the late Pleistocene to the late Holocene, exhibits a chloridoid peak centered on about 130cm, or about 5ka. At about 275-350 cm (14-13ka), an increase in C_4 panicoid types occurs, showing a change in the grass composition. Arboreal types appear within the zone, particularly about 11ka and earlier; coniferous types (e.g., *Picea* and *Pinus*) are likely at least part of this arboreal vegetation based upon macro- and other microbotanical information from the late Pleistocene record of the region. Zone A represents the upper part of the core which is dominated by the surface soil development and late Holocene sedimentation. C_4-type grasses dominate: panicoid and chloridoid subfamilies are high relative to the C_3 pooids.

Bala Cemetery

Biosilicate recovery from the Bala Cemetery site was very good to within the lowermost 40cm of the core, and the percentage of unknown phytoliths is relatively small (figs. 4, 5). Zone F (>240cm depth) is uninterpretable and consists of the lower Gilman Canyon Formation and limestone residuum. Zone E, 240-170cm, is distinguished by the relatively high frequencies of spiney spheres, the most common arboreal-type phytolith and indicative of a moist environment. Moreover, the predominance of pooid phytoliths suggests a cool, moist climate during this time. In Zone D, pooid types are at their highest percentage for the entire core, probably a reflection of the glacial maximum climate. There is, however, little evidence of trees during this time. The arboreal component appears to expand within Zone C, while the pooids decrease upward and chloridoids increase, i.e., a shift toward warmer, drier conditions. Zone B features a maxima in pooids, indicating decreased temperatures and a likely increase in available moisture. The uppermost zone (A), coinciding with the modern surface soil, exhibits a decrease in pooids and increase in the chloridoids and panicoids, i.e., warmer temperatures. The change from Zone B to A may be a function of either climate shift or historical cropping (e.g., sorgum). Age data for the site are yet pending and will provide the necessary time control for the interpretation and correlation with other sites.

Pump House Canyon

Phytolith preservation was extremely poor in samples collected at the Pump House Canyon site, except in the uppermost 40cm (figs. 6, 7). Despite poor preservation, four zones are definable. Zone D (>90cm) is uninterpretable due to the low concentrations of biosilicates, a likely result of corrosion due to elevated pH values. Zone C (70-90cm) is defined by high frequencies of spiney spheres, an arboreal type. Similar ratios of spiney spheres and pooid short cells were recognized at the same depths (75 and 85cm) in the Bala Cemetery core, indicating a similar type of vegetation at both sites. Zone B (40-70cm) is characterized by a decrease in chloridoid types and increase in pooids, with a diversity of arboreal types. The change in frequencies of the three types of grass phytoliths parallels that for the Bala Cemetery site. Pooids dominate Zone A, although chloridoids are well represented; panicoid types play a minor role throughout.

Manhattan Airport

On the basis of samples analyzed from the Manhattan Airport backhoe trench, the diagrams (figs. 8, 9) have been subdivided into three zones. The lowermost zone, C, has been differentiated on the basis of poor preservation and low concentration of phytoliths. Consequently, variations in phytolith values before about 13ka are meaningless and uninterpretable. Zone B, or the latest Pleistocene and Holocene, exhibits good phytolith preservation. The pooids are dominant, but decline in the upper part of the zone, i.e., as the Holocene progressed; chloridoids increased, but panicoids showed little change. Significant numbers of arboreal-type phytoliths, particularly spiney spheres, are present throughout this zone. Zone A, coincident with the surface soil and encompassing about the last 2,000 years, indicates a maximum in chloridoid grasses and a decrease in the arboreal component from Zone B.

Discussion

Landscape Evolution

A vision of how the upland landscape of Fort Riley evolved during the last 120,000 years or so is now beginning to emerge. Such a perspective provide a useful context for viewing environmental reconstructions. Evolution of the uplands is, of course, not a process isolated from the valley environment; the alluvial setting is extremely sensitive to environmental change and intimately linked to the upland change. Reconstructions of the alluvial history is presently underway and will likely provide data of greater temporal resolution than that of the uplands, at least for the Holocene.

During pre-Illinioan time, the ancestral Kansas River was cutting a strath in the limestones of the present valley walls, tens of meters above its present elevation. This bedrock surface and it covering of alluvial sands and gravels were subsequently mantled with Loveland loess, at a time when the river had entrenched to near its present level. Loess fall was relatively limited, which indicates that the source was either the alluvial plains to the northwest or the Missouri River valley to the east; either source is sufficiently removed spatially or meteorologically such that little loess fall would occur in this distal location. At the end of Illinoian time, the Sangamon soil developed in this thin mantle of loess. The alluvial phase of the Sangamon soil observed at the Manhattan airport site indicates the river was certainly no higher than this high terrace during the last interglacial.

From the ubiquity of the Sangamon soil in the study area as well as throughout the region, it is apparent that erosion was limited following Sangamon time, i.e., during the early Wisconsinan glacial stage. The Sangamon soil appeared to have survived and continued to dominate the landscape until the middle Wisconsin. At this time another loess unit was being deposited, i.e., the Gilman Canyon loess. The source of this loess is unknown, but it appears to different than that of the Loveland loess because it maintains a uniform thickness throughout the central Great Plains, rather than exhibiting a distance decay function as with the Loveland and Peoria loesses. Gilman Canyon loess fall was, however, exceedingly slow: the meter or so of loess accumulated over approximately 20,000 years, i.e., from before 40ka to about 20ka. Other than the relatively thin nature of the loess unit, the occurrence of relatively low depositional rates (surface stability) is indicated by the presence of two or more periods of soil development. The pedogenesis occurred primarily during the latter part of Gilman Canyon time (Farmdalian substage?) in that soil development typically dominates the upper half of the formation.

The contact between the Gilman Canyon Formation and the overlying Peoria loess is transitional. It appears that Gilman Canyon pedogenesis waned as the late-Wisconsinan loess began to fall such that accumulation rates eventually exceeded the rate of pedogenesis. Peoria loess deposition was, like the Loveland loess, relatively low in the study area. The loess unit thickens to the west and northwest in proximity to the sources, e.g., the Sand Hills of Nebraska, Platte River valley, and to a lesser extent adjacent to the other large river valleys crossing the central Great Plains, e.g., the Republican and Kansas. Accordingly, Peoria loess is thickest in the northwestern part of the reservation; over 2m appear to be present at the Bala Cemetery site.

Although no direct evidence has yet been found, some of the Peoria loess, as with other loess units, was probably lost, at least locally, during erosional maturation of the late-Wisconsinan landscape. Clearly, sites to the northwest such as Bala Cemetery have realized erosion of the Peoria loess and overlying Brady soil, assuming it developed in the area. At some localities, such as Sumner Hill, the total thickness of Peoria loess present at the end of the Pleistocene was completely involved in Brady soil development.

The next major pedogenic period (after the Gilman Canyon) occurred during the Pleistocene/ Holocene transition. The Brady soil, developed in the upper Peoria loess as deposition ceased, is, on the basis of excavations and coring, well represented in the southern part of the reservation. Consequently, the upland surface was largely stabilized during this time of environmental shift from the Wisconsinan glacial stage of the Pleistocene into the Holocene. Deposition of the Bignell loess, beginning during the early Holocene, buried the Brady soil. The southerly wind flow off the Kansas River valley apparently transported silts and fine sands from the alluvial bottom up onto the uplands. The rapid distance decay factor is suggested by the lack of Bignell loess at the Bala Cemetery site to the northwest of the Sumner Hill and Pump House Canyon sites. The contrast may be due to differential erosion, but this is deemed unlikely given that the two sites to the south adjacent to the Kansas River valley are in equally or more erosion-prone settings. Since the Brady soil is presumed to have developed to the northwest but is now apparently gone (e.g., Bala Cemetery), a thin layer of the Bignell loess was probably present but quickly eroded; now the Peoria loess is now exposed with its associated surface soil.

Throughout the reservation, surface soil development appears relatively recent, probably within the last 2,000 to 1,000 years and even less locally. This relative surface youth indicates that either erosion has occurred recently or that Bignell loess has been episodically or slowly accumulating on the landscape until relatively recently. The latter scenario appears correct, at least for areas along the southern margin of the reservation adjacent to the Kansas River valley, in that the Holocene upland surface has been periodically stable (slow or zero loess accumulation) such that soil formation could occur. These short periods of stability are manifest as weak or incipient soils developed in the Bignell loess, such as those indicated at the Pump House Canyon site. The preservation of cultural sequences, ranging from Clovis to Woodland and Protohistoric, beginning in the Brady soil and extending up through the Bignell loess into the surface soil at archaeological sites in these setting within the region (e.g., Johnson, 1996a) also indicates that there has been little erosion of the Bignell loess, at least locally.

Given the preservation and distribution of late-Quaternary deposits on the uplands, Fort Riley is an ideal study area for modeling landscape evolution in the context of cultural history. Further, the loess and alluvial deposits of the reservation permit ready extraction and interpretation of proxy data for the reconstruction of past environments.

Paleoenvironmental and Climatic Interpretation

The Regional Phytolith Record. Opal phytolith analysis of prehistoric sediments was first done in the Great Plains about ten years ago, but only a handfull of studies have been reported; only 17 studies have been done on a total of 15 sites in Nebraska, Kansas, Oklahoma and Texas. Of these studies, nine were seeking archaeological information (e.g., cultigen use), and seven for environmental reconstruction.

Of the five sites investigated in Nebraska, the Eustis ash pit is the most notable. Mining of volcanic ash up into the 1950s resulted in a 35m-high vertical face exposing loess and intercalated soils. Analysis of 32 samples collected from throughout the section demonstrated that phytoliths are relatively concentrated within the loess at that locality and that they are abundant in both the buried soils and in the intervening loess units (Fredlund et al., 1985). Results indicated that the soils formed during periods that were relatively warmer and drier than the periods of major loess deposition, including the modern surface soil. Two subsequent studies on samples from the site further articulated the utility of phytolith analysis in environmental reconstruction. An analysis by Bozarth of samples collected at close interval from the Gilman Canyon Formation and adjacent lower Peoria loess indicated a cool, moist grassland during early Gilman Canyon time which shifted to a warmer, drier grassland as soil development occurred and back to cool, more moist conditions again when the Peoria loess began to accumulate (Johnson et al., 1993a). A very good correlation exits between the $\delta^{13}C$ data and the aridity index computed from the phytolith data for the Gilman Canyon Formation; six magnetic parameters reflect the unique climatic episode as well through the climatic controls on weathering (fig. 10). In a novel approach, Fredlund (1993) extracted O, H, and C isotopic data from individual phytoliths to test the feasibility doing such to reconstruct past environments; in this pilot study, each of the samples analyzed produced results approximating the climatic conditions of the respective period.

The Sargent site, also in southwestern Nebraska, consists of a 6m-high vertical exposure within a gully cut into the loess-mantled upland along the south wall of the Platte River valley. Exposed is late Pleistocene Peoria loess overlain by Holocene Bignell loess containing at least five buried soils. Fossil biosilicates (phytoliths, sponge spicules, statospheres) were well preserved in all 34 samples analyzed (fig. 11; Bozarth, 1992b). The phytolith record indicates that a deciduous forest or open woodland was growing at the site about 10,640 years B.P. By about 10,080 years B.P., the site had been transformed into a pooid-dominated grassland, reflecting a relatively cool and moist growing season climate. After 10,000 years B.P., there was a relatively rapid increase in chloridoid grasses, reflecting increasing temperatures and reduced available moisture, i.e., appearance of the Holocene climatic regime. Conditions continued to become drier and warmer until about 3,800 years B.P., when a slight increase in available moisture occurred as indicated by the rise in panicoid phytoliths. The present dominance of chloridoid-type phytoliths reflects the situation of the site in the ecotone separating grama grass-buffalo grass from needle grass-wheat grass. A plot of the aridity index (after Diekmeyer, 1994) against $\delta^{13}C$ data from the site reveals a close correspondence between the two data sets and defines the increasing aridity throughout the Holocene (fig. 12).

The La Sena site is a Paleoindian archaeological site buried in a loess-mantled terrace in southwestern Nebraska and was exposed by wave action on Harry Truman Reservoir (May, 1991; May and Holen, 1993). The cultural deposit, consisting of a bed of fractured mammoth bones, appears to have been a butchering site at about 18,000 years B.P. Well-preserved phytoliths were isolated from four sediment samples. The dominance of pooid phytoliths in the mammoth bone bed, compared to a modern analog, indicated that the climate was much cooler and more moist than at present. Moreover, the presence of non-grass phytoliths suggested that deciduous trees were growing on or near the site at that time (Bozarth, 1991b).

Two disparate areas in Nebraska were investigated for the presence of cultigen phytoliths. Numerous well-preserved phytoliths were isolated from samples collected at three archaeological sites on Harlan County Lake in south-central Nebraska. No phytoliths derived from cultigens were found, but those grass phytoliths detected indicated an environment similar to that of today (Bozarth, 1987d). At archaeological site 23DX3, a Central Plains Tradition village discovered in northeastern Nebraska, Bozarth (1984b) identified silicified sunflower-like, multi-celled hair bases in phytolith isolates from sediment samples collected in cultural features.

Eight sites have been studied in Kansas, four environmental reconstructions and four archaeological investigations. Kurmann (1981, 1985) extracted phytoliths, as well as pollen and fungal spores from both a modern soil and a buried soil in alluvial fill of the Flint Hills region of east-central Kansas with the intent of obtaining a correlation with local vegetation. Pollen could distinguish woodland from prairie but was unable to differentiate short-grass from tall-grass prairie, but the phytolith study did so. In another environmental study, Bozarth (1986) collected nine samples between 20 and 213cm depth from within a cutbank of the Saline River on the upper reaches of Wilson Lake, central Kansas in order to characterize the environment at the time of alluvial deposition. Well-preserved phytoliths were isolated from all samples, including a modern analog. A comparison of the fossil phytoliths with the modern assemblage indicated that a period of alluviation after 5090 years B.P. and before 1940 years B.P. was drier than today, whereas the period of soil formation about 1940 years B.P. was more moist than at present. A comparison of the relative frequencies of chloridoid, panicoid, and pooid types indicates that the vegetation at the Wilson Lake area has been a mixed-grass prairie since that time.

In an attempt to reconstruct the diet of prehistoric bison of the central Great Plains, phytoliths were extracted from bison tooth calculus and impacta collected from a specimen found in Finney County, southwestern Kansas (Bozarth, 1993b). The phytolith assemblages derived from the calculus and impacta sample isolates indicated that the animal ate mostly short grasses (chloridoids) such as buffalo grass and blue grama (*Bouteloua gracilis*), but the presence of panicoid and pooid types provided evidence that the animal also grazed to some extent on the C_4 tall grasses and the C_3 cool-season grasses. The Jetmore mammoth site (14HO1), located in western Kansas, consisted of a buried mammoth bone deposit that dated to 9,235 years B.P. (Bozarth, 1984b). Analysis of phytoliths isolated from four subsurface samples and one modern (surface) analog sample indicated that most of the grasses growing concurrently with the mammoth(s) were panicoids, suggesting that the area was more humid than at present. Phytolith data also indicated that deciduous trees were growing at the locality during the early Holocene. Fossil pollen analysis conducted on the same samples suggested an open woodland environment.

Phytoliths were isolated from four sediment samples collected in two archaeological sites at Perry Lake, northeastern Kansas with the goal of documenting the aboriginal use of cultigens (Bozarth, 1990a). No economic species were identified from the phytolith record, but pooid, chloridoid, and panicoid grasses were common at both sites. Phytoliths from dicots, likely deciduous trees, were also present in small concentrations. Site 14MN328, representing a Great Bend Aspect village in central Kansas, was occupied by the late prehistoric Wichita, a group known for extensive cultivation some 500 years ago. Bozarth (1989, 1990c, 1993a) isolated phytoliths from samples collected in seven cultural features, with the intent of searching for the presence of cultigens. Three bean (*Phaseolus*) phytoliths were recovered from a sample representing a trash-filled storage pit, and a maize cob phytolith was identified from a hearth sample. The Hatcher site (14DO19) consists of the remains of a prehistoric habitation structure (Plains Village period) located adjacent to Clinton Lake in northeastern Kansas. Analysis of phytoliths isolated from archaeological and modern analog samples indicated that the site was covered with tall-grass prairie at the time of occupation. An associated examination of phytoliths contained within a daub concentration demonstrated that panicoid and pooid grasses were the most common grasses used in construction at the site (Bozarth, 1987b).

Another archaeological site studied, the Stigenwalt site (14LT351) in southeastern Kansas, revealed what has become known as the Stigenwalt complex, an early Archaic manifestation that occurred over a fourteen-hundred year period beginning about 8,810 years B.P. Analysis of well-preserved phytoliths from the site suggested that the primary vegetation community of the area at the time of occupation was one of mixed-grass prairie similar to that found to the west in central Kansas today (Bozarth, 1990b). This is a relatively dry type of grassland, significantly different from the more mesic mix of tall-grass prairie and cross-timbers forest common to the region at the present.

A single archaeological site in Oklahoma has undergone phytolith analysis: the Coody Creek site (34MS31) is a late Caddo village located in the eastern part of the state. Phytoliths were isolated from nine samples collected within four features as part of a paleoenthnobotanical study (Bozarth, 1984b). All samples yielded numerous well-preserved phytoliths, including squash-like groups of silicified epidermal cells

In a study of late-Quaternary alluvial fills in the Texas Panhandle, samples from eight sites were processed for their biosilicate contents, including phytoliths, siliceous bodies, and sponge spicules (Bozarth, 1995). Preservation was relatively good in samples from five of the sites: relatively high frequencies of chloridoid type phytoliths indicated that short-grass prairie was the dominant vegetation at Mustang Springs from about 10,000 years B.P. to the present, at the Edmonson site about 8,000 years B.P., at the Lubbock, Texas landfill about 3,100 years B.P., at the Sundown site about 2,100 year B.P., and at the Flagg site about 770 years B.P. Collectively, these findings suggest that short-grass prairie covered most of the Southern High Plains throughout the Holocene.

To date, phytoliths from very few sites have been analyzed, and most of those that have provide only a limited vignette of the past. Only the Sargent site in southwestern Nebraska provides a relatively detailed view of the Holocene; it is, however, much farther north than Fort Riley and therefore did not likely experience the extremes in moisture stress during the late Pleistocene and Holocene found to the south into Kansas. Fort Riley offers an opal phytolith record that complements the SIRA data and magnetic records (Johnson et al., 1994; Johnson, 1996b).

The Fort Riley Phytolith Record. Whereas the resulting diagrams depicting the phytolith data were characterized above ("Results"), the following provides an interpretation in the context of the preceding discussion of past environments for the central Great Plains and the Fort Riley area.

The Sumner Hill site provides a viable phytolith record of the latest Pleistocene and Holocene, although variance is relatively large, due in part to poor preservation and low concentrations (fig. 2, 3). During the latest Pleistocene, about 14 to 11ka, all three grass subfamilies seem to have been represented in a mixed grassland environment on this bluff-top landscape position. At least two samples from this interval indicate a strong panicoid grass representation, i.e., warm, but moist grassland climate. After about 11ka, the panicoid grasses stayed about the same, but the warm, dry-adapted C_4 grasses increased in percentage and the C_3 grasses slowly declined, an expression of the progressively increasing temperatures as the Holocene grassland climate was being defined. The aridity index (fig. 13) exhibits a fluctuating curve due to the variation in the percentages of the phytolith types. Not all of this is likely due to low phytolith counts; some is certainly a function of climatic variability during the Holocene. The overall pattern is, however, a decrease in the cool-season C_3 grasses and an increase in the warm-season C_4 grasses, particularly the chloridoid species, from the late Pleistocene to the present. Magnetic data display little correlation with the aridity index, an indication that the detail of the phytolith record is largely due to low concentration and preservation (fig. 13).

The Bala Cemetery site provides a quality phytolith reconstruction of the grassland environment on the broad uplands in the northeastern part of the reservation during the late Pleistocene (fig. 4, 5). C_3 grasses are well represented throughout the sedimentary record, reflecting cool growing season temperatures and relatively low moisture stress during loess deposition. Temperatures were also too low for the C_4 panicoid grass to successfully compete for a prominent role. Climate of the Last Glacial Maximum (LGM) is apparent in the pooid subfamily maximum (Zone D), which probably dates from about 18 to 16ka. Spiney sphere concentrations indicate that deciduous trees were represented prior to and subsequent to the LGM; during the LGM coniferous trees may have replaced the deciduous varieties, but are not represented or recognized in the phytolith record. The drop in C_3 grasses occurring at the top of Zone 3 is marked by a slight increase in the C_4 chloridoid grasses; this may be an indication of a short-term perturbation in climate. The upper 50cm of the profile indicates a slight warming as both types of C_4 grasses increase. The rise in the panicoid C_4 species indicates that moisture levels are staying relatively high, i.e., the main change is an increase in temperature. The decline in available moisture is pronounced in the Holocene, which is unrepresented here. The aridity index (fig. 14) depicts the gradual increase in moisture stress as the Pleistocene ends and the Holocene approaches. The brief drying at the top of Zone C is represented by the jump in the curve at about 70cm depth. The frequency dependence curve approximates the aridity curve, with the magnetic increase at about 1.5m possibly relating to soil development during the LGM (fig. 14).

The phytolith record at the Pump House Canyon site is abbreviated, encompassing only the last 2-3,000 years, based on the single Holocene age of 5,810 years B.P. (fig. 6, 7). For a brief period, possibly several hundred years, the strath was occupied by C_3 grasses and deciduous trees. A probable interpretation of the relatively high pooid grass population and trees is that the strath environment has a microclimate slightly less severe than the exposed upland sites above. Shortly thereafter the C_3 grasses declined and the xeric C_4 grasses dominated. This was a short-lived period: xeric C_4 grasses decreased slightly and then varied little to the present; and C_3 grasses expanded their role, with little subsequent change. The curve of the aridity index (fig. 15), restricted to the upper 90cm, illustrates two dry peaks, which may correlate with soil formation identified in the frequency dependence curve.

Phytolith stratigraphy from the Manhattan Airport site provides insight to conditions on the flood plain and terraces of the late Pleistocene and Holocene. The viable phytolith record begins about 13ka. Cool-season grasses dominated the alluvial surface at this site from the Pleistocene-Holocene transition until the middle Holocene; these C_3 grasses may, however, be edaphic, i.e., adapted to wet flood plain conditions, and not represent the broad scale climatic patterns. Deciduous arboreal species were also a major component during this time. Sometime prior to 2,080 years B.P. the C_3 grasses declined and both xeric and mesic C_4 grasses increased; the arboreal component also declined dramatically. Without higher-resolution age control, the timing of this change can only be estimated, but ages of 5,000 or 3,500 years B.P. are reasonable guesses given the peak of alluvial activity evident in the stratigraphic record at this time (Johnson and Martin, 1987; Johnson and Logan, 1990; Mandel, 1995; Johnson et al., 1996). This latter change may again be at least in part a response to changing local conditions, e.g., lowering watertable with channel entrenchment, rather than to only regional climate shift. Nonetheless, the change is nicely documented and would impact aboriginal activity on this particular alluvial surface. The aridity curve (fig. 16), valid for only the upper 2m, illustrates the amelioration in climate following the Pleistocene-Holocene transition and corresponds well with the magnetic data curves.

Summary and Conclusions

The stratigraphy of loess deposits which have been investigated on Fort Riley exhibits the same sequence of loess units and intercalated buried soils as is found elsewhere in the region, but adds detail unique to the reservation. Composite stratigraphy of the late-Quaternary loess deposits preserved on the reservation consists of the basal Sangamon soil of the Last Interglacial (Illinoian age; c. 120-110ka), Gilman Canyon Formation (a pedogenically-altered, middle Wisconsinan loess; c. >40-20ka), Peoria loess (late-Wisconsin; c. 20-10ka), Brady soil (Pleistocene/Holocene transition; c. 11-10ka), Bignell loess (Holocene; c. 9-?ka), and modern surface soil.

The record of climatic change during the late Quaternary has been retained in the loess deposits of the central Great Plains, and such a record is extractable using some relatively new approaches. The loess sequences at Fort Riley represent a nearly continuous time series of climatically-forced environmental change for the late Quaternary. Application of two analytical techniques, magnetics and SIRA, is providing proxy data sets that represent a time series of climatically regulated pedogenesis/weathering and botanical composition (Johnson, 1996b). This investigation, however, presents the results of an expanded study of the opal phytolith record contained within sites from various landscape setting on the reservation (cf. Johnson et al., 1994).

Study localities included two upland sites, Sumner Hill and Bala Cemetery; a valley-wall site, Pump House Canyon; and a valley-bottom site, Manhattan Airport. Sumner Hill, a bluff top site on the north side of the Kansas River valley, has the thickest loess accumulation of all sites examined (c. 13m), extending from the Sangamon soil to the surface soil, whereas Bala Cemetery site, in the extreme northwestern part of the reservation, consists of less than 3m of loess above the limestone bedrock. Pump House Canyon site has a relatively thick accumulation of loess, probably in excess (Gilman Canyon and Peoria loesses) of 5m. Sediments at the Manhattan Airport are alluvial, not eolian, and exhibit the alluvial phase of two or more of the soils found in the upland record (Brady and Sangamon and/or Gilman Canyon).

Quality of the opal phytolith record varied appreciably among the sites. Sumner Hill produced only a marginal record for the upper 3m, i.e., the latest Pleistocene and all of the Holocene. After some abrupt and brief fluctuations in grass composition at the end of the Pleistocene, the xeric warm-season grasses expanded as the Holocene climate and its grassland biome evolved. At the Bala Cemetery site, the Holocene record appears to be absent, with the phytolith assemblages indicating the amelioration of climate at the closure of the Pleistocene, i.e., the mesic, cool-season grasses were gradually yielding their dominance to the more xeric warm-season grasses. The quality of record at this upland site is good. The Pump House Canyon site provided the poorest phytolith record of all sites investigated thus far; phytolith preservation and concentration were good only in the upper 30-40cm. The limited data indicate a mixed grass environment for the last 2-3,000 years on this strath. The Manhattan Airport site, a high terrace consisting of late-Pleistocene to late-Holocene alluvium, provided a good phytolith record for the upper 2m and offers an interesting mix of regional and local environmental signals. The mesic cool-season grasses slowly yield to the more xeric warm-season grasses as the Pleistocene comes to an end, but the relatively high percentage of C_3 grasses may be exaggerated in the record due to the microclimate of the alluvial bottom (high soil moisture and arboreal shading).

The composite of these data portrays climatic change at the end of the Pleistocene and much of the Holocene: C_3 grasses, such as the bromes, fescues and bluegrasses, gave up their dominance and the C_4 grasses took the advantage, particularly the more xeric chloridoid types, such as the gramas, buffalo grass, and muhly grasses. These data provided no clear indication of major, long-term events during the Holocene, in particular the Altithermal. Notable events, both long- and short-term, are very likely part of the climatic history for the region and the reservation, but the quality of the Holocene phytolith record for the upland loess deposits is limited to the marginal record obtained at Sumner Hill. As a result, possible correlation with the magnetic record is limited.

Despite the limitations of the phytolith data obtained to date, more is known about the late-Quaternary climatic history of Fort Riley, particularly as it relates to landscape evolution and to changing environments and resources of the aboriginal peoples occupying the reservation landscape since the late Pleistocene. Grassland composition and associated gathering potential and game changed, and trees came and went from the landscape. Also, the upland environment was different from the strath and from the alluvial bottoms, and rates and magnitude of change varied among the different landscape elements.

Recommendations

Based on the results of this and past studies at Fort Riley (Johnson et al., 1994; Johnson, 1996b), it is recommended that further opal phytolith analysis be conducted in order to realize the potential of the upland paleoenvironmental record and to generate a geoarchaeological model of sufficient temporal and spatial resolution to provide the basis for development and/or testing of the computer-based geoarchaeological model under development by R. Bras and colleagues at MIT. Specific recommendations include:

1) Selection and analysis of one or more upland sites containing expanded Holocene records in order to improve on the quality of record obtained from the Holocene phytolith record of the Sumner Hill site. After potential sites are identified, a spot check analysis of phytolith isolate quality would be made to locate sites with good records. The Bala Cemetery site indicates that quality *phytolith records exist in the upland late-Pleistocene (Peoria) loess deposits, but such a quality record needs to be found in Holocene-age (Bignell) loess.*

2) Samples should be collected and analyzed from within and adjacent to archaeological sites located on both the uplands and stream terraces. Phytolith analysis has shown itself to be an outstanding techniques for reconstructing paleoenvironments and cultural resources use from archaeologically enriched stratigraphy.

Opal phytolith analysis is a relatively new technique and the University of Kansas Palynology Laboratory is one of the few places in the world where it is being done. Application of the technique to the loess deposits of Fort Riley has, therefore, been a learning experience upon which one can build.

Since phytolith analysis was referenced in recommendations included in the review draft of the magnetic and isotope analyses on the reservation (Johnson, 1996b), those recommendations are restated:

1) In that the latest Pleistocene and entire Holocene are of primary interest in development of a cultural resources and landscape evolution model, further emphasis should be placed on this portion of the upland stratigraphic record. The environmental transition at the late-Pleistocene/Holocene boundary needs to be better characterized, as does the detail of the Holocene (e.g., the Altithermal, soil-forming periods). Specifically, high-resolution (e.g., 2cm contiguous) sampling should be undertaken on two or more new sites with expanded Holocene sequences and on cores from the existing sites of Sumner Hill and Pump House Canyon for the purpose of SIRA of carbon. Further, these sites should be sampled at close interval (c. 25cm) for ^{14}C dating; subdividing the samples in order to individually date the total humates and humic acid and humin fractions would be helpful in deciphering the chronostratigraphy (Martin and Johnson, 1995; Johnson and Valastro, 1994). Magnetic analyses (susceptibility and frequency dependence) should also be conducted at the new sites selected.

2) The analysis of trace element content has been used in a number of applications related to the reconstruction of late-Quaternary environmental conditions (e.g., research by Muhs et al., 1990, 1994; Reheis, 1990; and Diekmeyer, 1994). Trace elements (e.g., cerium, strontium, yttrium, zirconium) contained with in the loess and intercalated soils provide a "fingerprint" of the sediment and provide information regarding paleowind direction and loess source region. Data obtainable include not only presence or absence, but also the concentration of individual trace elements. Recent research by Johnson and Muhs (unpub. data; Johnson et al., 1993b) has documented the difference in source regions for loess deposits of eastern Colorado versus those of southern Nebraska and Kansas. For Fort Riley, two levels or orders of information can probably be derived from the loess sequences. First, trace element analysis can indicate the stratigraphic level (and time, with sufficient ^{14}C control) at which the winds shifted from the north-northwest to the south-southwest, i.e.,

determine whether the trace element signature is that of the Platte River sediments or of the Kansas River sediments. The second order variations in the data will document the more subtle changes within the trace element time series. These variations will indicate small order changes in wind intensity and direction at a given location, as well as define the distance-decay function along a sampled transect. The stronger the winds, the higher the concentration of the trace elements, since they are not only rare but also represent the heavy mineralogic fraction. Short-term changes in direction will also show up in the signature. Trace element analyses should be conducted on two or more of the sites referred to in the first recommendation, with sampling being conducted at close interval (c. 2cm). Previous research (Johnson et al., 1993b, Diekmeyer, 1994, Muhs and Johnson, 1996) has indicated that a suite of eleven trace elements contains the optimal amount of information for examining Pleistocene and Holocene loess of the central Great Plains. Facilities for preparation of samples for analysis exist within this investigator's laboratory, and those for analysis exist both in the Department of Geology, University of Kansas, and at the U.S. Geological Survey, Denver, Colorado (D.R. Muhs' laboratory).

3) Extensive archaeological surveys currently being conducted on the reservation are yielding new upland prehistoric sites, one or more of which have been found to contain evidence for Clovis (latest Pleistocene/earliest Holocene) occupation. Selected archaeological sites should be analyzed to articulate the specific paleoenvironmental record. As a corollary to the first recommendation, these sites, depending on their location relative to the nonarchaeological sites, should receive high-resolution magnetic, isotope, *opal phytolith* and perhaps trace-element analyses.

4) Results of D.L. Johnson's survey of the upland loess mantle (D.L. Johnson, 1996) should be used as base line data in developing a detailed map of surface materials in order to provide the precision of information needed to articulate a detailed sequence of landscape evolution on Fort Riley. A combination of shallow coring and ground-penetrating radar surveys can be employed to provide the information for input into a GIS in order to create a detailed map of surficial geology and the spatial variation in thickness of the various map units. Shallow coring can be conducted using the trailer-mounted Giddings soils probes operated by the Department of Geography, University of Kansas. Ground-penetrating radar, conducted by dragging a skid-mounted radar system behind an all-terrain vehicle, and personnel are available through both the Department of Geology, University of Kansas and the Kansas Geological Survey.

Research conducted to date by this investigator (Johnson et al., 1994; Johnson, 1996b; this report) and by D.L. Johnson (D.L. Johnson, 1992, 1994, 1996) has defined the late-Quaternary lithostratigraphy of the reservation and documented the first-order trends in the chrono-, magneto-, and biostratigraphy (SIRA, opal phytoliths). Now that the potential for reconstruction of the late-Quaternary environmental record has been realized and broadly defined, high-resolution study needs to be undertaken at selected existing and new sites in order to extract the interpretable detail.

Table 1. Radiocarbon (^{14}C) Ages.

Site Name (Core Number)	Depth(cm)	ISGS no.[1]	$\delta^{13}C$ (‰)	Corrected Age (yr BP)[2]
Sumner Hill (21)[3]	150-160/pit 6	3101	-16.2	6,140±130
	170-180/pit 6	3140	-17.1	7,770±220
	200-210/pit 6	3142	-16.8	11,230±310
	240-250/pit 6	3131	-16.2	11,980±270
	260-270/pit 6	3143	-15.9	13,240±310
	280-290/pit 6	3146	-15.6	13,470±430
	215-225/pit 2	3164	-15.5	19,170±650
	310-320/pit 5	3159	-16.0	20,760±750
	255-265/pit 2	3160	-15.4	21,580±740
	235-245/pit 2	3163	-15.1	21,610±590
	330-340/pit 5	3158	-17.0	22,560±990
	350-360/pit 5	3157	-17.2	23,000±1,100
Bala Cemetery (19)[4]	150-220	2622	-17.3	19,070±280
Pump House Canyon[5]	130-140	3165	-15.6	5,810±180
	290-300	3167	-21.6	18,830±450
	330-340	3166	-18.4	23,010±850
Manhattan Airport (33)	40-48	2996	-14.4	2,080±70
	163-168	2997	-20.4	11,550±440
	525-528	3004	-16.5	11,580±570
	228-233	3003	-18.3	13,740±540
(20)	168-240	2623	-19.4	19,990±450
No name (28)	23-28	3005	-15.5	1,000±100
	28-33	3006	-15.4	1,730±110
	147-152	3008	-19.0	14,140±770
	202-207	3007	-17.3	15,110±590

1 Illinois State Geological Survey Radiocarbon Laboratory number.

2 Radiocarbon age corrected for the effects of isotopic fractionation (Taylor, 1987).

3 Pits 2 and 5 overlapped stratigraphically, and the same unit (Gilman Canyon Formation) was dated in both pits, but at slightly different subunit stratigraphic positions.

4 This age was determined by D.L. Johnson (1994) on a core from the area; ages are pending from the location where phytolith and isotope samples were collected in this study.

5 A second suite of ages is pending for this site.

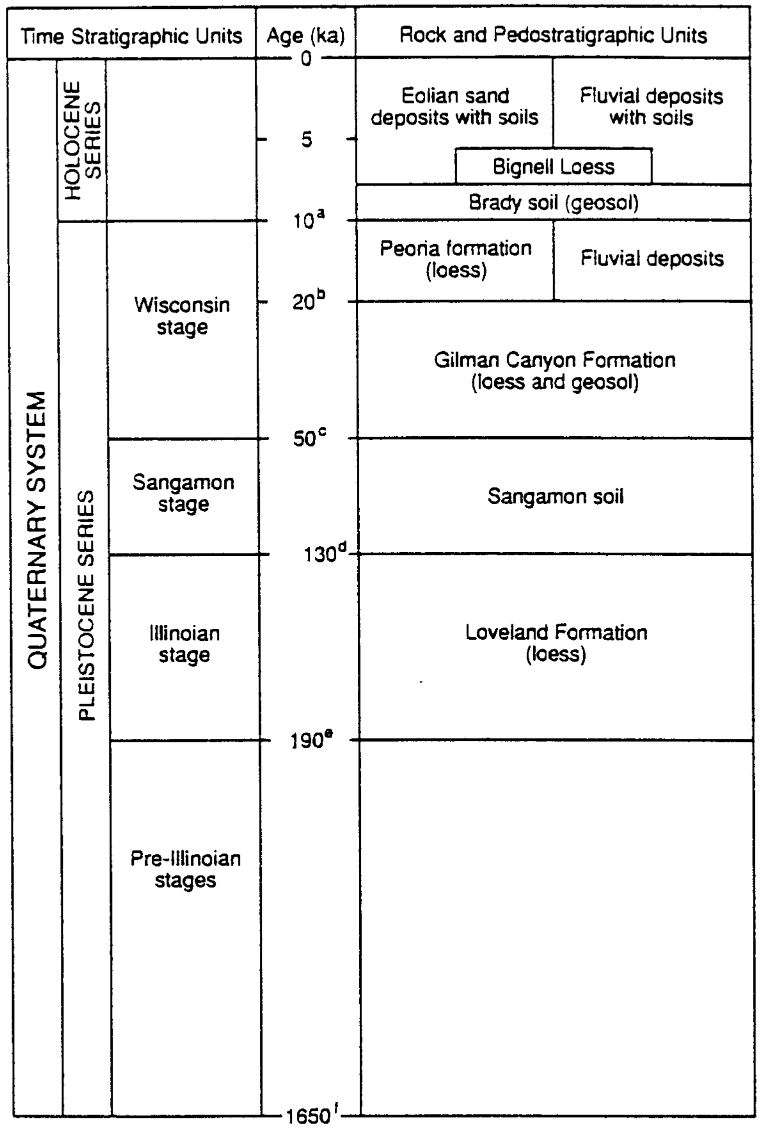


Figure 1. Late-Quaternary stratigraphic succession for the nonglaciated part of Kansas. Age designation rationale consists of the following: Pleistocene-Holocene boundary after Hopkins (1975); (b) the uppermost Gilman Canyon Formation ^{14}C dates at about 20ka (Johnson, 1993a, b); (c) the age of 50ka is an estimate for the beginning of Gilman Canyon time, not the end of Sangamon soil development, *i.e.*, the boundary is believed to be an unconformity; (d) 130ka is consistent with the sea-level record and the age of marine isotope boundary 5e/6, the presumed end of the Illinoian glacial stage and with recent themoluminescence age determinations in southcentral Nebraska and central Kansas (Maat and Johnson, 1996; Johnson and Muhs, 1996); (e) age of the marine isotope boundary 6/7 as determined by Martinson and others (1987); and (f) age of the Pliocene-Pleistocene boundary at the Virca, Italy section (Aguirre and Pasini, 1985).

Figure 2. Frequency of Phytolith Types from the Sumner Hill Site.

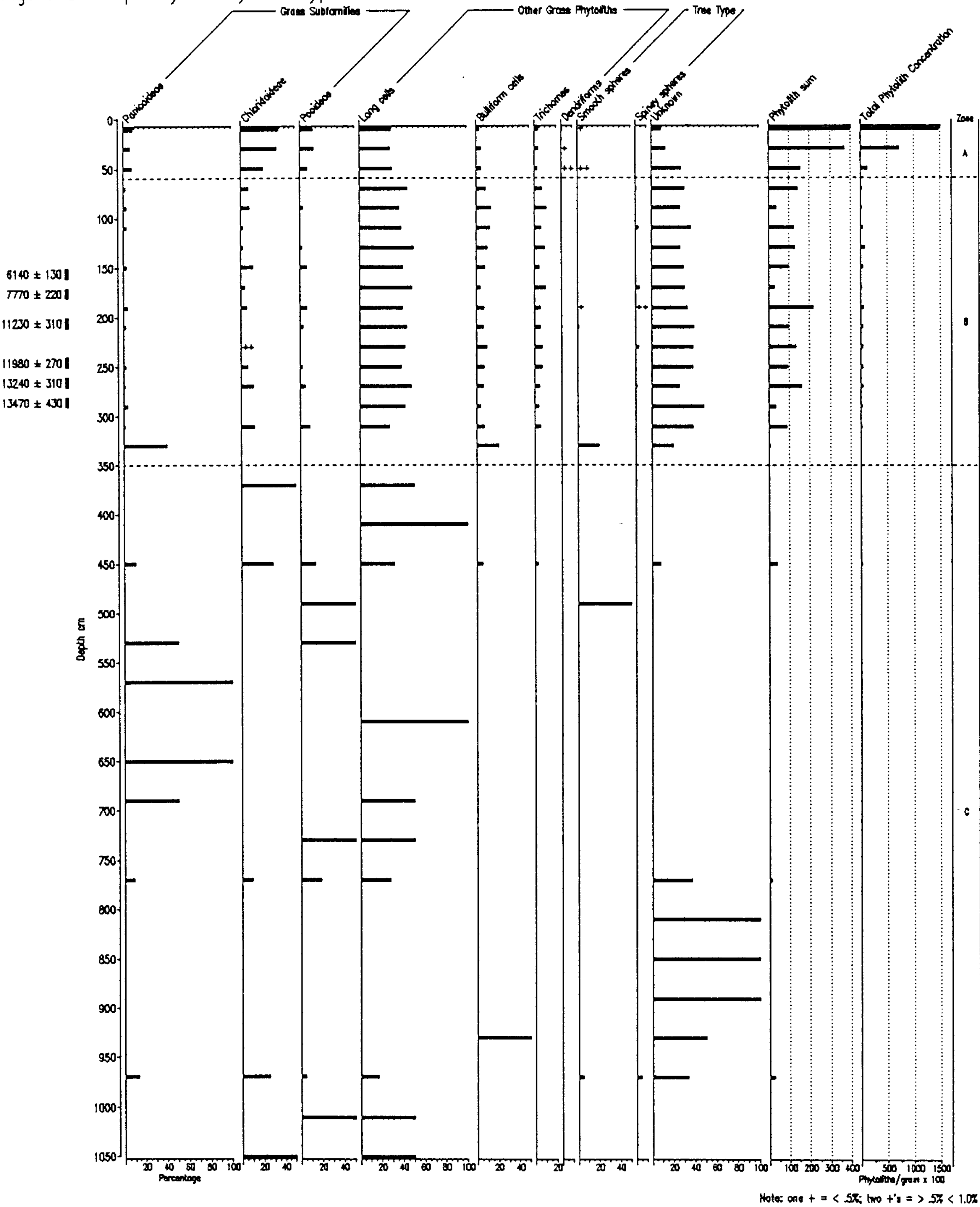


Figure 3. Frequency of Grass Short Cells and Arboreal Type Phytoliths from the Sumner Hill Site.

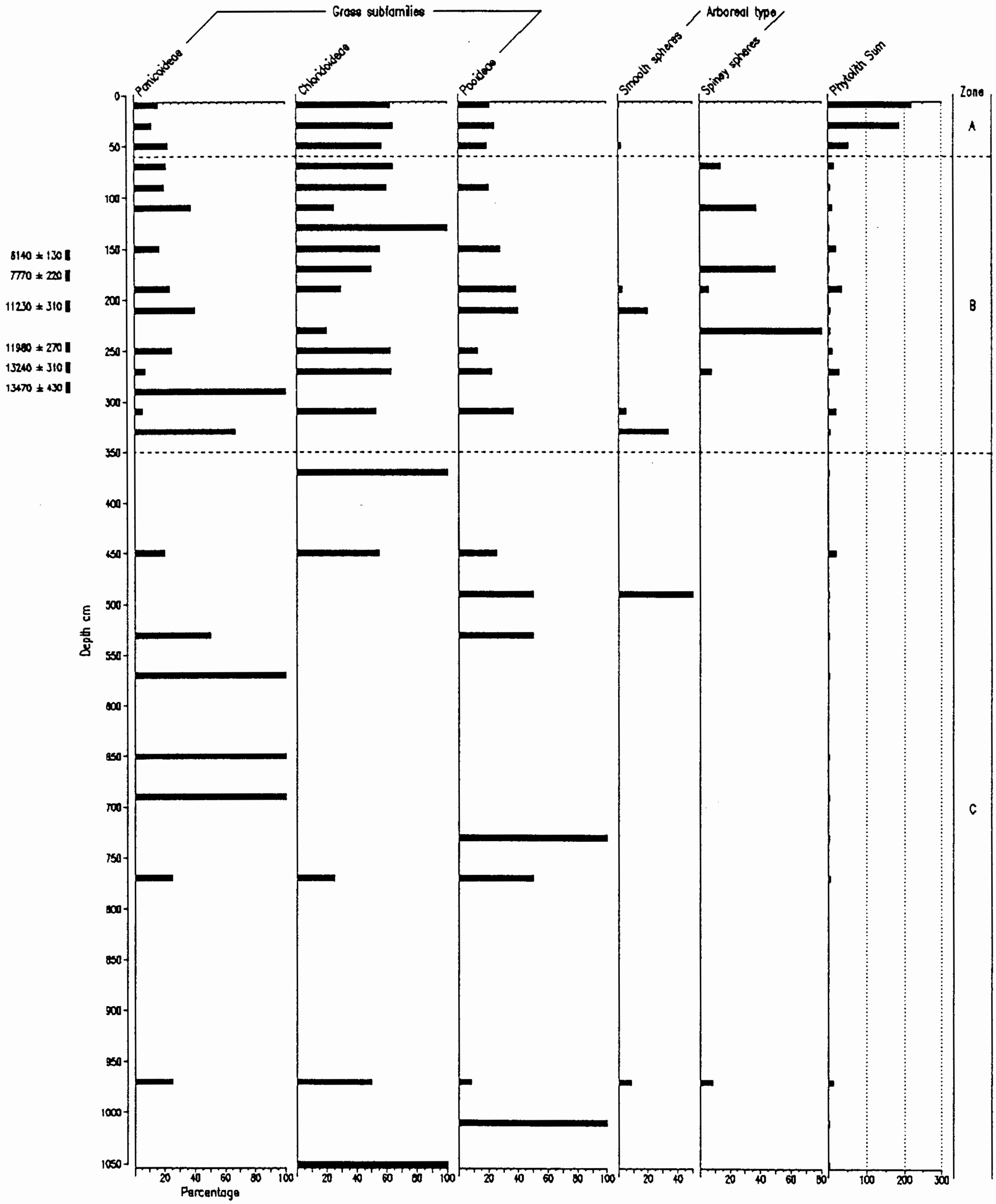


Figure 4. Frequency of Phytolith Types and Sponge Spicules from the Bala Cemetery Site.

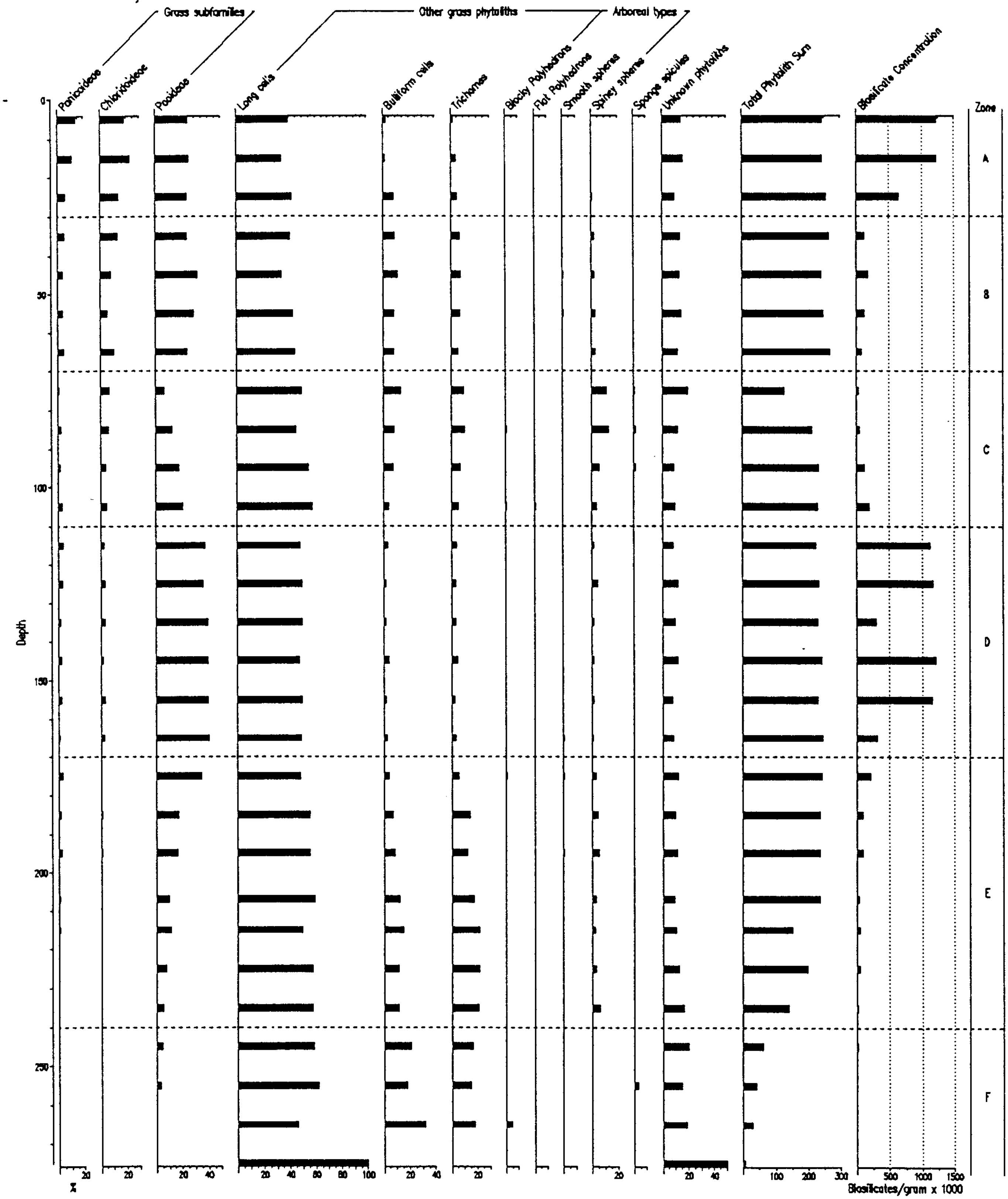


Figure 5. Frequency of Grass Short Cells and Arboreal Type Phytoliths from the Bala Cemetery Site.

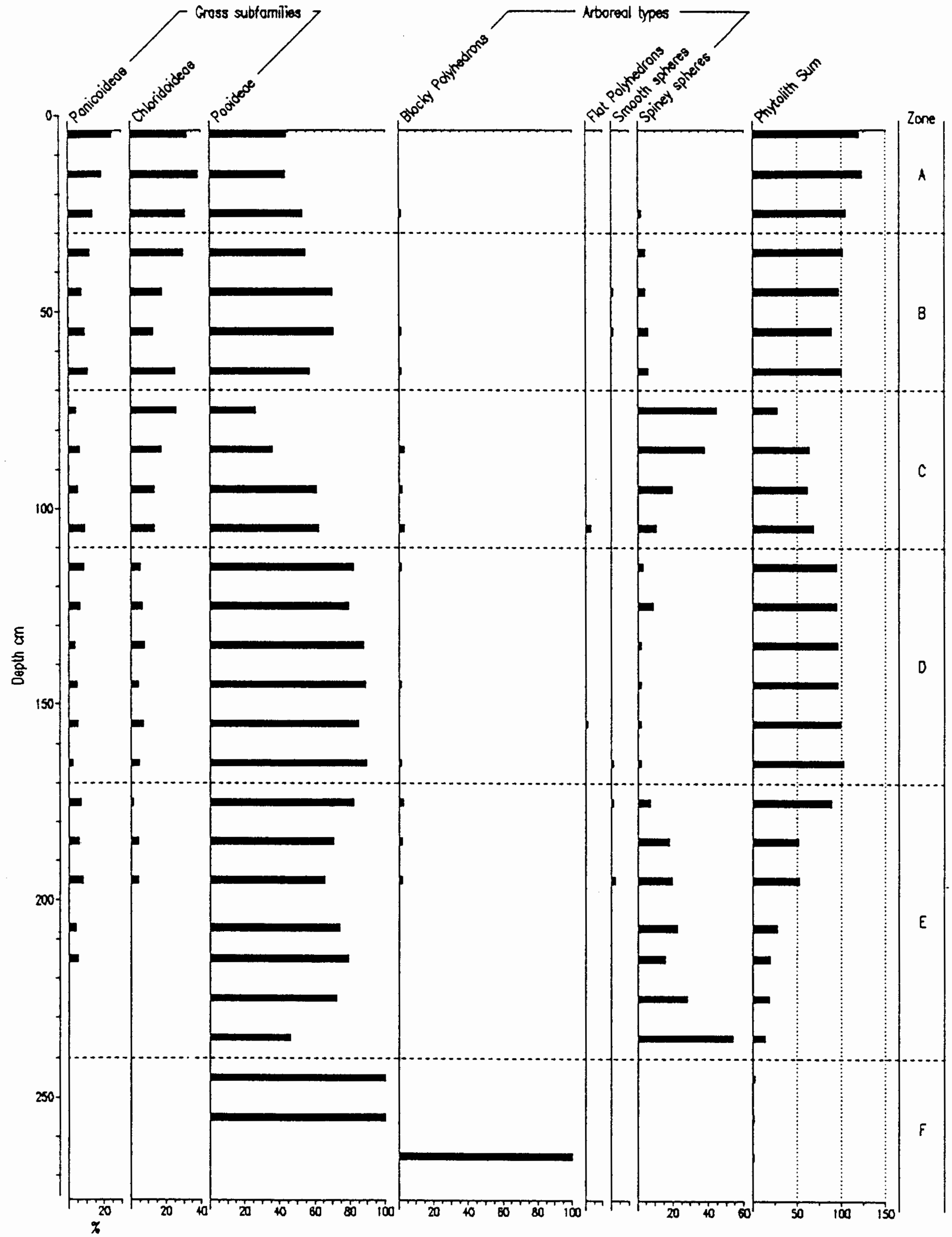


Figure 6. Frequency of Phytolith Types and Sponge Spicules from the Pump House Creek Site.

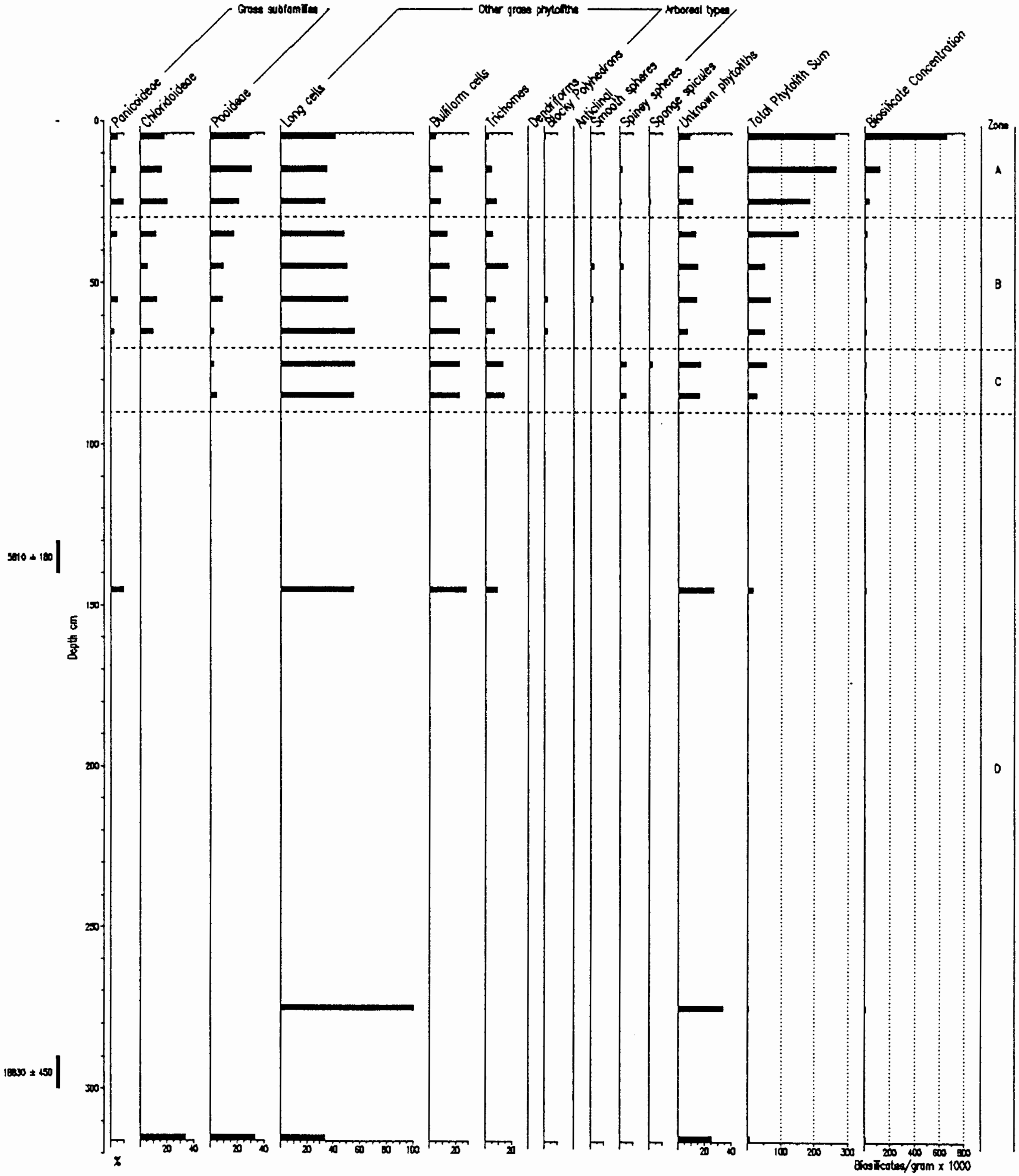


Figure 7. Frequency of Grass Short Cells and Arboreal Type Phytoliths from the Pump House Creek Site.

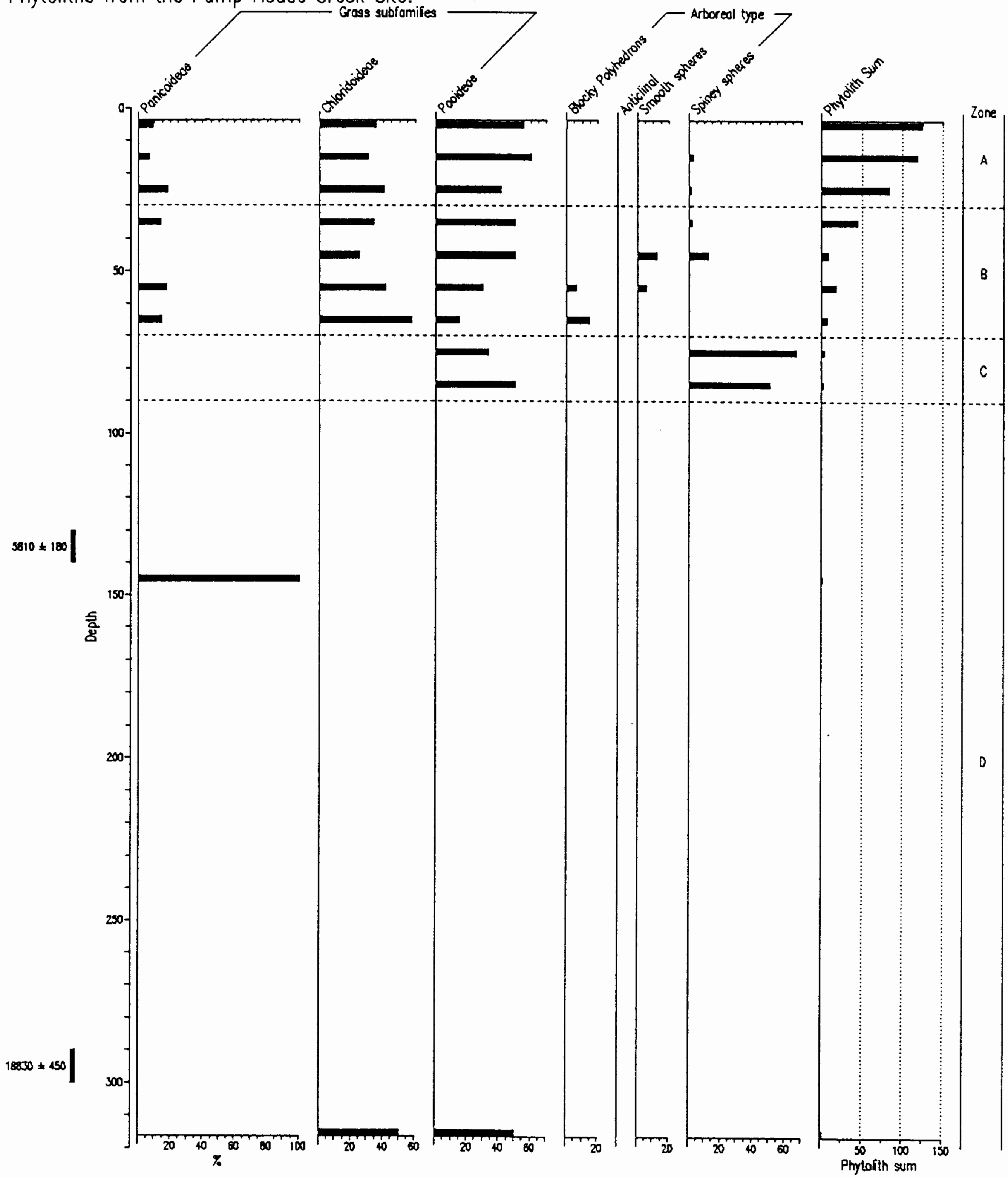


Figure 8. Frequency of Phytolith Types and Sponge Spicules from the Manhattan Airport Site.

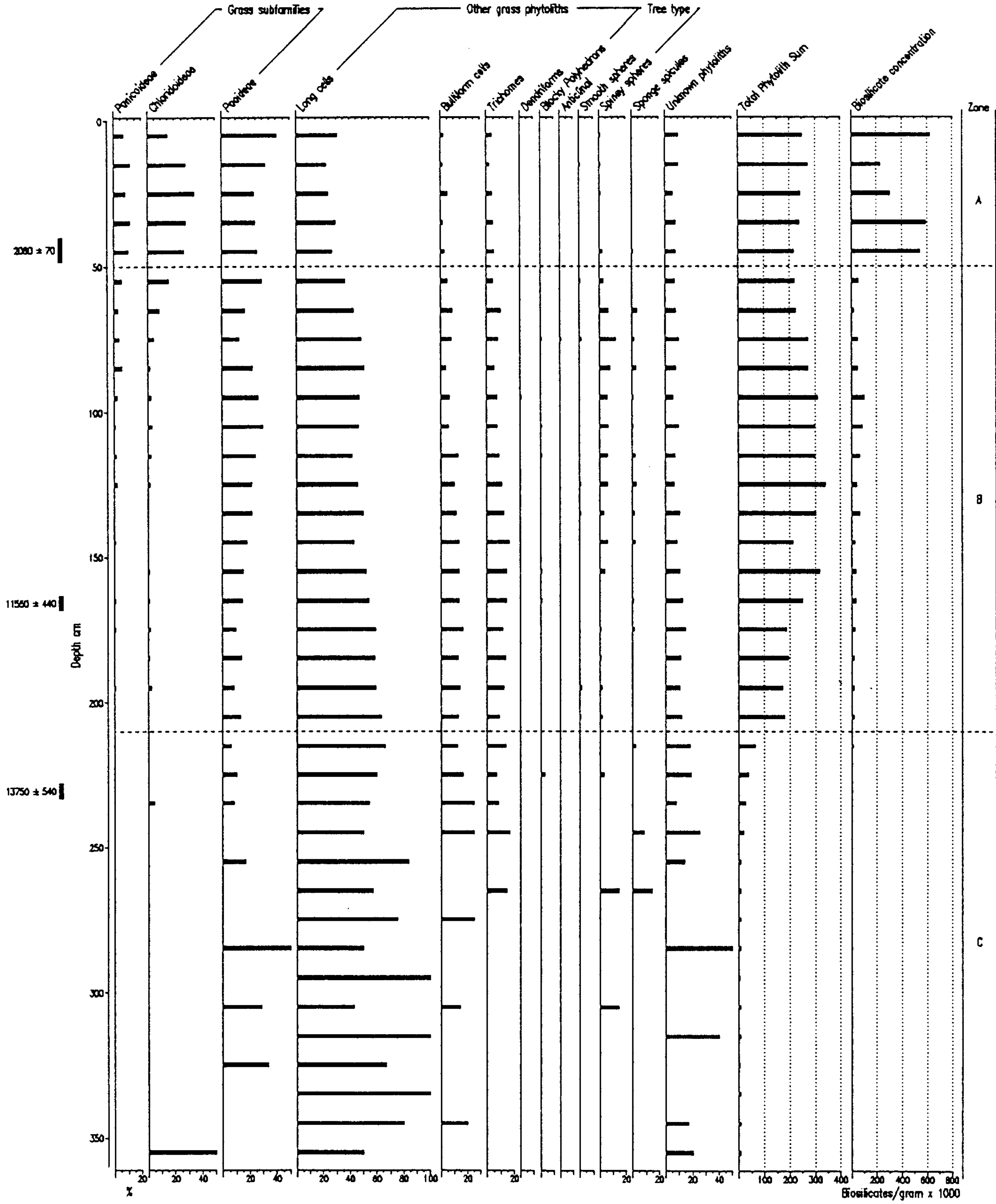
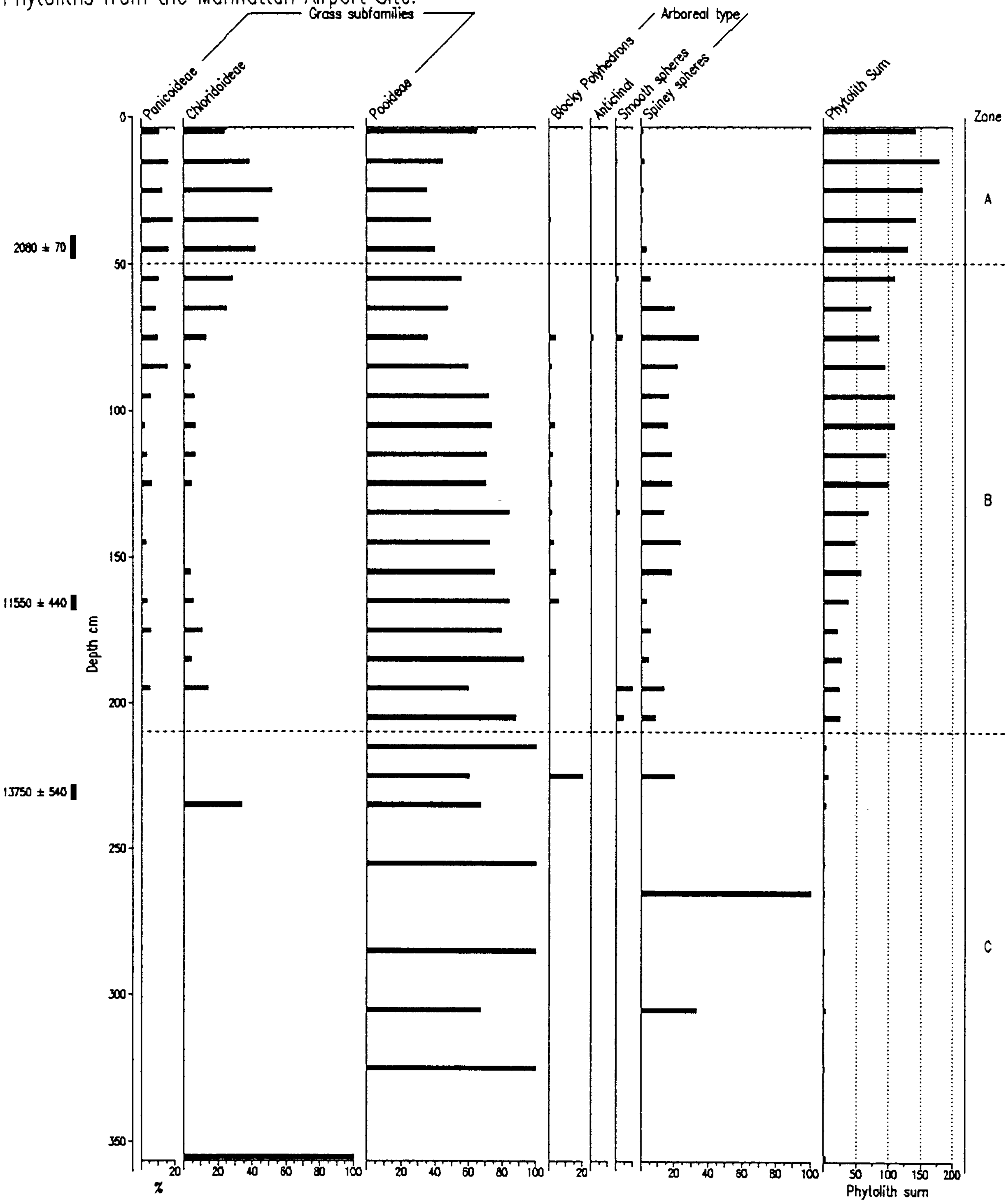


Figure 9. Frequencies of Grass Short Cells and Arboreal Type Phytoliths from the Manhattan Airport Site.



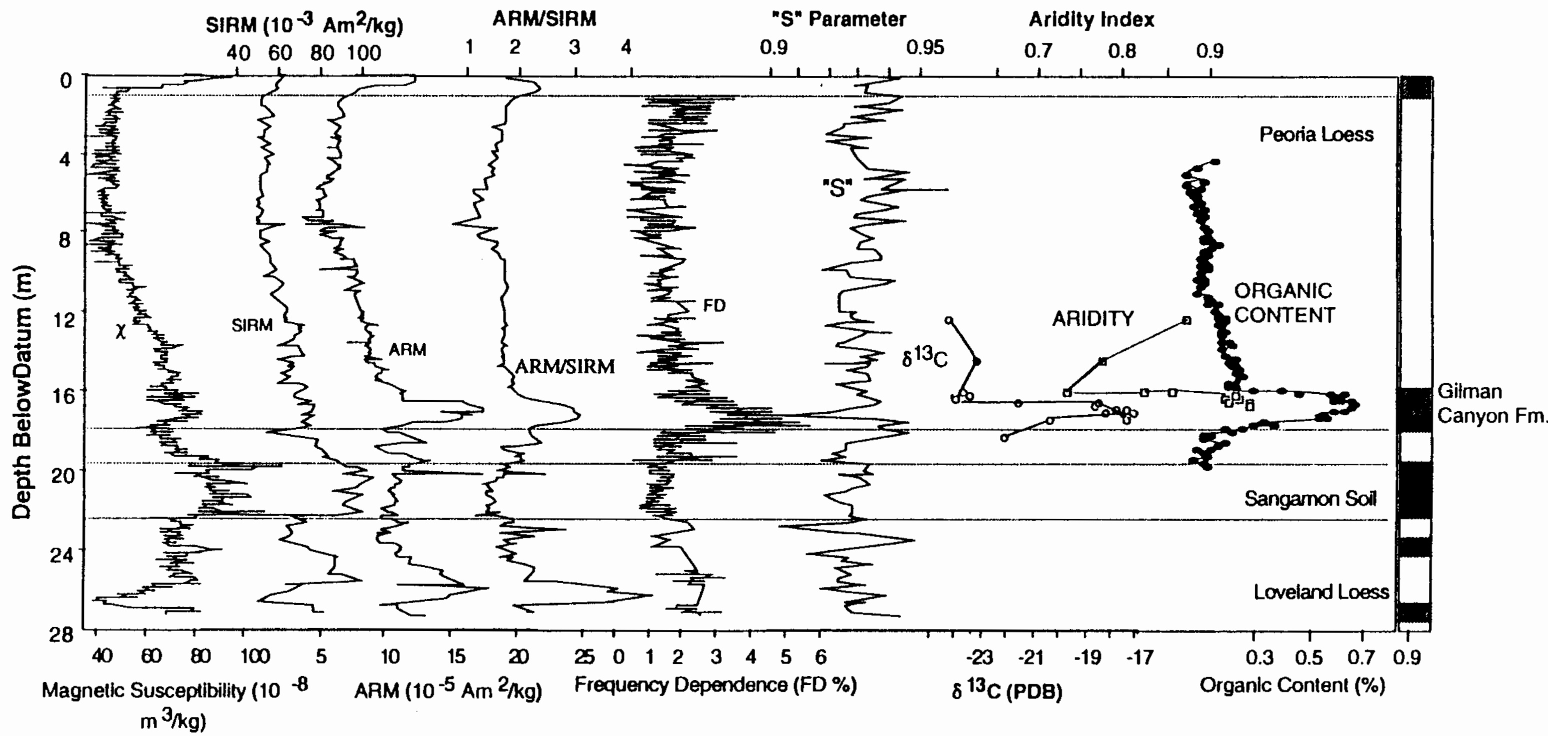
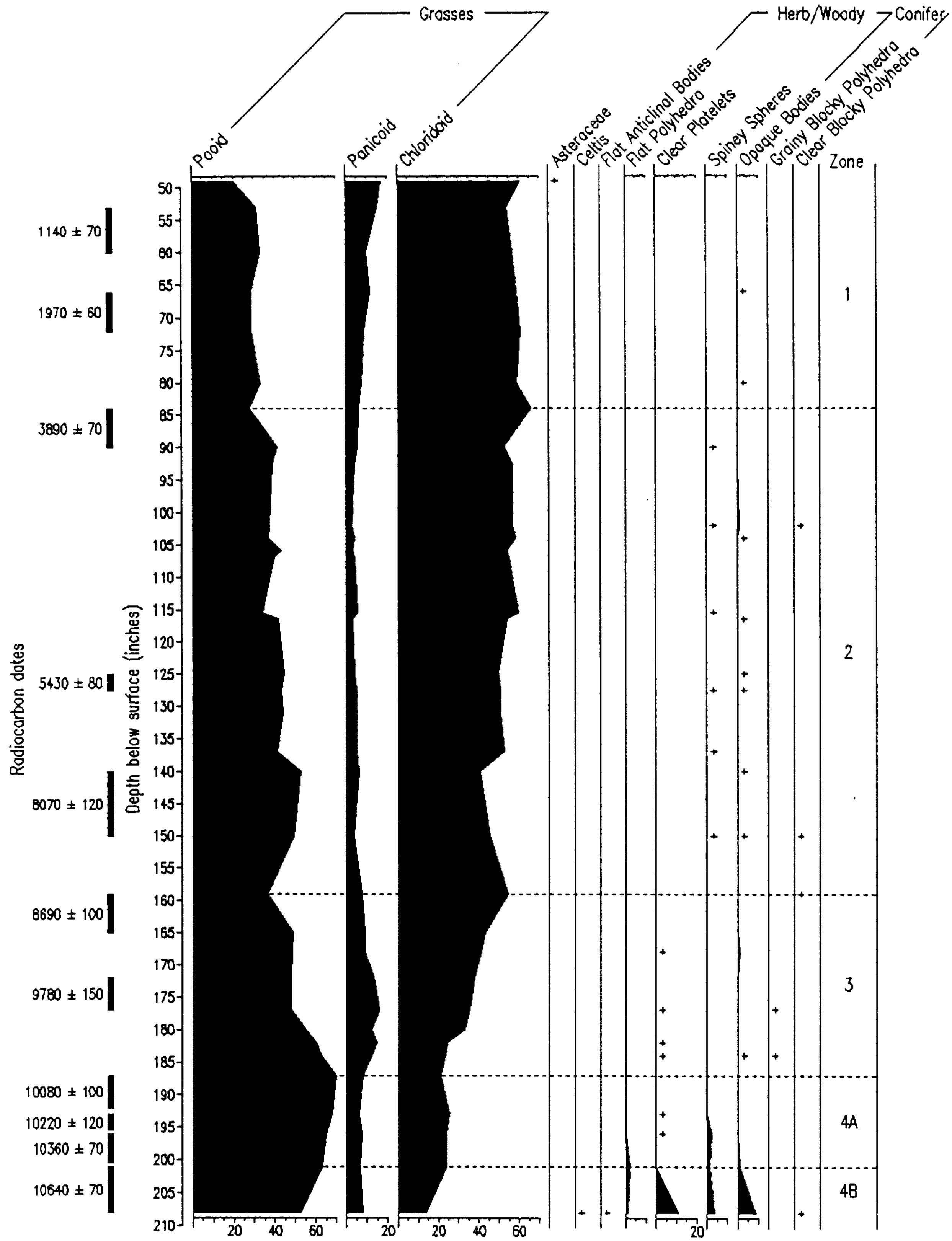


Figure 10. Stratigraphic plots of magnetic parameters (χ, SIRM, ARM, ARM/SIRM, FD, and S) compared with stratigraphic plots of $\delta^{13}C$ values derived from bulk organic carbon, an aridity index derived from opal phytolith data, and percent organic matter, all from the Eustis ash pit in southwestern Nebraska (Johnson, 1996b).

Figure 11. Frequency of Occurrence (%) of Phytoliths at the Sargent Site, Southwestern Nebraska (Bozarth 1992b).



Note: Conifer = conifer type; + represents frequencies less than 1 %

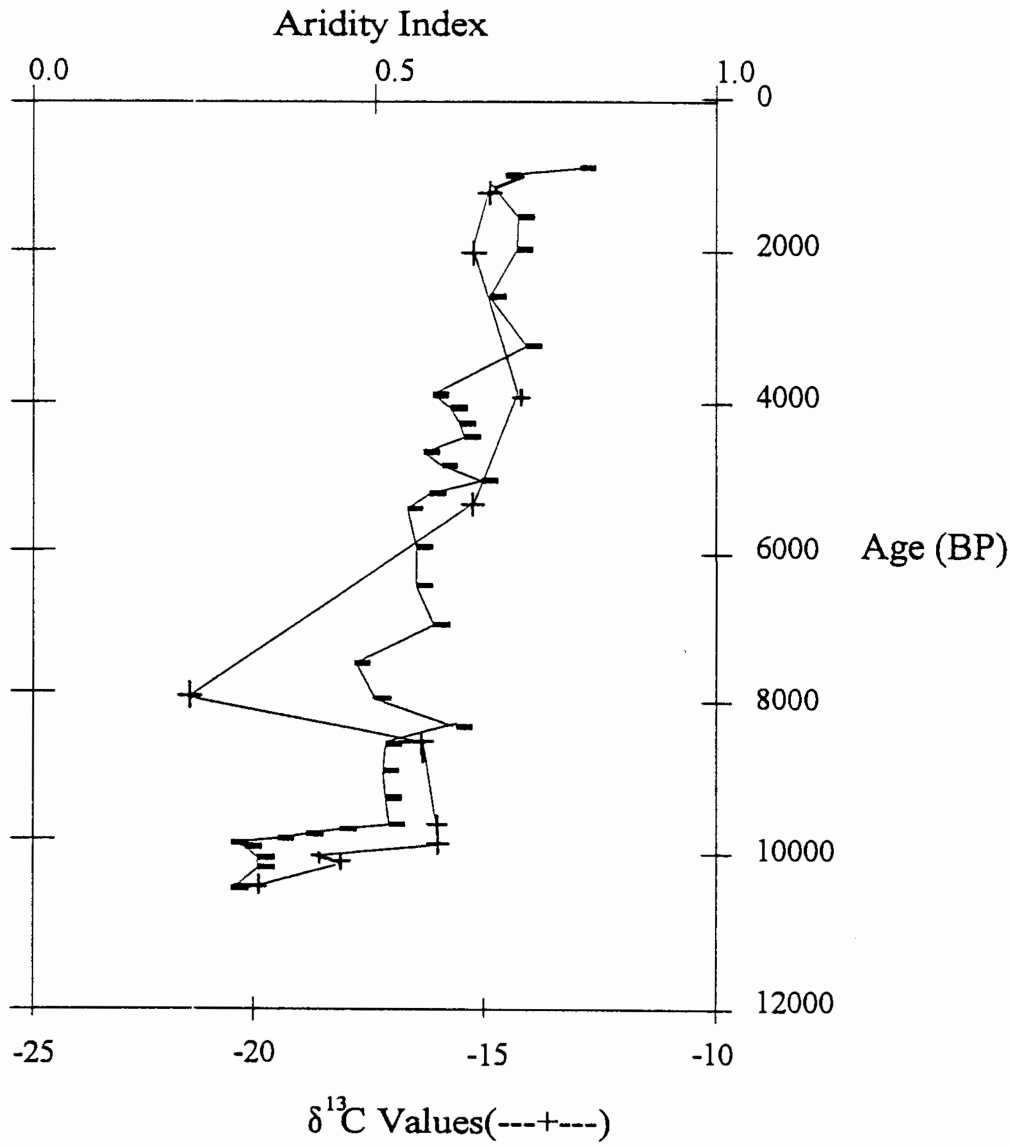


Figure 12. Aridity index (after Diekmeyer, 1994) and $\delta^{13}C$ data curves for the Sargent site, southwestern Nebraska (Bozarth, 1992b; Dort, 1996).

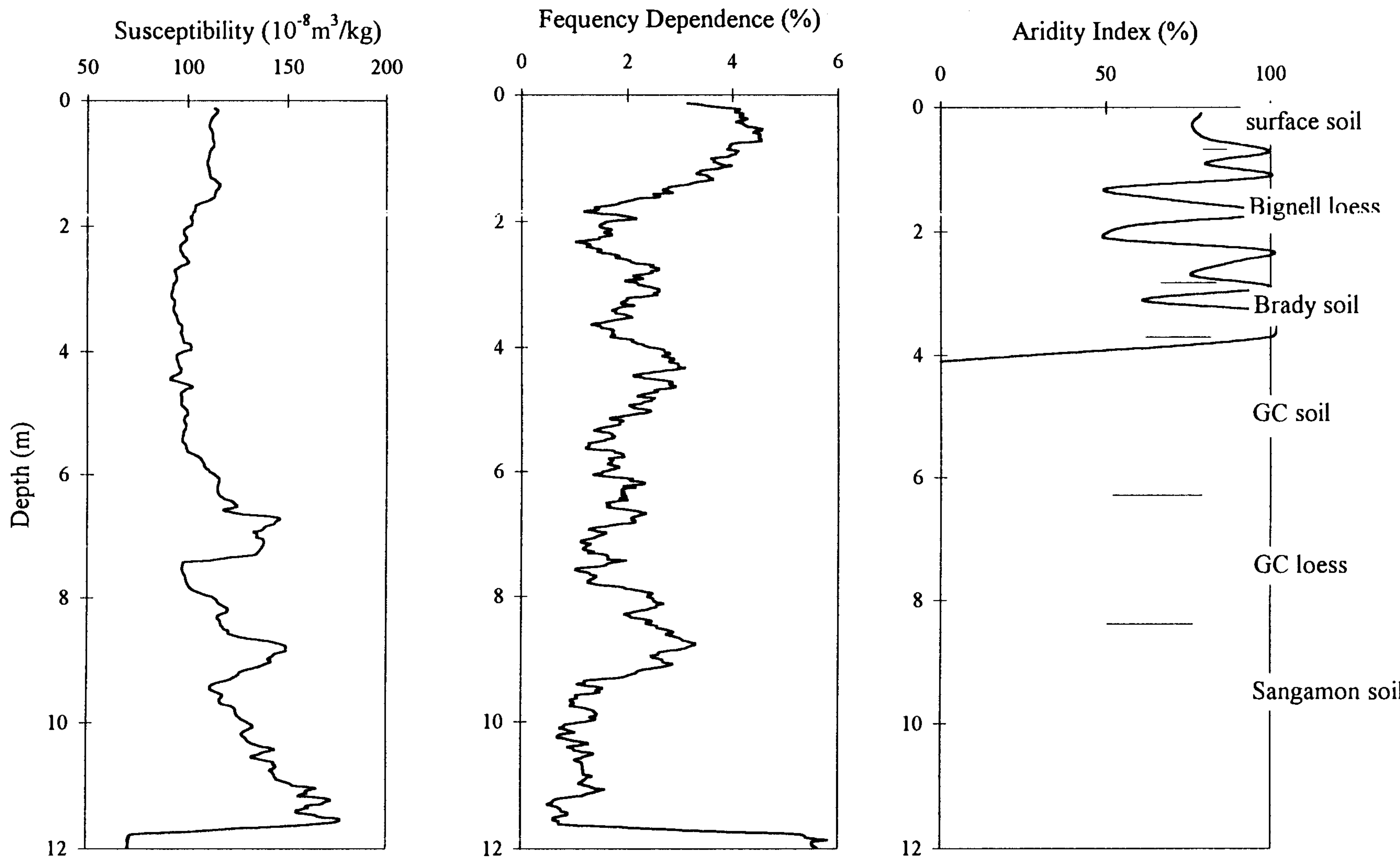


Figure 13. Magnetic susceptibility and frequency dependence and aridity index data curves for the Sumner Hill site.

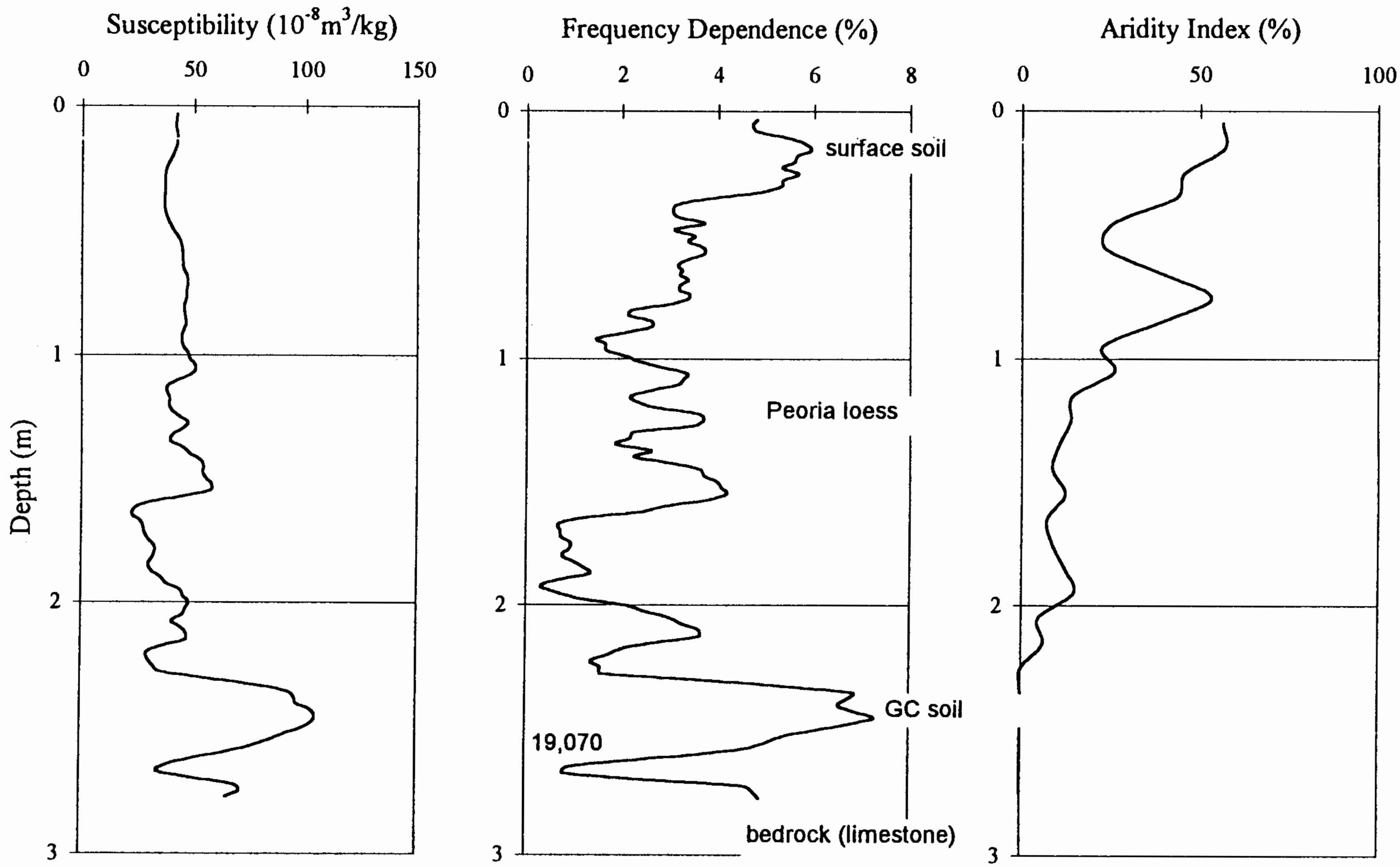


Figure 14. Magnetic susceptibility and frequency dependence and aridity index data curves for the Bala Cemetery site.

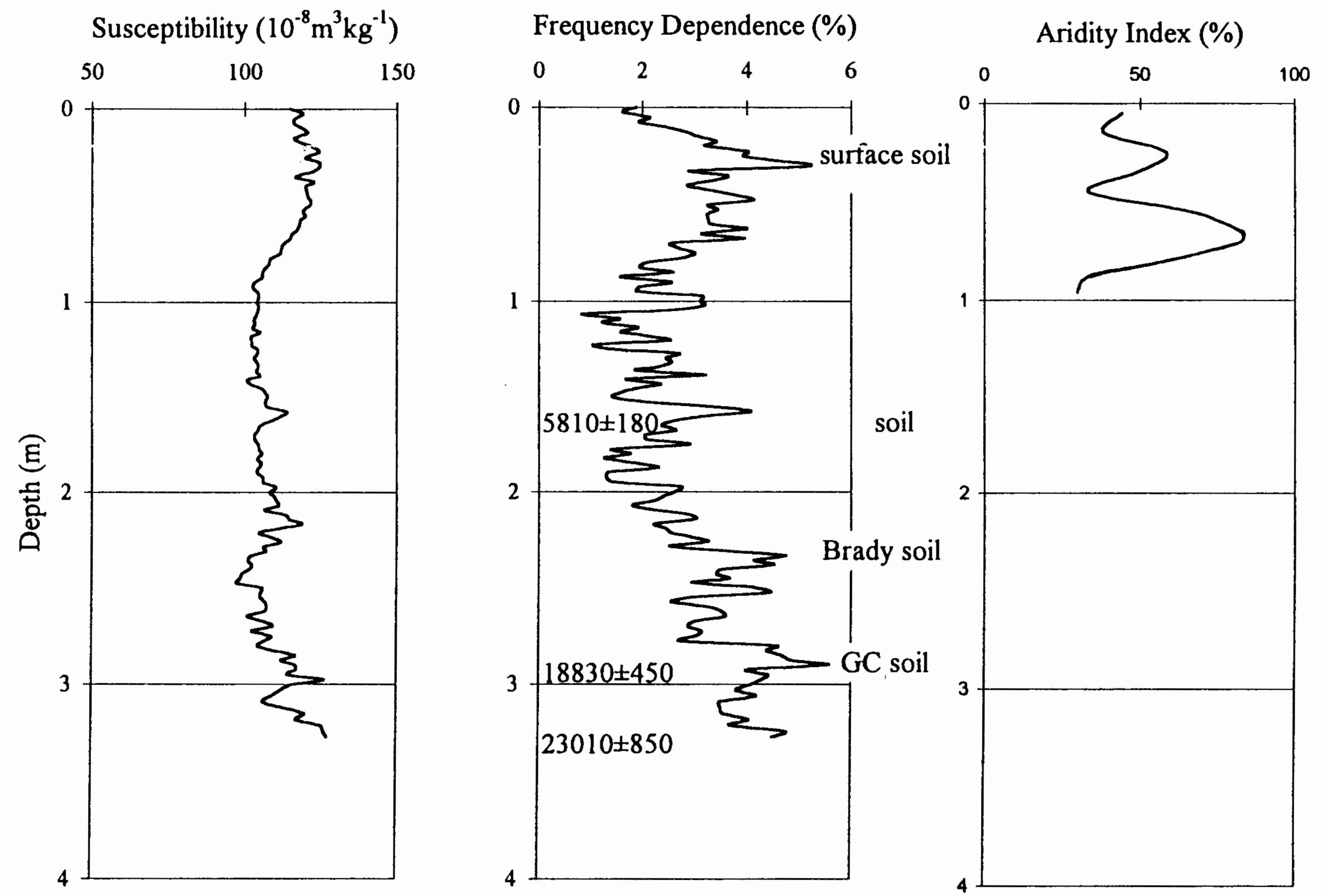


Figure 15. Magnetic susceptibility and frequency dependence and aridity index data curves for the Pump House Canyon site.

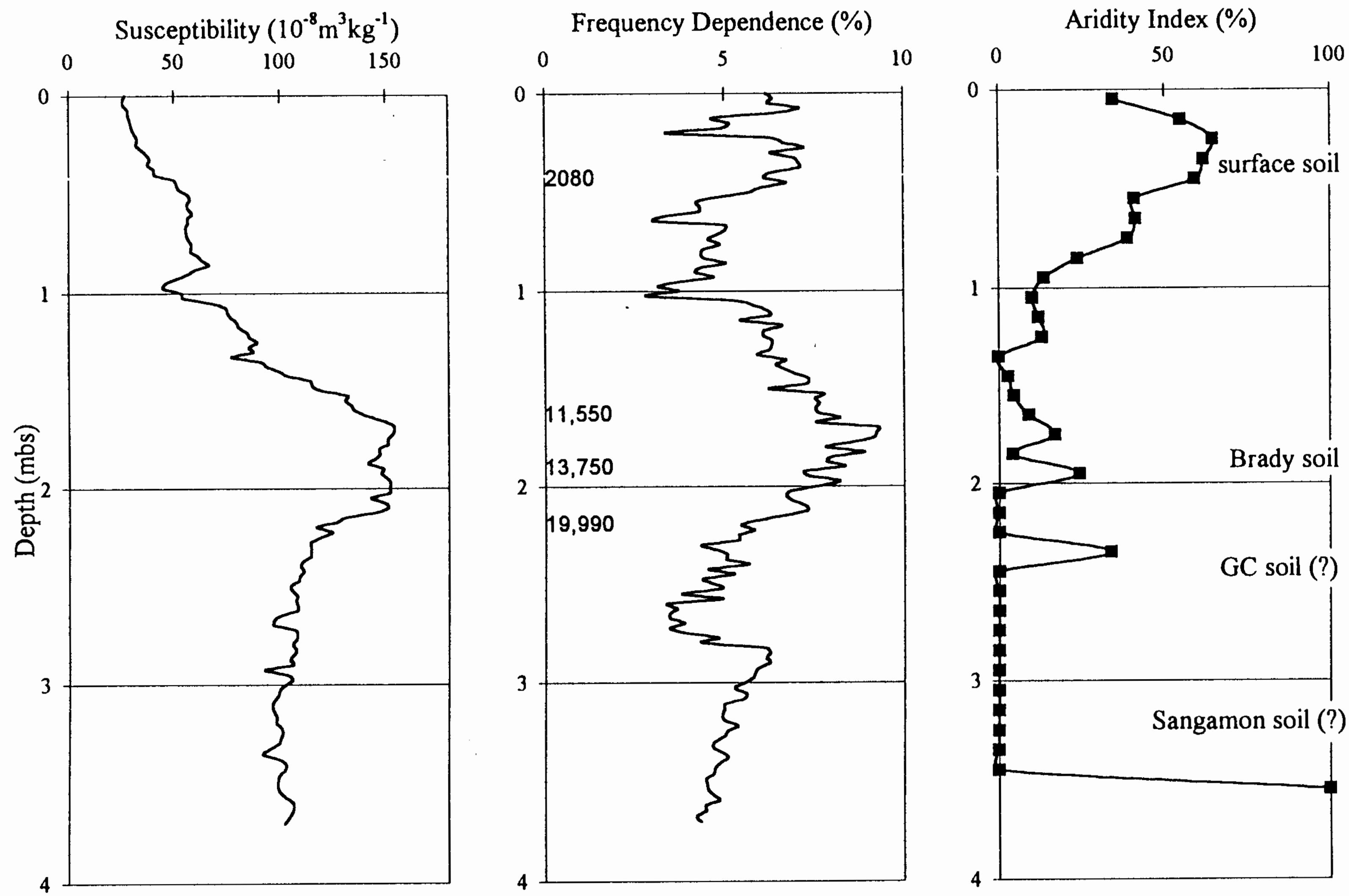


Figure 16. Magnetic susceptibility and frequency dependence and aridity index data curves for the Manhattan Airport site.

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