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# 200,000 years of climate change recorded in eolian sediments of the High Plains of eastern Colorado and western Nebraska

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## INTRODUCTION

Loess and eolian sand cover vast areas of the western Great Plains of Nebraska, Kansas and Colorado (Fig. 1). In recent studies of Quaternary climate change, there has been a renewed interest in loess and eolian sand. Much of the attention now given to loess stems from new studies of long loess sequences that contain detailed records of Quaternary glacial-interglacial cycles, thought to be a terrestrial equivalent to the foraminiferal oxygen isotope record in deep-sea sediments (Fig. 2). Loess is also a direct record of atmospheric circulation, and identification of loess paleowinds in the geologic record can test atmospheric general circulation models. Until recently, eolian sand on the Great Plains had received little attention from Quaternary geologists. The past decade has seen a proliferation of studies of Great Plains dune sands, and many studies, summarized below, indicate that landscapes characterized by eolian sand have had dynamic histories.

On this field trip, we will visit some key eolian sand and loess localities in eastern Colorado and southwestern Nebraska (Fig. 1). Stratigraphic studies at some of these localities have been conducted for more than 50 years, but others have been systematically studied only in the past few years. Many of the data which appear in this guidebook have been derived from previous studies (Swinehart and Diffendal, 1990; Madole, 1994; Loope and others, 1995; Maat and Johnson, 1996; Muhs and others, 1996, 1997a, 1999; Mason and others, 1997; Aleinikoff and others, 1999), but some are presented here for the first time.

## LOESS STRATIGRAPHY IN THE CENTRAL GREAT PLAINS

Four middle-to-late Quaternary loess units, from oldest to youngest, Loveland Loess, the Gilman Canyon Formation, Peoria Loess and Bignell Loess, have been identified and correlated on the Great Plains (Schultz and Stout, 1945; Frye and Leonard, 1951). Loveland Loess is usually no more than a few meters thick and is often the oldest loess unit exposed at many localities. In places, it appears to have an eolian sand facies. Loveland Loess is identifiable by the presence of the last interglacial Sangamon Soil in its upper part. This paleosol is usually relatively thick (1-2 m), frequently exhibits 7.5YR hues, and has well-developed prismatic or subangular blocky structure with clay films in the upper part of the B horizon and carbonate coatings or nodules in the lower part of the B horizon. The Gilman Canyon Formation is thin (usually <2 m) and typically has an organic-rich soil developed in it. Commonly, the soil developed in the Gilman Canyon Formation is welded to the upper part of the Sangamon Soil. In places, including one locality to be visited on this field trip, the Gilman Canyon Formation has two buried soils. The Gilman Canyon Formation is overlain by Peoria Loess, which is the thickest (up to ~48 m) and areally most extensive of the Great Plains loess units. A dark, organic-rich buried soil, referred to as the Brady soil, caps the upper part of the Peoria Loess, separating it from the overlying Bignell Loess. Bignell Loess is usually no more than ~2 m thick and has a patchy distribution. It has been found in Nebraska, Kansas and Colorado, but has not been reported east of the Missouri River.

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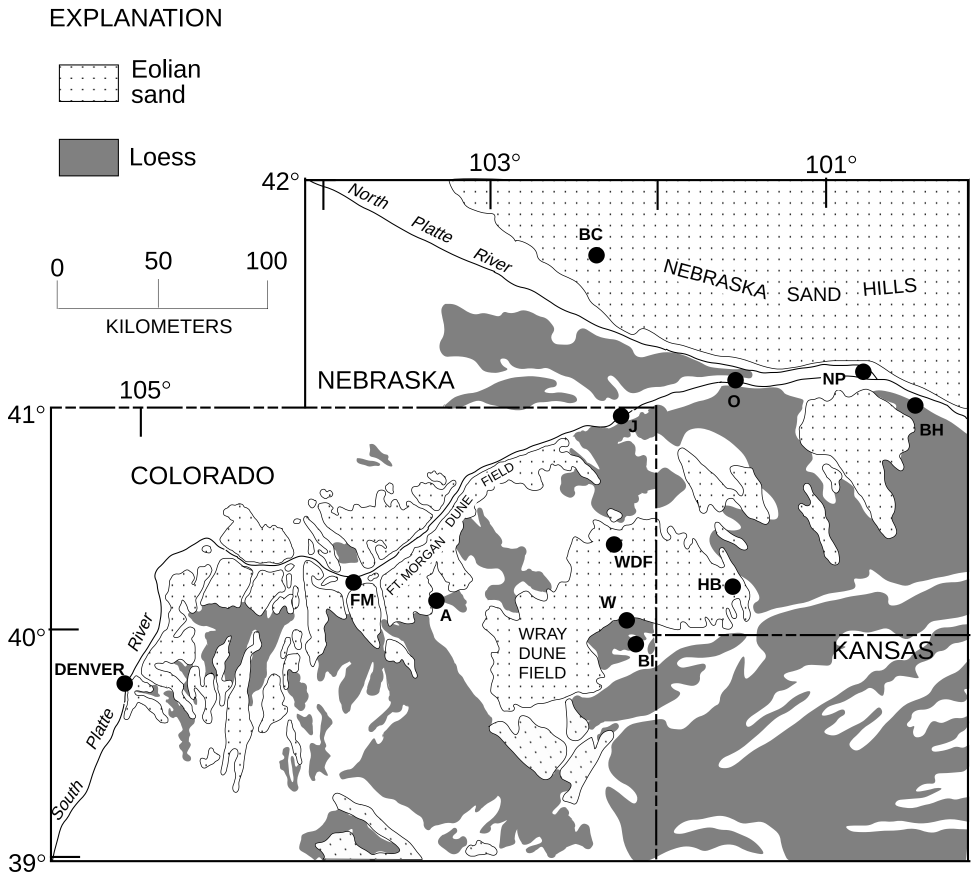


Figure 1. Map showing distribution of late Pleistocene loess and eolian sand in northeastern Colorado, southwestern Nebraska and northeastern Kansas, and field trip stops. Compiled from Swinehart (1990), Swinehart and others (1994), Kuzila and others (1990), Ross (1991), and Muhs and others (1996, 1999). FM, Fort Morgan (STOP 1); A, Akron; J, Julesburg; WDF, Wray dune field overlook (STOP 2); W, Wray (STOP 3); BI, Beecher Island (STOP 4); HB, Hoover blowout (STOP 5); BH, Bignell Hill (STOP 6); NP, North Platte; O, Ogallala; BC, Eldred Camp and Blue Creek (STOPS 7 and 8).

Geochronological studies indicate that the uppermost loess deposits on the Great Plains span the last interglacial-glacial cycle (Fig. 2). Most recent age estimates of Great Plains loesses have been from localities in Nebraska. At Eustis, Nebraska, there are numerous pre-Loveland loesses and intercalated paleosols, and a carbonate nodule from the Btk horizon of the youngest well-developed paleosol below the Loveland Loess gives a U-series age of 184,000 ± 5000 yr (analysis by B.J. Szabo, communicated to D.R. Muhs), which is a minimum-limiting age for this buried soil. Thermoluminescence (TL) ages that average about 163,100 yr have been made on the Loveland Loess itself, also at Eustis (Maat and Johnson, 1996). The TL ages and the underlying U-series age indicate that Loveland Loess could have been deposited during the penultimate glacial period, equivalent to deep-sea oxygen isotope stage 6, in good agreement with TL ages on Loveland Loess from the paratype locality in Iowa (Forman and others, 1992b). The Sangamon Soil, therefore, could have developed over a period from sometime after ~160,000 yr BP until deposition of the earliest Gilman Canyon Formation sediments, a timespan that may correspond to all of oxygen isotope stage 5 and perhaps part of stage 4 (Fig. 2). Based on radiocarbon ages of soil organic matter reported by Martin (1993), May and Holen (1993), Maat and Johnson (1996) and Muhs and others (1999), and TL analyses by Pye and others (1995) and Maat and Johnson (1996), the age of the Gilman Canyon Formation is ~40,000 to ~22,000 $^{14}$C yr BP. The Gilman Canyon Formation, therefore, corresponds in time to the mid-Wisconsin interstadial period and has its equivalent in the deep-sea record as oxygen isotope stage 3 (Fig. 2). Charcoal from spruce (Picea), as well as bone, snails, and detrital organic matter found within Peoria Loess give ages ranging from ~21,000 to ~10,000 $^{14}$C yr BP (Wells and Stewart, 1987; Martin, 1993; May and Holen, 1993; Feng and others, 1994; Maat and Johnson, 1996), and TL dating of Peoria Loess in Nebraska gives ages ranging from ~24,000 to ~12,000 cal yr BP (Pye and others, 1995; Maat and Johnson, 1996). Direct dating of probable Peoria Loess at a locality in eastern Colorado using TL methods gives ages ranging from ~20,000 to ~15,000 cal yr BP (Forman and others, 1995). All these ages indicate that Peoria Loess found in the Great Plains, as with Peoria Loess east of the Missouri River, correlates to the late Wisconsin glacial period and deep-sea oxygen isotope stage 2 (Fig. 2). Maximum-limiting ages of Bignell Loess are based on radiocarbon ages of organic matter from the Brady soil, and range from ~11,800 to ~8,000 $^{14}$C yr BP (Martin, 1993; Maat and Johnson, 1996; Muhs and others, 1999). Direct dating of Bignell Loess using TL gives ages ranging from ~9,000 to ~3,000 cal yr BP (Pye and others, 1995; Maat and Johnson, 1996).

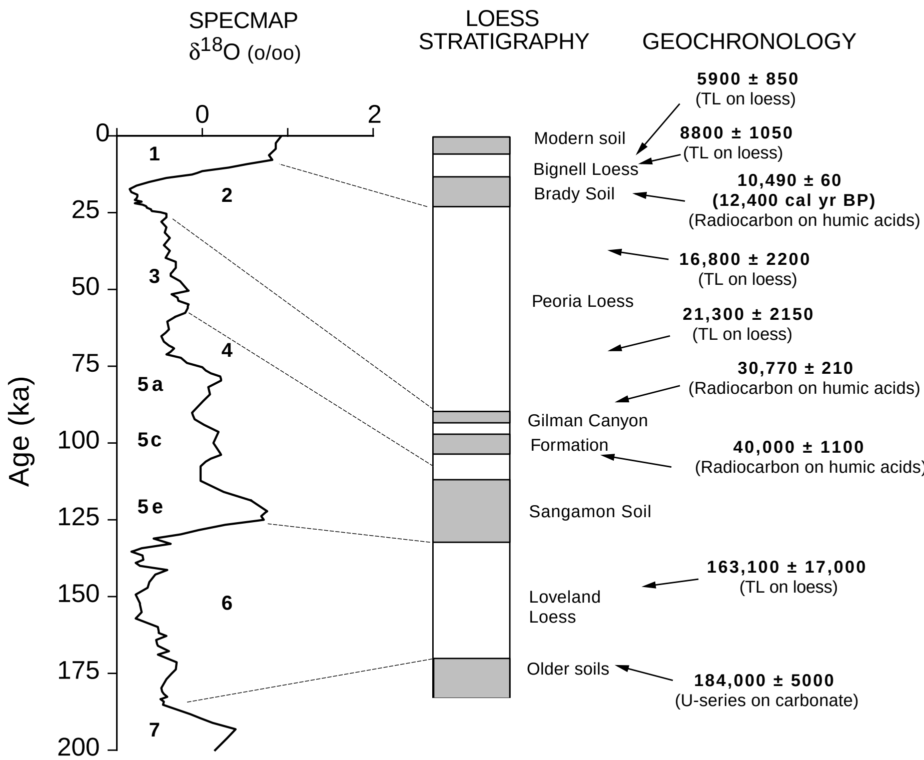


Figure 2. Generalized loess stratigraphy of the western Great Plains, with age estimates for deposits and soils and possible correlation to the deep-sea oxygen isotope record. Age data for western Great Plains loess from Maat and Johnson (1996) and Muhs and others (1999); oxygen isotope data from Martinson and others (1987).

## EOLIAN SAND STRATIGRAPHY IN THE CENTRAL GREAT PLAINS

The most extensive eolian sands in North America are found in the central and southern Great Plains. Limited radiocarbon ages and degree of soil development suggest that eolian sand sheet and dune deposition took place during the last glacial period in Nebraska and Colorado. Late glacial eolian activity is supported by maximum-limiting ages of ~13,000 $^{14}$C yr BP for some of the largest barchanoid ridges in the Nebraska Sand Hills (Swinehart and Diffendal, 1990) and evidence for eolian sand movement in the southwestern part of the dune field (Loope and others, 1995), seen on this trip. In Colorado, radiocarbon and soil evidence indicate that eolian sheet sands were deposited over large areas during the last glacial period (Madole, 1995; Muhs and others, 1996).

The mid-Holocene (~8000-5000 $^{14}$C yr BP) has long been considered to be a dry period in central North America (Webb and others, 1993), and would seem to be an optimum time for eolian sand movement. However, there are actually few records of paleoclimatic conditions from the Great Plains itself during this period, and there is not widespread evidence for mid-Holocene eolian sand activity in the region. Radiocarbon ages from a few localities in the Nebraska Sand Hills indicate some probable mid-Holocene eolian activity (Loope and others, 1995; Stokes and Swinehart, 1997). In Colorado, Forman and co-workers (Forman and Maat, 1990; Forman and others, 1992a, 1995) infer that the mid-Holocene was an important time of eolian sand movement, but only one locality really has definitive evidence for this. In Texas, mid-Holocene eolian sand is found in the stratigraphic record of dry valleys (Holliday, 1989; 1995b). Elsewhere in the Great Plains, the record for mid-Holocene eolian sand deposition is scanty, and is based on indirect lines of evidence such as dunes with a degree of soil development that is greater than that on late Holocene eolian sand, but less than that on late Pleistocene deposits (Madole, 1995; Muhs and others, 1996). Although it seems likely, from independent evidence, that conditions were optimal for eolian sand activity over the central Great Plains during the mid-Holocene, extensive late Holocene eolian activity may have removed much of the geomorphic record.

Both radiocarbon and luminescence methods demonstrate that eolian sands over much of the Great Plains have been active in the past 3,000 yr (Ahlbrandt and others, 1983; Swinehart and Diffendal, 1990; Madole, 1994, 1995; Holliday, 1995a, 1997a, 1997b; Forman and others, 1992a, 1995; Loope and others, 1995; Muhs and Holliday, 1995; Arbogast, 1996; Muhs and others, 1996, 1997a, 1997b; Wolfe and others, 1995; Stokes and Swinehart, 1997). In addition, most of these studies have stratigraphic data indicating multiple periods of eolian activity in the late Holocene. The number of radiocarbon ages and their analytical uncertainties do not yet make it possible to test the hypothesis of regional synchroneity of activity. However, these observations indicate that, contrary to earlier beliefs, eolian sands in this region can be active under an essentially modern climatic regime.

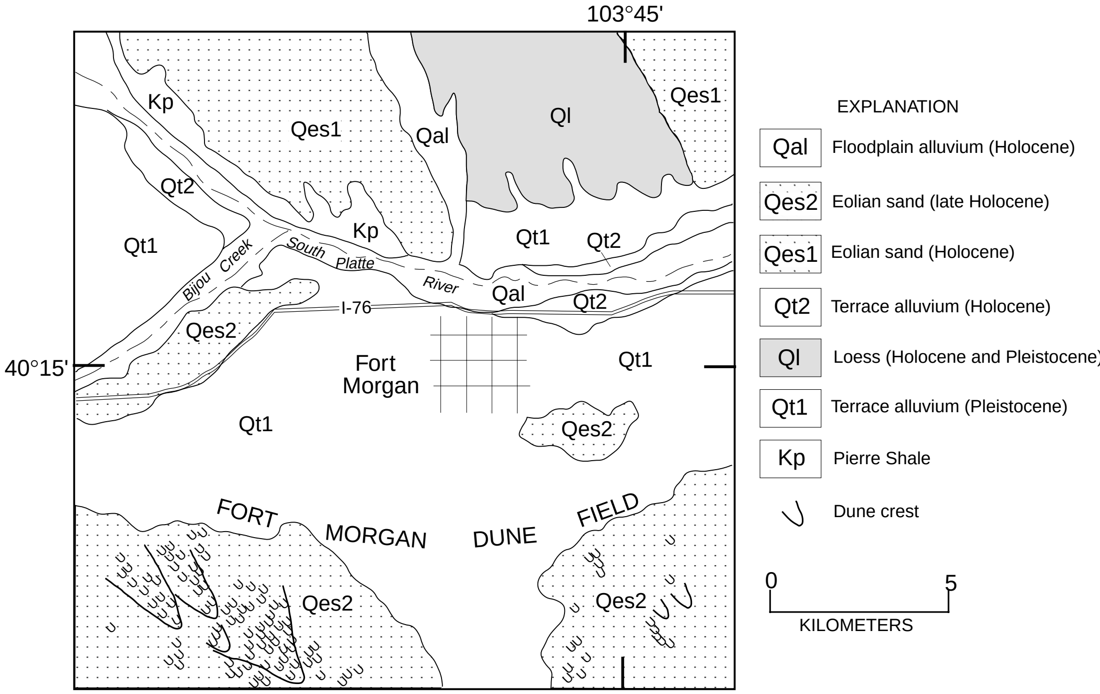


Figure 3. Simplified geologic map of the Fort Morgan, Colorado area (STOP 1), based on unpublished aerial photograph interpretation and field mapping by D.R. Muhs, and soil survey data of Spears and others (1968).

## BETWEEN DENVER AND STOP 1

Eolian deposits can be seen shortly after passing the town of Hudson on I-76. East of Hudson, at the Kersey Road exit, there is a cut made for a railroad visible on the right. The sediments exposed in this cut have been studied by Forman and others (1992a, 1995) and Madole (1995). Between Hudson and Keenesburg, the landscape you see is covered with 2-3 m of late Wisconsin (Peoria) loess. A few miles east of Keenesburg, as you climb the hill, the Fort Morgan dune field will be visible, on both sides of I-76. The landforms here are low-relief parabolic dunes and eolian sand sheets, and can be easily distinguished by the abundance of Artemisia (sage) cover and frequent blowouts. The soils here belong mostly to the Valent series, and have simple A/AC/C profiles. Between mileposts 45 and 46, just before entering Roggen (exit 48), partially active dunes can be seen on both sides of the highway, but particularly on the north side. These dunes were active during the 1930s drought; just after passing through Roggen, there is a good view to the north of other dunes that were also active in the 1930s and are barely stable now. After passing by the town of Wiggins, I-76 traverses part of the Broadway-Kersey-Qt1 terrace (see discussion below); just outside of Wiggins, the terrace can be viewed to the north, eolian sand to the south.

## STOP 1: SOUTH PLATTE RIVER NEAR FORT MORGAN

The major drainage for northeastern Colorado is the South Platte River, which heads in the Colorado Rocky Mountains near Colorado Springs and joins the North Platte River just east of the city of North Platte, Nebraska. Because the river figures prominently in the origin of both eolian sand and loess in the region, our first stop will be a short landscape view of the South Platte River near Fort Morgan. Fort Morgan itself and much of I-76 in this area are built on what has been called the "Broadway" (Scott, 1978), "Kersey" (Holliday, 1987) and "Qt1" (Muhs and others, 1996) terrace, of probable late-glacial age. It is as much as 8 km wide in places (Fig. 3). To the south, compound parabolic dunes of the Fort Morgan dune field, with their minimally developed soils (A/AC/C profiles) are visible and overlie this terrace. To the northwest and northeast, more subdued dunes and sand sheets of the Sterling dune field, with their relatively well developed soils (A/Bt/Bk/C profiles), can be seen. Immediately north of Fort Morgan, the highest part of the landscape is an alluvium-mantled upland overlain by loess.

Isotopic analyses (discussed at STOP 4) indicate that South Platte River sediments were a contributing, but not sole source of loess in eastern Colorado (Aleinikoff and others, 1999). Geochemical and isotopic analyses demonstrate that the most likely source of sediment for both late Pleistocene and Holocene eolian sand in the Fort Morgan and Wray dune fields (Fig. 1) was the South Platte River (Muhs and others, 1996).

Stratigraphic and radiocarbon studies by Madole (1994) show that the most recent episodes of eolian sand movement in the Fort Morgan dune field occurred in the past ~1500 yr (Fig. 4), and helped build the compound parabolic dunes

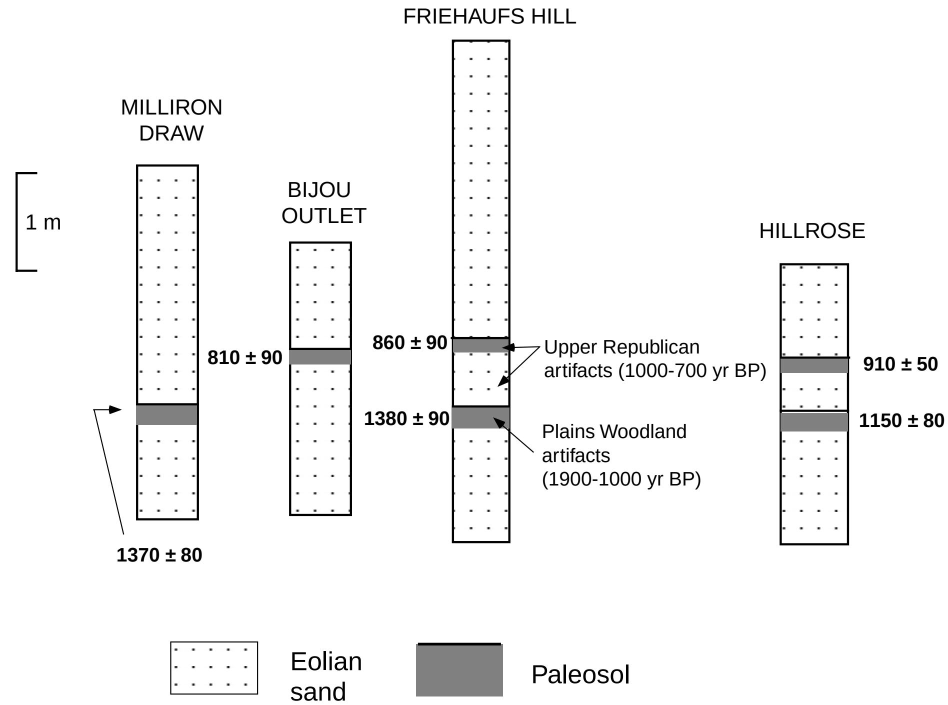


Figure 4. Stratigraphy and radiocarbon ages of late Holocene eolian sands from sections in the Fort Morgan dune field. Redrawn from Madole (1994).

shown in Fig. 3. Stratigraphic studies through augering show that the late Holocene parabolic dunes are underlain by eolian sheet sands that also occur at the surface in interdune areas and have well-developed soils (Muhs and others, 1996). These older sheet sands have maximum-limiting radiocarbon ages of ~27,000 yr BP, suggesting they were probably deposited during the last glacial period.

## BETWEEN STOPS 1 AND 2

From the city of Brush (immediately east of Fort Morgan) to ~25 km southwest of Julesburg, on the Colorado/Nebraska state line, I-76 is built on late Holocene sands of the Fort Morgan dune field. This dune field parallels the South Platte River (which provided the source sediments) for more than 100 km northeast of Brush (Fig. 1). Near Julesburg, we turn south on U.S. 385. Note the exposure of carbonate-cemented sand and gravel in Miocene Ogallala Group rocks behind the gas station on the right immediately after turning onto highway 385. Lugn (1968) and Hunt (1986) considered sediments derived from Ogallala Group rocks to be the most important source of eolian sands on the Great Plains, a concept no longer supported, at least for dunes in northeastern Colorado (Muhs and others, 1996). About 10 km south of the turnoff onto 385, we ascend the High Plains surface on Ogallala Group rocks. Scott (1978) mapped loess on much of this surface. New field work by Muhs and others (1999) confirms the presence of this loess, but it is patchy and frequently only about a meter thick. About 9 km south of Holyoke, we enter the Wray dune field.

## STOP 2: GEOMORPHOLOGY OF THE WRAY DUNE FIELD

This stop is an overview of landforms in the Wray dune field, the largest eolian sand body in Colorado and southwestern Nebraska. Sediments in this dune field range from medium to fine sand, and show a general northwest-to-southeast fining, consistent with dune orientations that indicate northwesterly winds at the time of formation (Fig. 5). At this stop, we are near the northern edge of the Wray dune field, but the landscape here has some of the finest examples of parabolic dune forms in eastern Colorado. At the intersection of the county line and highway 385, we are standing on the left arm (as you look downwind) of a small parabolic dune that is part of a much larger compound parabolic dune (Fig. 6). Other small dunes that are part of this megadune are visible to the southwest and U.S. Highway 385 itself cuts through two arms of such a dune. Simple parabolic dunes are also found in the area, and are of approximately the same dimensions as those that make up the compound dunes (Fig. 6). Soils on both the simple and compound parabolic dunes belong to the Valent series, an Ustic Torripsamment with a simple A/AC/C profile, indicating relatively young deposits, similar to those in the Fort Morgan area.

Interdune areas in this region are occupied by eolian sand sheets, such as the one with the center-pivot irrigation system visible to the southwest. These eolian sands have much better developed soils with A/Bt/Btk/C profiles, indicating a considerably greater age than late Holocene. Although stratigraphic studies have yet to confirm it, it is likely that, as with the Fort Morgan dune field, these older eolian sheet sands underlie the parabolic dunes.

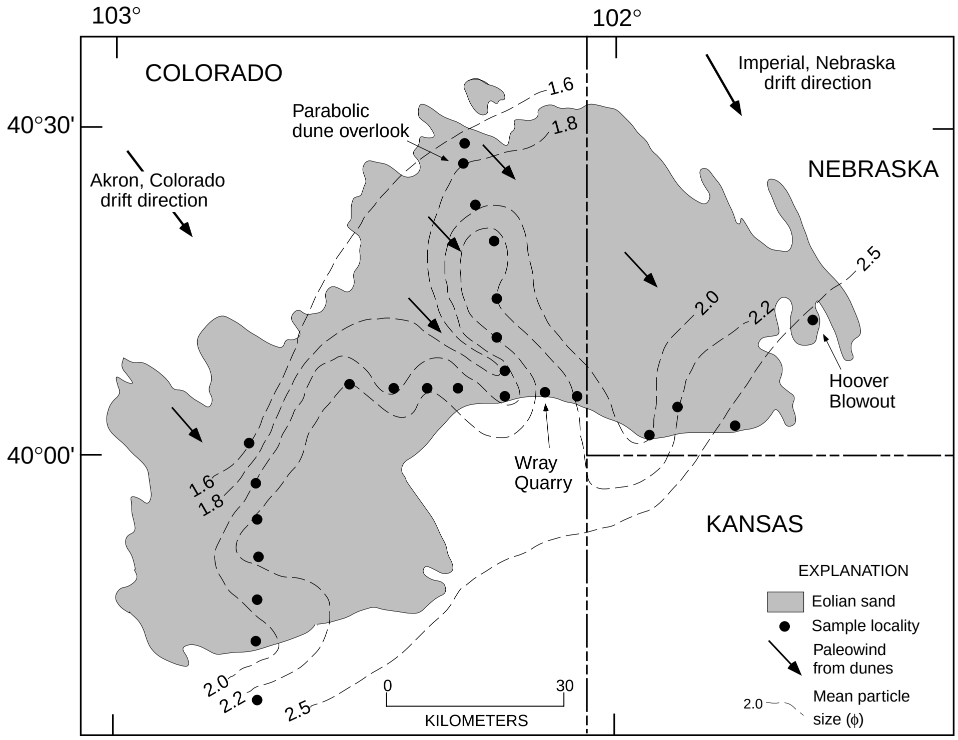


Figure 5. Map of the Wray dune field, Colorado and Nebraska, locations of STOPS 2, 3, and 5, and mean particle sizes for late Holocene eolian sands. Particle size data are plotted geographically for the first time here, but are from Muhs and others (1996).

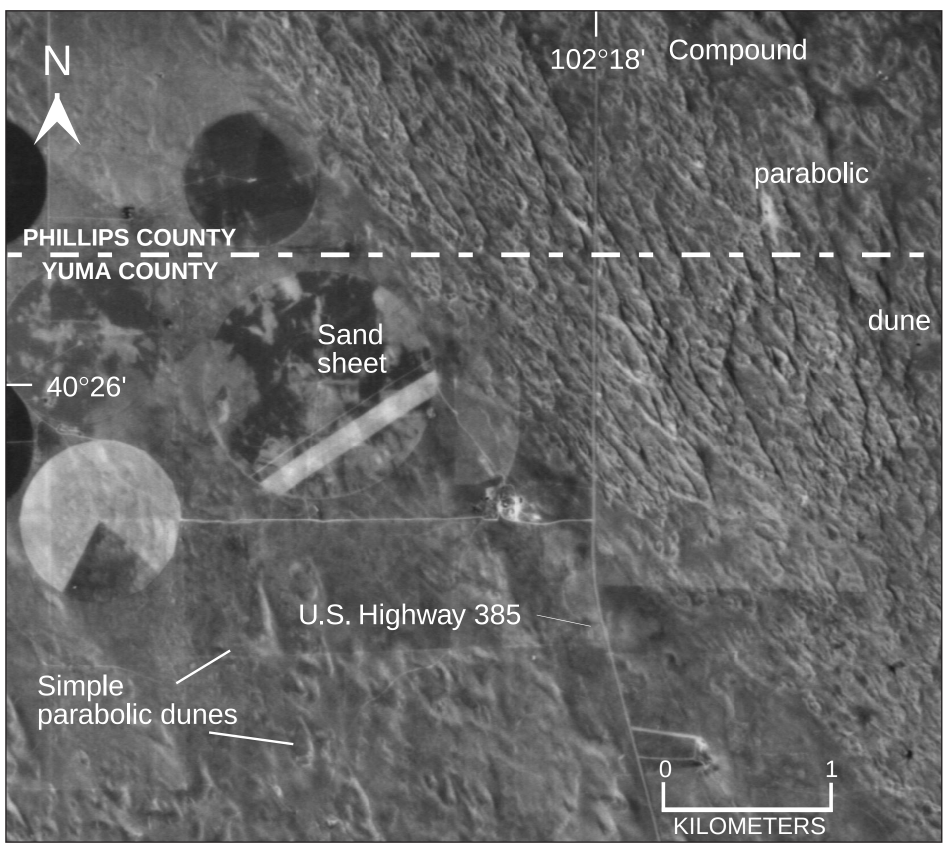


Figure 6. Aerial photograph of a portion of the northern part of the Wray dune field (STOP 2), showing simple and compound parabolic dunes and eolian sheet sands. Note erosion of sheet sands since last cultivation of area in circular field.

## STOP 3: STRATIGRAPHY AND SEDIMENTOLOGY OF A PARABOLIC DUNE NEAR WRAY, COLORADO

Having seen the geomorphology of the Wray dune field at its northern end, at the next stop we will examine the stratigraphy and sedimentary structures in the left arm (looking downwind) of a parabolic dune in a quarry exposure immediately north of the North Fork of the Republican River, near the town of Wray, Colorado (Figs. 7, 8). The lowermost unit is a medium-to-fine sand that is thinly bedded and has apparent gentle (5-7°) dips to the east. About 3 m of this eolian sand is exposed and it is characterized by prominent clay lamellae (also called "clay bands," cf., Gile, 1985 and "dissipation structures," cf. Ahlbrandt and Fryberger, 1980) in the lower part of the exposure. At least nine bands, 0.5-1.5 cm thick, can be found over an ~80-cm depth zone. The lowermost eolian unit has a minimally developed A/AC/C profile, but one which is more than a meter thick. Organic matter content in this soil is extremely low and not much different from the underlying sand (Fig. 7), but it is easily identifiable by its darker colors (10YR 5/2, dry), the presence of roots, and abundant krotovina (infilled rodent-sized burrows). Its combination of minimal horizon development (i.e., lack of B horizon formation) but relatively great thickness suggests that the final stages of deposition of eolian sand were slow, such that pedogenesis kept pace with sedimentation. The contact between the buried soil and the younger eolian unit above it is unusually sharp, suggesting that the soil profile, thick as it is, has been truncated by later deflation.

The middle eolian unit is light brown or light yellowish brown (10YR 6/3, 6/4, dry) sand about 5 m thick and also has beds that dip gently (7-10°) to the east (Fig. 7). Particle size is extremely variable and ranges from coarse to very fine sand. "Pin stripe" or "wood grain" ripple strata 1-3 mm thick are common, particularly in the upper part of the unit. Secondary structures are also common. Root casts are visible in the upper beds

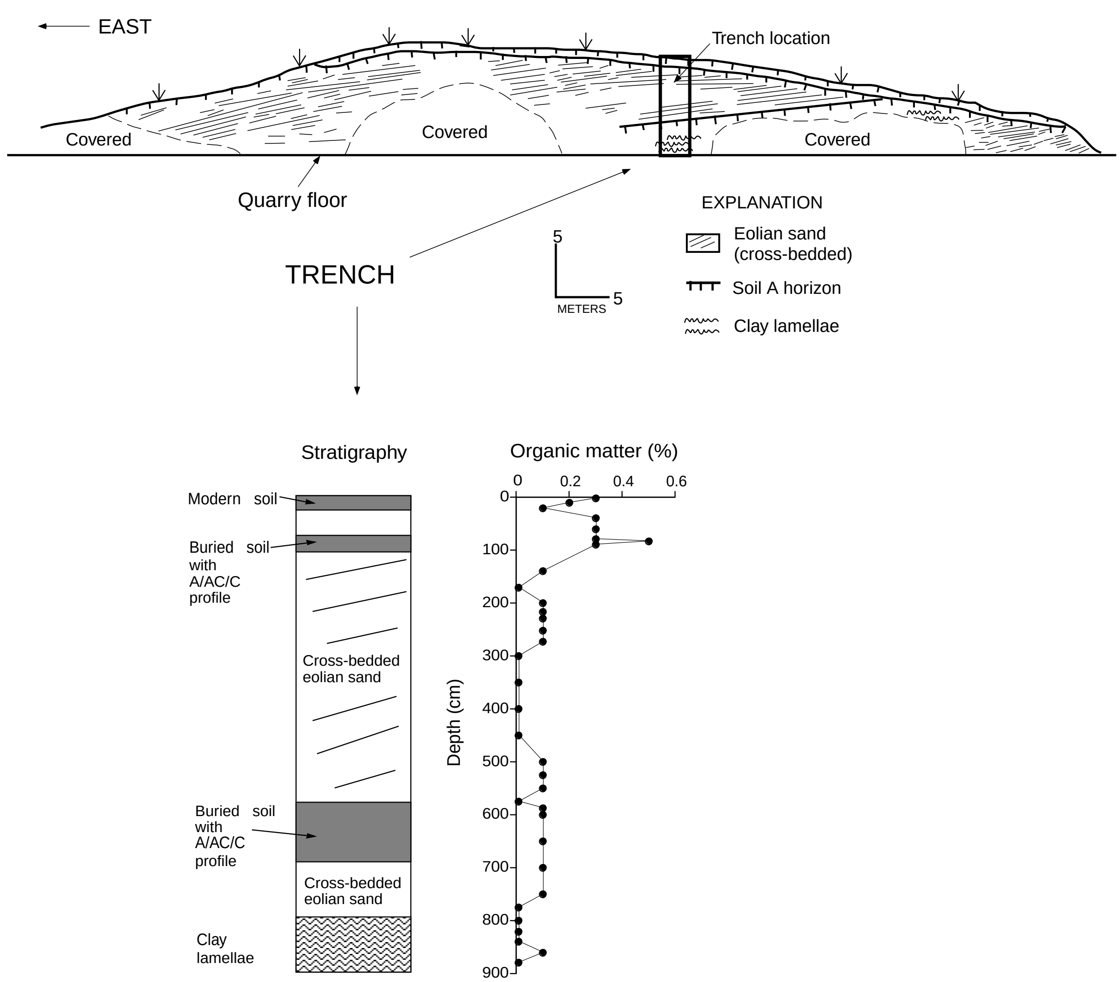


Figure 7. Upper: stratigraphy and structures of the left arm of a parabolic dune exposed in a quarry at Wray, Colorado (STOP 3), and location of trench shown in detail. Lower: detail of trench stratigraphy shown above, and organic matter content (done by the Walkley-Black method) of eolian sand and soils shown as a function of depth. Previously unpublished data of the authors.

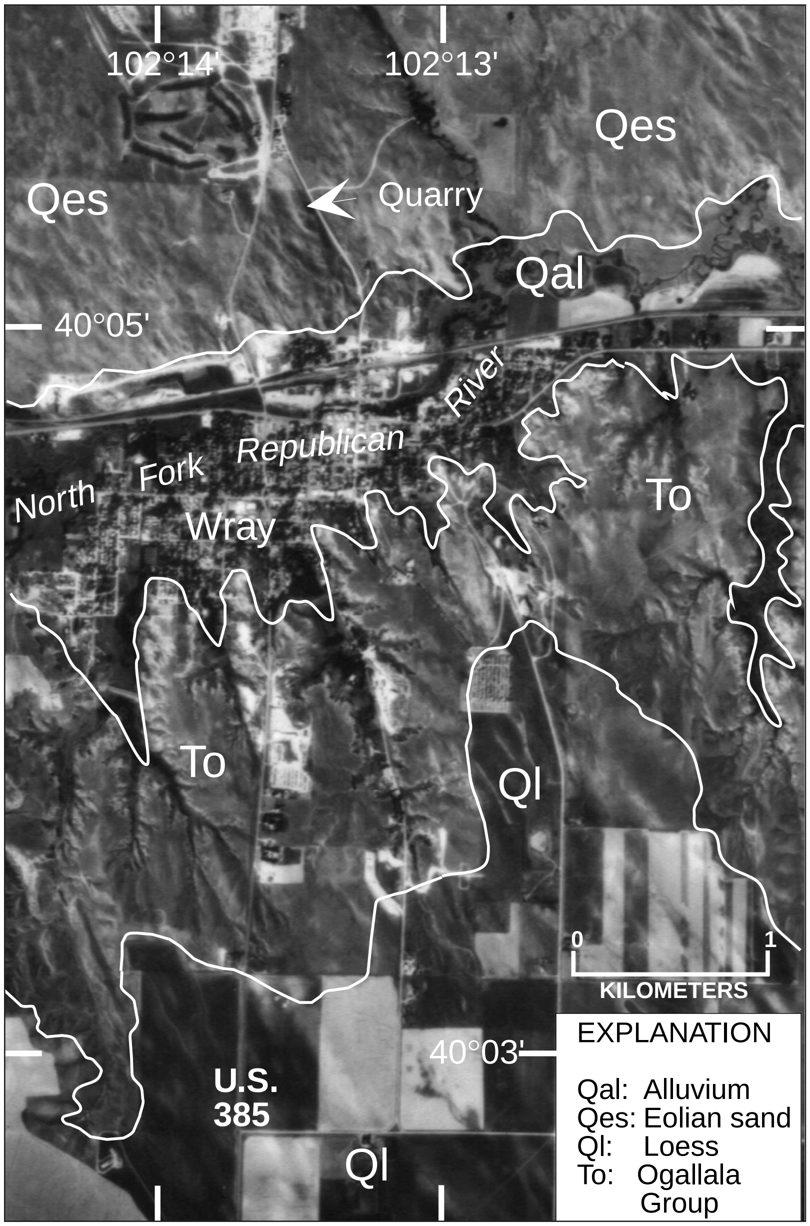


Figure 8. Aerial photograph of the Wray, Colorado area with surficial geology superimposed and location of quarry (STOP 3). Previously unpublished geologic mapping by D.R. Muhs.

and suggest that some vegetation was present at the time of sedimentation. In addition, possible bison hoof prints (cf. Loope, 1986) can be seen about 0.5 m above the lower paleosol and possibly in the some of the finer-grained strata upsection. The middle eolian unit is capped by a thin (10-15 cm), dark (10YR 5/2, dry) paleosol with an A/AC/C profile that has higher organic matter content than the underlying sand (Fig. 7). A meter or less of very young eolian sand occurs above the thin paleosol and could be historic, because only the simplest of A horizons has developed in the upper 5 cm or so.

The stratigraphy and degree of soil development in this dune suggests that all three eolian units were probably deposited in the late Holocene. The lowermost paleosol closely resembles those in a similar stratigraphic position in the Fort Morgan dune field that have radiocarbon ages of ~800 to 1400 yr BP (Madole, 1994, 1995). The thick, lower paleosol, and the presence of root casts in primary structures above it, suggest that vegetation colonization with soil development and sedimentation were competing processes. Despite this, the volume of sand that was transported in late Holocene time is considerable, given the thickness seen in section here and the size of the dune itself.

South of Wray, the North Fork of the Republican River has cut magnificent cliffs into Ogallala Group rocks, underlain by White River Group rocks (Fig. 8). These cliffs, which are visible to the south of STOP 3, are mantled with loess but little or no eolian sand. Dune sand of the Wray dune field apparently did not cross the North Fork of the Republican River. However, loess thickness here is considerable, and is the subject of the next stop.

## STOP 4: LOESS STRATIGRAPHY AT BEECHER ISLAND, COLORADO

Loess in northeastern Colorado is the westernmost part of an almost continuous loess blanket in the North American midcontinent. Loess is distributed widely but discontinuously to the southeast of the South Platte River, with only isolated occurrences to the north of this major drainage (Figs. 1 and 9). The thickest loess we have observed in eastern Colorado is about 12 m. Thicknesses of 2-5 m are more typical, and in the northeasternmost part of the state, thin loess occurs in a patchy distribution on the surface of Ogallala Group bedrock.

Mean particle sizes of loess in northeastern Colorado vary from fine silt in the western part of the area to coarse silt in the eastern part of the area (Fig. 9). This eastward coarsening is in part a function of decreasing clay content to the east, which is as high as 30-40% in the western part of the region and less than 10% near Julesburg (Muhs and others, 1999). The latter workers suggested that much of the clay in eastern Colorado loess may have been eroded from high-clay bedrock units such as the Pierre Shale, which are exposed more widely in the west than they are in the east, and where deflation hollows have been mapped (Colton, 1978). This clay may have been transported as silt-sized clay aggregates. Silt-sized grains in eastern Colorado loess require other sources, which we discuss below.

A roadcut near Beecher Island, Colorado (Fig. 9), has one of the thickest exposures of loess yet reported for eastern Colorado. Buried soils can be found in this section and are identifiable on the basis of morphology, organic matter maxima, $CaCO_3$ minima, and clay maxima (Fig. 10). At this locality, grey-green calcareous clays of unknown origin (not visible in the roadcut, but accessible by augering) are overlain by eolian (?) sands and silts in which a strongly expressed buried soil

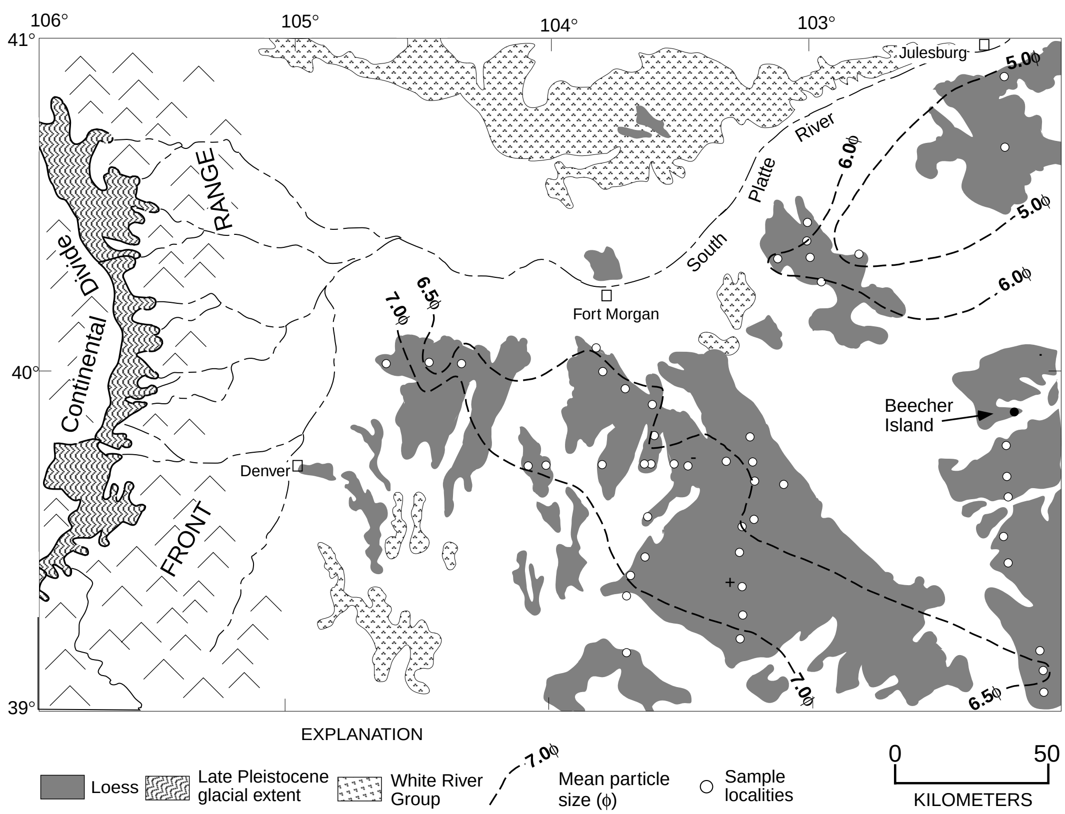


Figure 9. Distribution and mean particle size data for Peoria loess of northeastern Colorado, distribution of rocks of the White River Group, and extent of last-glacial (Pinedale) glaciers on the east side of the Continental Divide. Particle size data are from Muhs and others (1999), but are plotted geographically for the first time here; plus/minus symbols indicate anomalous samples. Distribution of White River Group rocks from Scott (1978); extent of Pinedale glaciers from Madole and others (1998).

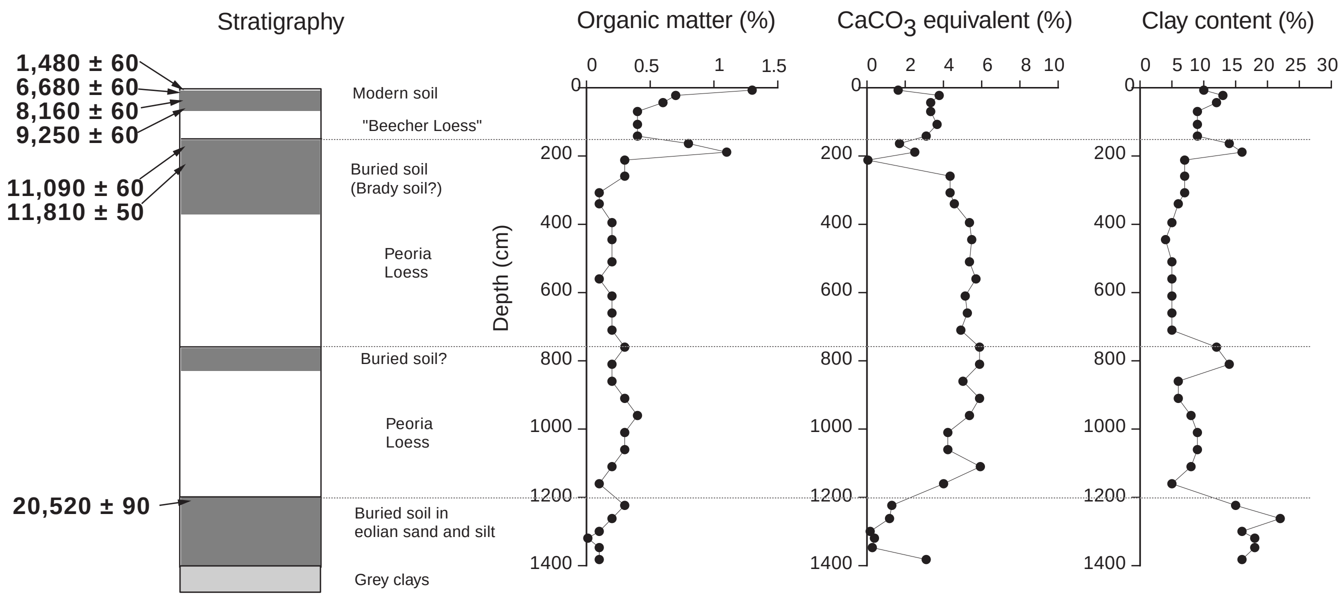


Figure 10. Stratigraphy, AMS radiocarbon ages, organic matter content, $CaCO_3$-equivalent content, and clay content as a function of depth in the loess section at Beecher Island, Colorado (STOP 4). From Muhs and others (1999).

developed. This buried soil has a subangular-to-angular blocky structure with well-expressed clay films, and up to 22% clay in the Bt horizon. Approximately 10 m of what is interpreted to be Peoria Loess overlies this buried soil, although a possible thin buried soil, with strong, coarse prismatic structure and 12-14% clay, is found at a depth of ~8 m. Between a depth of ~1.5 m and 3.5 m, there is a thick, buried soil with an A1/A2/AB/Bw1/Bw2/C profile (Fig. 10) that may be equivalent to the Brady soil of Nebraska and Kansas. This buried soil is in turn overlain by stratified eolian silt and sand with a modern soil, characterized by an A/Bw1/Bw2/C profile, in its upper part. Land snails (*Succinea grosvenori* Lea) can be found in both the upper part of the Peoria Loess and in the younger loess above the uppermost buried soil.

AMS radiocarbon dating of carefully extracted humic acids from paleosols, following the methods given in Abbott and Stafford (1996) and reported by Muhs and others (1999), provides a chronology of loess deposition at Beecher Island (Fig. 10). Humic acids from the upper part of the lowermost buried soil give a radiocarbon age of 20,520 ± 90 $^{14}C$ yr BP, and those from the A1 and A2 horizons of the buried soil (between 1.5 and 3.5 m depth) give ages of 11,090 ± 60 and 11,810 ± 50 $^{14}C$ yr BP (~13,000 and ~13,700 cal yr BP), respectively. Collectively, the radiocarbon ages indicate that Peoria loess deposition occurred between about 20,000 and 12,000 $^{14}C$ yr BP. AMS radiocarbon age determinations were also made on humic acids from the A, Bw1, Bw2, and upper C horizons of the modern soil, in order to provide a minimum age for the youngest, sandy loess at this locality. These ages are consistently younger up through the modern soil profile, and suggest that the youngest loess was deposited between about 11,000 and 9,000 $^{14}C$ yr BP (~13,000 and 10,000 cal yr BP). We had originally assumed that this younger loess was the Bignell Loess of Nebraska and Kansas (seen at the type locality, STOP 6), but the radiocarbon age of the deepest part of the modern soil suggests either that it is slightly older or that deposition of Bignell Loess was time-transgressive.

The complex origin of silts in loess of eastern Colorado, alluded to earlier, can be demonstrated by examination of Pb-isotopic data from Peoria Loess at Beecher Island. One likely source of silt-sized particles during the last glacial period in Colorado is glaciogenic silt derived from alpine glaciers of Pinedale age in the Front Range. Rock flour from Front Range glaciers likely would have been carried by the South Platte River and its tributaries to the Great Plains and deposited in what are now sediments of the Qt1-Broadway-Kersey terrace. Such an origin was proposed for western Great Plains loess by Bryan (1945), Frye and Leonard (1951), Swineford and Frye (1951), and Pye and others (1995). A less obvious source of loess is the White River Group of Eocene-Oligocene age (Fig. 9), which is rich in silt-sized particles. These two sources can be distinguished from one another because the Pb-isotopic compositions of K-feldspars are distinctly different (Aleinikoff and others, 1999). K-feldspars in modern South Platte River silts, derived from Precambrian crystalline rocks of the Front Range (ages of 1.0, 1.4, and 1.7 Ga), have $^{206}Pb/^{204}Pb$ values of 17.0 to 17.7, whereas silt-sized K-feldspars of the White River Group (age of volcanism is ~34 Ma) have $^{206}Pb/^{204}Pb$ values that are much more radiogenic, and range from 18.1 to 19.6. In addition, U-Pb ages of silt-sized zircons from these two sediment groups are distinctly different (Proterozoic vs. Tertiary).

At Beecher Island, K-feldspars from Peoria Loess have Pb-

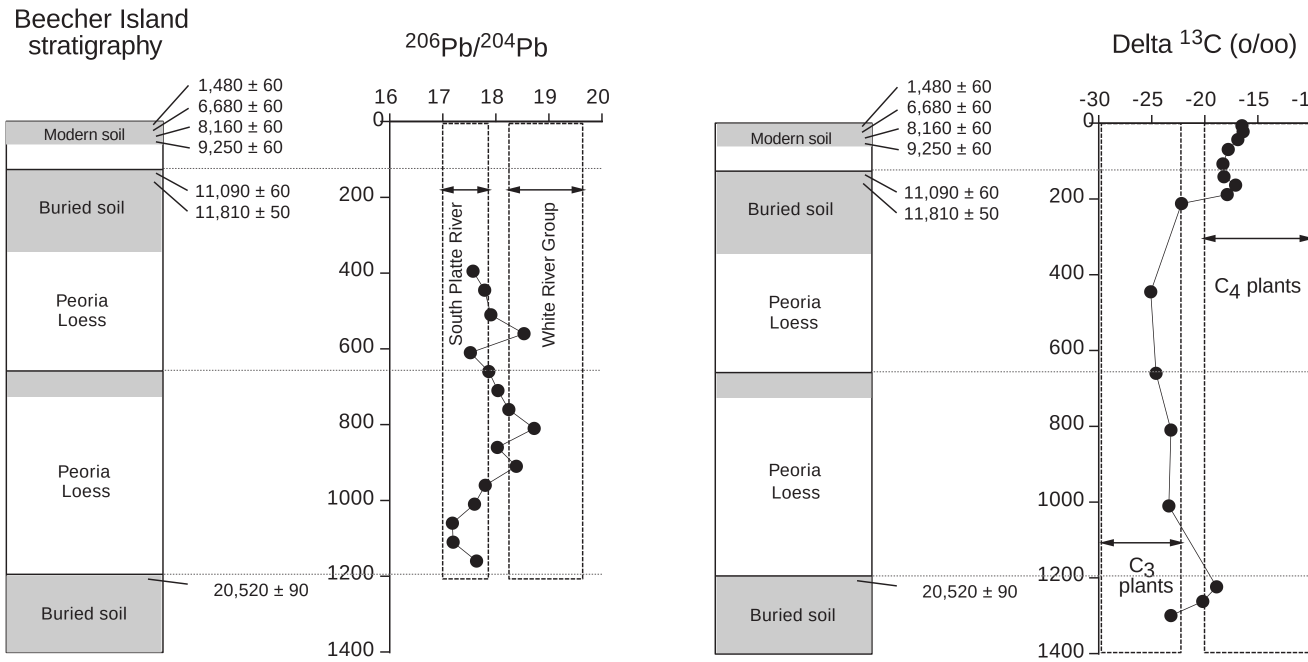


Figure 11. Pb isotopic composition of K-feldspars, ranges of these values in modern South Platte River sediments and sediments of the White River Group, and carbon isotopic composition of soil organic matter and detrital organic matter in loess at Beecher Island. Pb isotope data from Aleinikoff and others (1999); carbon isotope data from Muhs and others (1999).

isotopic compositions that span the entire range of ratios measured in both possible sources (Fig. 11). This indicates that loess was derived from both silt from the South Platte River (and therefore could be glaciogenic) and silt from the White River Group. The isotopic ratios vary systematically within the loess section at Beecher Island and within another section found farther west, near Last Chance, Colorado (Aleinikoff and others, 1999). In both sections, loess just above the ~20,000 yr old paleosol has Pb isotopic compositions within the range of values measured in K-feldspars from the South Platte River. Upsection, the ratios increase (to values corresponding to those found in the White River Group) and decrease twice. Aleinikoff and others (1999) suggested that under relatively cold conditions within the last (Pinedale) glacial period, valley glaciers of the Front Range advanced and glaciogenic silt derived from Proterozoic crystalline rocks was entrained within the ice, with relatively little silt released to streams during short summer ablation periods. Concomitantly, vegetation, which even now is fairly sparse on rocks of the White River Group, decreased, thereby destabilizing the volcaniclastic sediments and making them more susceptible to erosion. Although there may have been some eolian erosion directly from sediments of the White River Group, it is more likely that reduced vegetation cover would allow greater fluvial erosion and delivery to tributaries of the South Platte River. As conditions became warmer during the Pinedale glacial period, vegetation cover increased in eastern Colorado, including that on sediments of the White River Group, and Pinedale glaciers of the Front Range receded, generating greater amounts of outwash. The Pb-isotopic data from Beecher Island suggest the occurrence of an earlier cycle of warming and cooling between peak late Pinedale glaciation and final deglaciation (Fig. 11).

Carbon isotopic composition of organic matter in buried soils and loess show changes in dominant vegetation types over the last glacial-interglacial cycle (Fig. 11). The lowermost buried soil, dated at ~20,000 $^{14}C$ yr BP, has $\delta^{13}C$ values of -19‰ to -23‰, indicating a probable mix of $C_3$ and $C_4$ vegetation. In the overlying Peoria Loess, detrital organic matter has $\delta^{13}C$ values ranging from about -23‰ to -25‰, indicating a dominance of $C_3$ vegetation at the time of loess fall. However, the A horizon of the ~11,000 $^{14}C$ yr BP buried soil and the loess and modern soil above it have $\delta^{13}C$ values of -16‰ to -17‰, indicating a dominance of $C_4$ vegetation. Overall, the carbon isotopic compositions indicate a dominance of $C_3$ vegetation from about 20,000-12,000 $^{14}C$ yr BP, and a dominance of $C_4$ vegetation after ~12,000 $^{14}C$ yr BP. Two vertebrate faunal localities near Beecher Island studied by Graham (1981) provide additional details of the environment at the time of loess deposition. Fossil ungulates and rodents recovered from Peoria Loess at these localities led Graham (1981) to conclude that grassland was the predominant vegetation in eastern Colorado during the time of Peoria Loess deposition. The combined faunal and carbon isotope data indicate that, during the last glacial period, eastern Colorado supported a cool grassland, perhaps similar to that found today in southern Canada, Montana, or the Dakotas.

## STOP 5: HOOVER BLOWOUT, WRAY DUNE FIELD, NEBRASKA

In the eastern part of the Wray dune field a deep deflation hollow called the Hoover blowout (Madole, 1995; Muhs and others, 1997a) exposes ~8 m of eolian sand within the nose area of a northwest-trending compound parabolic dune (Fig. 12). The eolian sands contain two buried soils with A/AC/C profiles and are underlain by pond or lacustrine sands, silts, and clays. Fossil mollusk shells (*Stagnicola palustris*) from the paludal sediments gives a radiocarbon age of 13,130 ± 295 $^{14}$C yr B.P. (DIC-2198) and humus from the lowermost buried soil gives a radiocarbon age of 7870 ± 240 $^{14}$C yr B.P. (DIC-2270) (Madole, 1995). Impressions that we interpret as bison hoof marks (cf. Loope, 1986) can be found in the middle eolian unit and transect primary bedding structures. At a depth of ~4.5 m, within the middle eolian unit, we recovered a long bone fragment attributable to Bison. The bone gives a carboxyl age of 290 ± 60 $^{14}$C yr B.P. (490-0 cal yr B.P.) and a collagen age of 360 ± 60 $^{14}$C yr B.P. (515-290 cal yr B.P.) (Muhs and others, 1997a). From all of these radiocarbon ages, we infer that eolian sedimentation began sometime after ~13,000 $^{14}$C yr B.P. and was episodic, based on the presence of the two buried soils. The two most recent episodes of eolian sedimentation began sometime after ~7870 $^{14}$C yr B.P., and one episode apparently occurred within the past ~500 cal yr. Thus, the data indicate that the eastern Wray dune field, like the Fort Morgan dune field and the Nebraska Sand Hills, has been active in the past 1000 cal yr, and could even have been active in historic time.

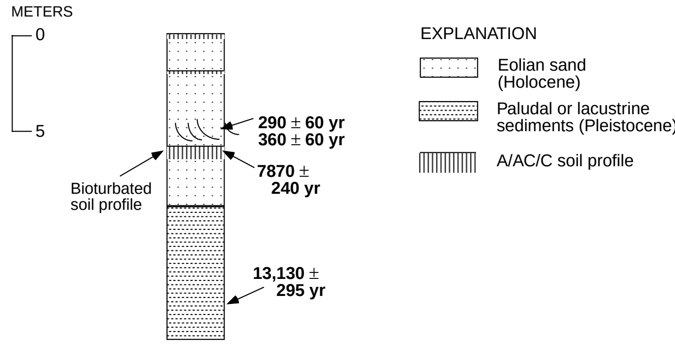


Figure 12. Stratigraphy and radiocarbon ages in eolian sands at the Hoover blowout (STOP 5). Data from Madole (1995) and Muhs and others (1997b).

## STOP 6: BIGNELL HILL, NEBRASKA

Bignell Hill, Nebraska, is one of the most famous loess sections in the midcontinent of North America. The locality has been studied by Quaternary geologists for more than 50 years, and results of this work have been reported by Schultz and Stout (1945), Frye and Leonard (1951), Dreeszen (1970), Johnson (1993), Feng and others (1994), Maat and Johnson (1996) and Muhs and others (1999). It is the type locality for the Brady soil, developed in the uppermost Peoria Loess, and the Bignell Loess, which overlies the Brady soil (Schultz and Stout, 1945).

Bignell Hill contains what may be the thickest (>50 m) late Quaternary loess section in North America. At the northern end of the roadcut, a reddish-brown paleosol can sometimes be seen cropping out on the road surface. Above this, and visible in the roadcut itself (to about 4.5 m above road level), is a well developed brown (10YR 5/4, dry) paleosol with a Btk/Bt/C profile almost a meter thick that has developed in eolian (?) silty sands (Fig. 13). Both paleosols are undated, but they may correlate to some part of the Sangamon interglacial period. Overlying these two paleosols is the Gilman Canyon Formation, consisting of two loesses that both contain minimally developed but organic-rich soil A horizons that are darker (10YR 5/2 and 4/3, dry) than the underlying and overlying loess (10YR 6/3, 6/4, dry). The Gilman Canyon Formation contains abundant evidence of burrowing in the form of krotovina. Overlying the Gilman Canyon Formation is one of the thickest (~48 m) exposures of Peoria Loess yet described from North America. This loess has distinct laminae, particularly in the upper 20 m of the unit, which Feng and others (1994) suggest may represent annual layers. At the top of the Peoria Loess, there is a distinctive paleosol (10YR 3/2 dry colors in the A horizon) with an A/Bw1/Bw2/BC/C profile called the Brady soil, first reported by Schultz and Stout (1945). The Brady soil is overlain by about 2 m of Bignell Loess (2.5Y 6/3, dry) and a modern soil.

There have been many geochronological studies of the loesses at Bignell Hill (Fig. 13). Maat and Johnson (1996) reported an age of 30,970 ± 780 $^{14}$C yr BP on organic matter from the oldest of the two paleosols from the Gilman Canyon Formation. Muhs and others (1999) used the humic acid extraction method of Abbott and Stafford (1996) in an attempt to minimize contamination from both older (reworked) and younger carbon, and obtained an age of 40,600 ± 1100 $^{14}$C yr BP for the same paleosol. Feng and others (1994), Maat and Johnson (1996) and Muhs and others (1999) also dated the uppermost Gilman Canyon paleosol. The earlier studies reported an age of 28,130 ± 610 $^{14}$C yr BP; Muhs and others reported an age of 30,770 ± 210 $^{14}$C yr BP. Maat and Johnson (1996) reported concordant total-bleach and partial-bleach TL ages of 28,300 ± 5100 cal yr BP and 28,200 ± 8400 cal yr BP, respectively, for the Gilman Canyon Formation. All age estimates for the uppermost paleosol are in good agreement with one another within analytical uncertainties and possible radiocarbon-to-calendar year calibration uncertainties (which are unknown for this time period). However, there is a significant difference (almost 10,000 yr) in the ages reported for the lowermost Gilman Canyon paleosol. Part of the difference may be due to the portion of the paleosol that was sampled in the two studies as well as the extraction methods themselves.

There are fewer ages of the Peoria Loess at Bignell Hill. Feng and others (1994) found a charcoalized twig of Picea glauca from the upper part of the Peoria Loess, ~3.5 m below the base of the Brady soil, and reported an age of 11,880 ± 90

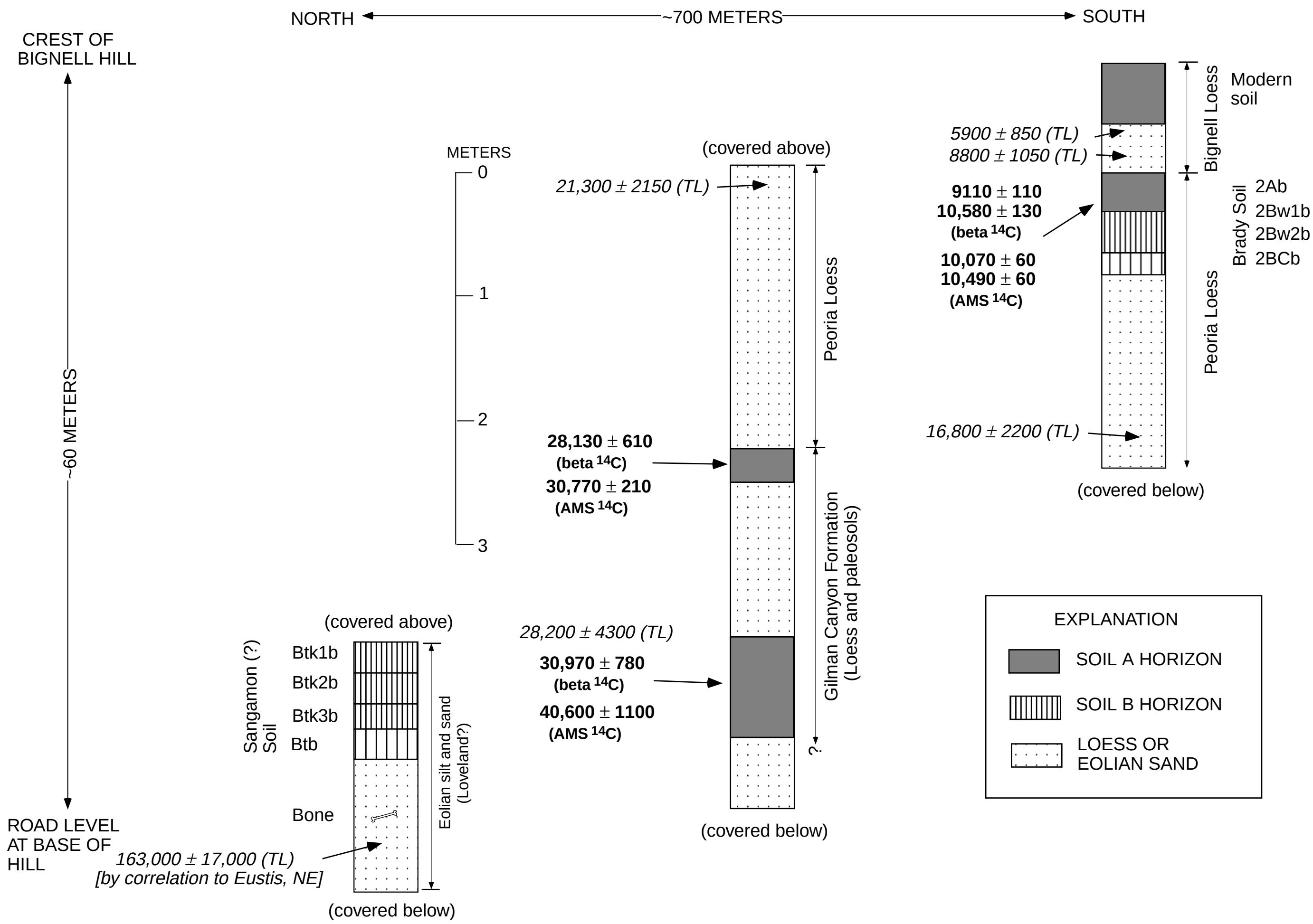


Figure 13. Stratigraphy, soils and age estimates of deposits at Bignell Hill, Nebraska (STOP 6). Stratigraphy from Maat and Johnson (1996), soils data from the present authors, and age data from Maat and Johnson (1996) and Muhs and others (1999).

$^{14}$C yr BP. This radiocarbon age translates to a calendar-year age of ~13,600-14,200 yr BP. Maat and Johnson (1996) reported basal Peoria Loess TL ages of 21,700 ± 3200 (total bleach) and 20,900 ± 3000 cal yr BP (partial bleach). These same workers also reported TL ages for the uppermost Peoria Loess (~4 m above Feng and others' sample depth) of 17,900 ± 2500 (total bleach) and 12,400 ± 5000 (partial bleach) cal yr BP.

The Brady soil and overlying Bignell Loess have received the most attention for geochronological studies at Bignell Hill, starting with Dreeszen (1970). Johnson (1993) and Maat and Johnson (1996) redated the upper and lower parts of the Brady soil and obtained ages of 9110 ± 110 $^{14}$C yr BP and 10,580 ± 130 $^{14}$C yr BP, respectively. Muhs and others (1999) also redated the upper and lower Brady soil and obtained ages of 10,070 ± 60 and 10,490 ± 60 $^{14}$C yr BP (11,007-12,054 and 12,176-12,590 cal yr BP), respectively. Maat and Johnson (1996) reported TL ages of 8300 ± 1200 (total bleach) and 10,400 ± 2300 (partial bleach) cal yr BP for the lower Bignell Loess and 6100 ± 900 (total bleach) and 4500 ± 2200 cal yr BP (partial bleach) for the upper Bignell Loess. The TL ages obtained by both methods are concordant and stratigraphically consistent with the radiocarbon ages of the underlying Brady soil. Collectively, all the age estimates indicate that the Bignell Loess was deposited during the Holocene, in agreement with TL ages of this unit reported from elsewhere in Nebraska (Pye and others, 1995).

The radiocarbon and TL results from Bignell Hill indicate that while there is general agreement in the timing of last-glacial loess deposition in various parts of the midcontinent, there are differences in detail. Loess deposition in Iowa, Illinois, and areas eastward was, to a great extent, a function of source sediment availability from the Laurentide ice sheet via the Mississippi and Missouri Rivers, and the timing of loess deposition closely followed the history of movement of the ice sheet (see Grimley and others, 1998). Based on the ages from Bignell Hill, Peoria Loess deposition in western Nebraska could have begun earlier (sometime just after ~30,000 $^{14}$C yr BP) and continued until significantly later (~10,500 $^{14}$C yr BP) than in areas to the east (Ruhe, 1983; Curry and Follmer, 1992; Grimley and others, 1998). Furthermore, eolian silt of Holocene age, such as the Bignell Loess, has not been reported from areas east of the Missouri River. These observations suggest that sources of loess in the central Great Plains are unrelated to the specific dynamics of the Laurentide ice sheet, a conclusion supported by recent isotopic data found in Aleinikoff and others (1998, 1999).

## STOP 7: ELDRED CAMP, NEBRASKA SAND HILLS

The 50,000 km$^2$ Nebraska Sand Hills area (Fig. 1) is the largest dune field (active or stabilized) in North America. Diverse stabilized eolian landforms are found in the Nebraska Sand Hills (Smith, 1965; Ahlbrandt and Fryberger, 1980; Swinehart, 1990). Barchanoid-ridge dunes are as much as 50 km long and are easily visible on Landsat imagery. Barchans, linear dunes, parabolic dunes, dome-like dunes, and sand sheets all are found in the Nebraska Sand Hills. Measurement of slip-faces of stabilized dunes and high-angle foreset bed dip azimuths indicate that paleowinds originated from the north-west and north, similar to modern wind regimes (Warren, 1976; Ahlbrandt and Fryberger, 1980). In the next two stops we will examine the field evidence for at least two episodes of eolian sand movement that were dramatic enough to dam drainages within the western Nebraska Sand Hills (Figs. 1 and 14).

At Eldred Camp (STOP 7), the steep groundwater gradient between Crescent Lake and the springs at the head of Blue Creek (1:115) contrasts with the 1:450 gradient of Blue Creek and the 1:1100 slope of the groundwater table north of the lake (Fig. 14). Crescent Lake, the southernmost of hundreds of lakes in the western Sand Hills, lies 3 km north of and 25 m higher

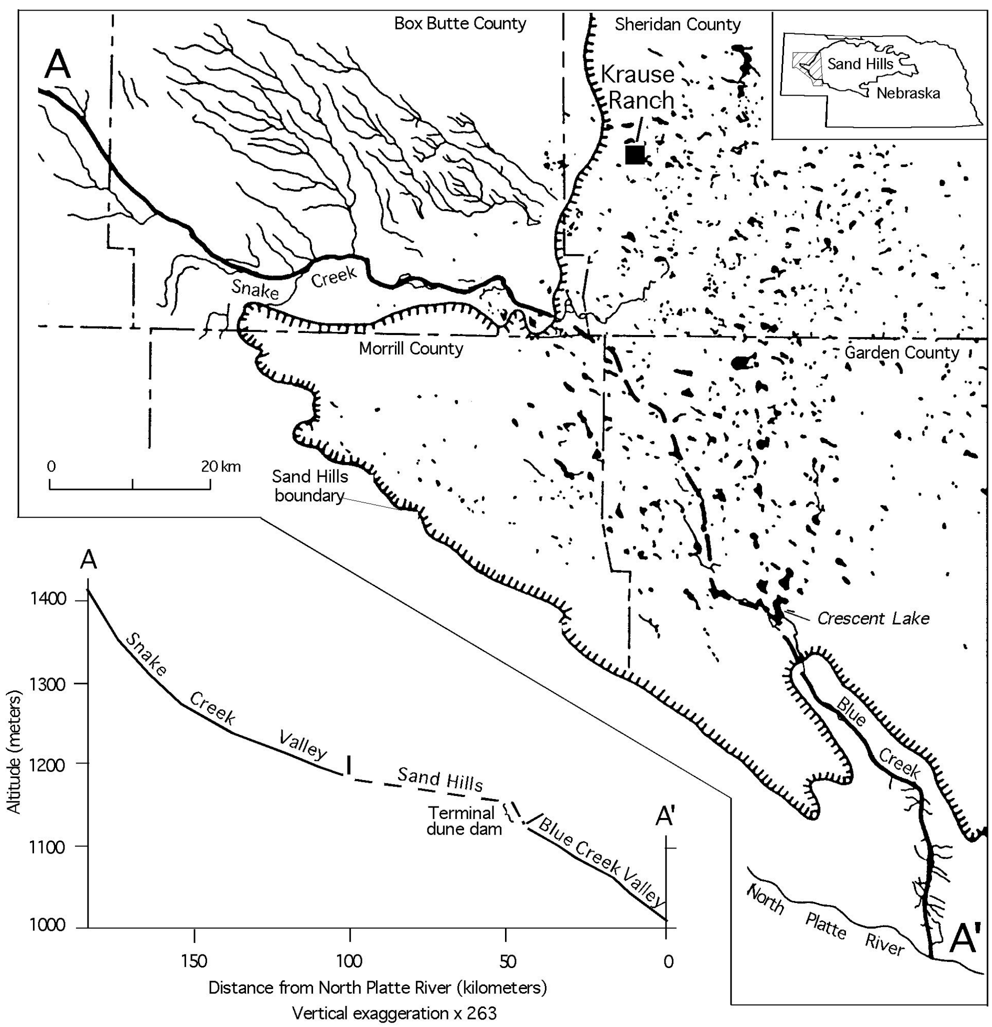


Figure 14. Western Nebraska Sand Hills showing lakes and present Snake Creek and Blue Creek drainages. Section A-A' was drawn down the primary Snake Creek and Blue Creek valleys and across the Sand Hills (dashed line) along a probable trace of the dune dammed valley. The position of the buried valley is well known only in the area of Crescent and Swan lakes (Fig. 15).

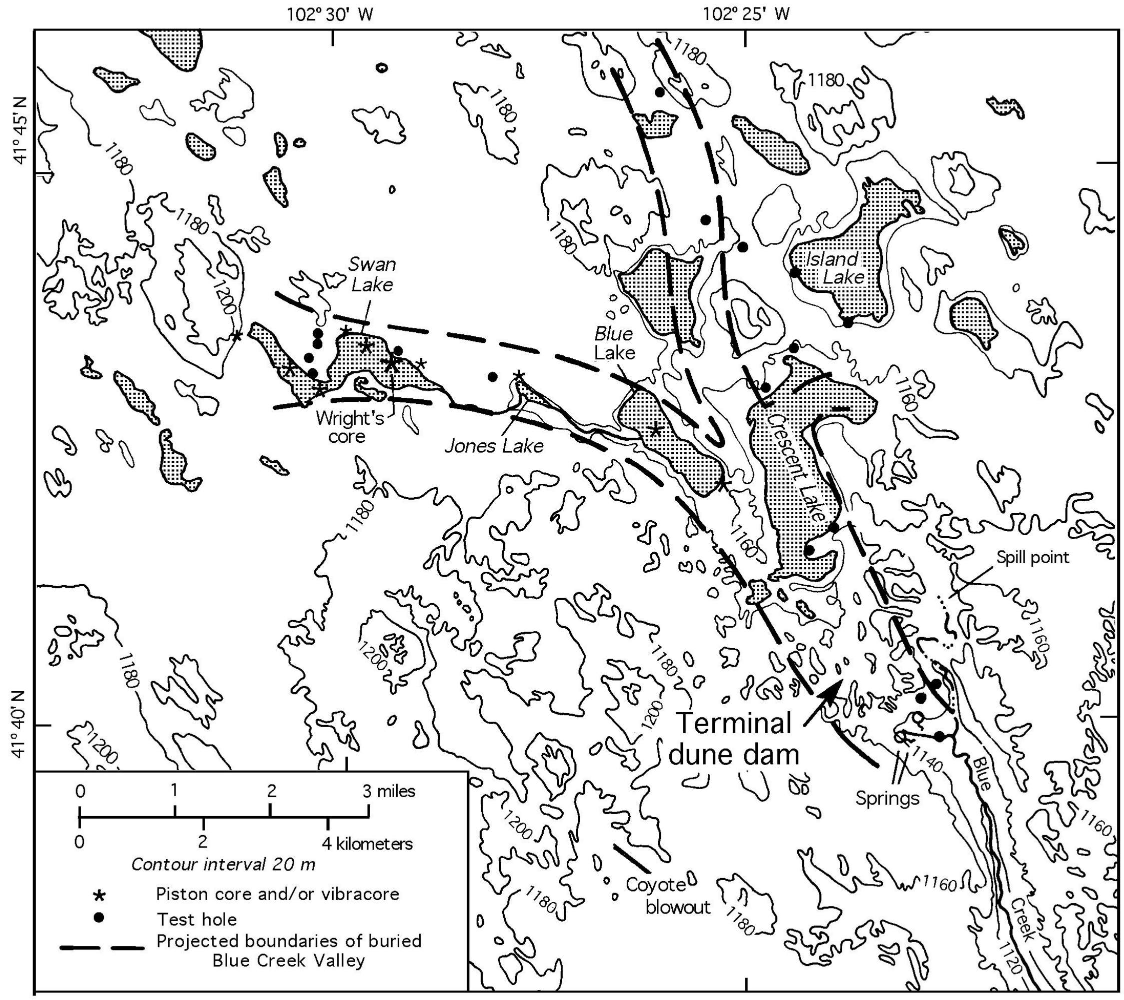


Figure 15. At the southwest margin of the Sand Hills, springs at the head of Blue Creek (lower right) emerge from a sand dam, the southernmost of numerous sand bodies that block this major paleovalley. Configuration of the buried valley system is based on vibracores, rotary drilled test holes, outcrops, long axes of lakes, and the dimensions of Blue Creek valley. The high water table behind the dam creates lakes in interdune positions. None of the lakes presently has natural surface water inlets or outlets. Note position of the intermittent stream course just east of the terminal dune dam that acted as a spillway when the level of Crescent Lake was about 2 m higher.

than the springs at the head of Blue Creek (Figs. 14 and 15). This portion of the valley of Blue Creek was not formed by headward sapping; dune sand clearly dammed a through-going extension of Snake Creek that occupied this valley (Loope and others, 1995; Mason and others, 1997).

Although dune sand is presently mounded across the position of the paleovalley, an abandoned spill point for the system lies on Ogallala Group bedrock about 1 km east of the sand-filled paleovalley, at an elevation 2 m higher than the adjacent lake surface (Fig. 15). A sinuous channel below this spill point that is cut into Ogallala Group rocks testifies to overflow during a former lake high stand. Catastrophic drainage of the lake immediately behind the dam did not take place because the spillover is floored by partially lithified material, not dune sand. Seepage through the dune dam must have taken place at a sufficient rate to prevent massive overflow and deep entrenchment.

Re-establishment of Blue Creek as a through-going stream would require that many dune dams be removed. If the southernmost dune dam were overtopped, then the southernmost lake or cluster of a lakes would drain, and the head of Blue Creek would migrate upstream several kilometers to the base of the next dam. A counterintuitive conclusion of our work is that a positive change in the water budget could cause a drastic drop in the water table throughout the catchment area due to overtopping and removal of the dune dams.

## STOP 8: BLUE CREEK DUNE DAM, NEBRASKA SAND HILLS

Across broad areas of the Sand Hills, interdune surfaces intersect the groundwater table, forming extensive wetlands. Paradoxically, the part of the sand sea with the least precipita-

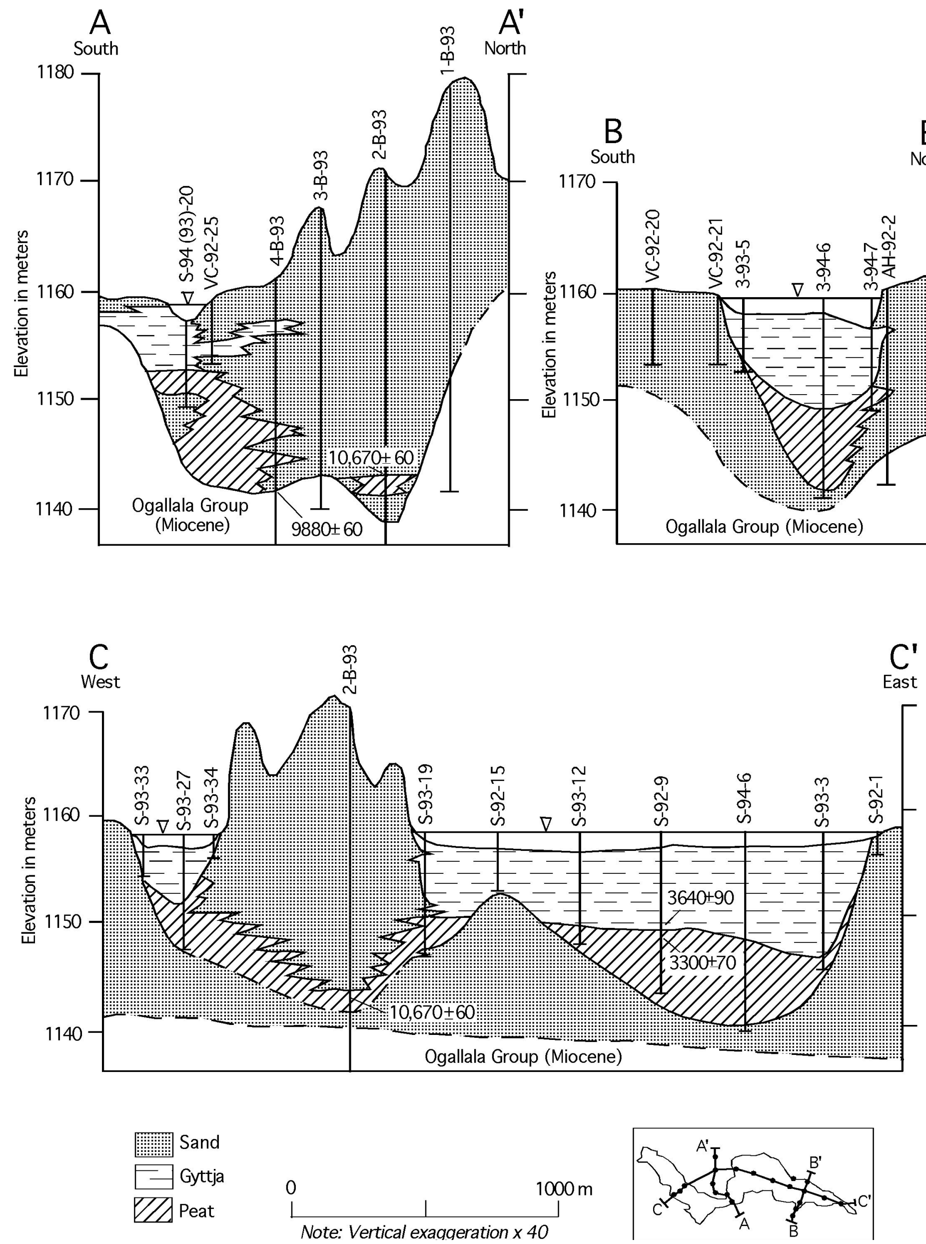


Figure 16. Swan Lake basin cross-sections constructed from lake piston cores and probes, vibracores (VC), rotary test holes (1 to 4-B-93), and an auger test hole (AH), and radiocarbon ages. After Mason and others (1997).

tion—the western Sand Hills—contains the greatest number of lakes. Estimates of the total number of interdune lakes in the region vary from 1500 to 2500.

Eastward-flowing water courses on the dune- and lake-free tableland west of the Sand Hills disappear when they reach the western margin of the sand sea (Fig. 14). At the southern edge of the Sand Hills, Blue Creek, a perennial, spring-fed stream that occupies a valley cut into Miocene bedrock, emerges from dune sand (Figs. 14 and 15). This "terminal dune dam" is the southernmost of scores of valley-blocking sand bodies; modern lakes and Holocene lacustrine and wetland sediments occupy the parts of the paleovalleys not filled by dune sand.

Based on rotary drilling, vibracores, piston cores, outcrops, the orientation of the long axes of lakes, and the dimensions of Blue Creek's valley, we project a 1200 to 2200-m-wide, 30-m-deep buried fluvial paleovalley between Swan Lake and the head of Blue Creek, 8 km southeastward (Fig. 15). Sixteen radiocarbon ages on various materials in cores from three different lake basins (Figs. 16, 17, 18) establish a chronological framework for correlation and interpretation (Loope and others, 1995; Mason and others, 1997).

At least two distinct episodes of blockage are required to explain the history of sedimentation in Swan, Blue and Crescent Lakes. Recently, we obtained five radiocarbon ages from

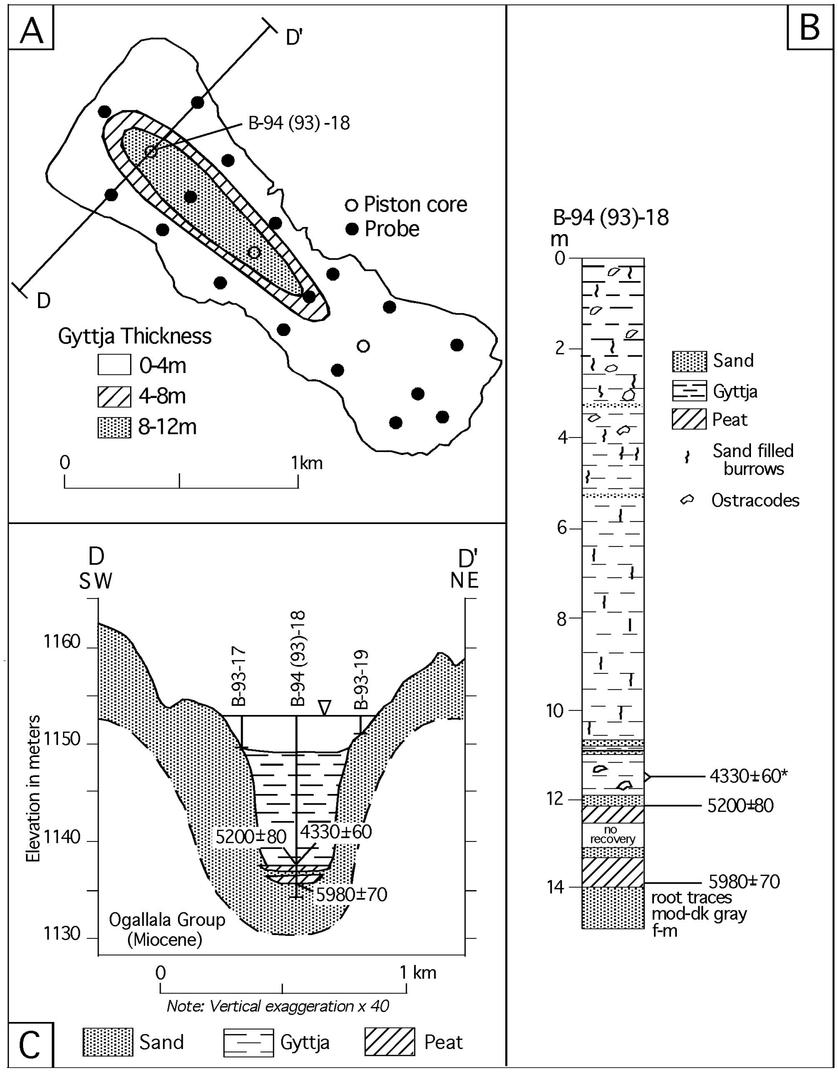


Figure 17. Blue Lake basin gyttja thickness and stratigraphy. (A) Gyttja thickness. (B) Composite section of two piston cores taken at the same location. The AMS radiocarbon age of 4330 yr BP (marked by asterisk) was obtained from plant fragments in gyttja. (C) Cross-section D-D'. After Mason and others (1997).

basal lacustrine/wetland sediments in cores taken near Krause ranch (Fig. 14), about 60 km north-northwest of Crescent Lake (Sweeney and others, 1998). These sites were part of the ancestral Blue Creek drainage basin prior to the latest Wisconsin blockage. The radiocarbon ages range from 12,160 to 12,360 $^{14}$C yr BP and suggest dune blockage of the drainage just prior to 12,000 radiocarbon years ago. It appears that a significant arid interval well after the last glacial maximum led to a major episode of dune blockage in the Blue Creek drainage. We postulate that the blockage at Swan Lake formed prior to 10,600 $^{14}$C yr BP and possibly as early as 12,000 $^{14}$C yr BP since the 10,600 $^{14}$C age comes from the upper part of a 2-m-thick peat. The dune dams that created Blue and Crescent basins were emplaced during remobilization of this part of the dune field during the mid-Holocene (about 6000 $^{14}$C yr BP).

The second blockage event most likely reflects the middle Holocene period of minimum effective moisture in the Great Basin and Colorado Plateau summarized by Thompson and others (1993). Lake and pollen data from the north-central U.S. also support a middle Holocene period of minimum effective moisture (Webb and others, 1993). Holliday (1989) gave evidence for a prolonged drought and widespread eolian activity on the Southern High Plains between 6500 and 4500 $^{14}$C yr BP, while Stokes and Swinehart (1997) presented direct evidence of middle Holocene eolian activity in the northern Sand Hills based on an optically stimulated luminescence age of ~5700 cal yr BP.

We postulate that as the Sand Hills area became increasingly arid and vegetation became sparse, dune sand would have covered a larger and larger proportion of the interfluves. Infiltration of precipitation into the permeable dune sand would have reduced the magnitude of runoff events, and as rainfall diminished further, surface flow would have ceased when subsurface flow through the unconsolidated, highly per-

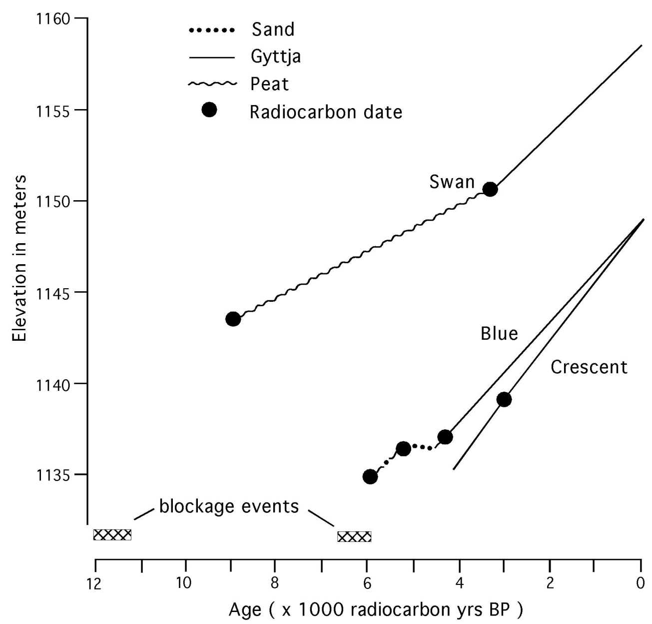


Figure 18. Estimated sedimentation rates and water level changes in the three lake basins based on radiocarbon ages and sedimentary facies distribution from piston cores. After emplacement of the initial dune dam 10,500 to 12,000 $^{14}$C yr BP, peat accumulation in Swan Lake basin kept pace with the water table rise. Following a second episode of dune blockage south of Blue and Crescent Lakes, the water table gradient was lowered through the Swan basin dam and led to the formation of an open-water lake. Lacustrine deposition in Swan Lake basin lagged about 1000 years behind that of Blue and Crescent basins.

meable alluvium on the floor of the channel (derived in large part from the Broadwater Formation) and the underlying poorly consolidated sediments of the High Plains aquifer could accommodate all the water input to the drainage basin. Given the high hydraulic conductivity of the materials that underlie the Sand Hills, it seems likely that prolonged drought would have eventually eliminated surface flow in many reaches of the streams of the region. The sand-carrying capacity of the wind is reduced in the lee of obstacles to air flow such as cliffs and stream banks (Greeley and Iversen, 1985). This tendency probably led to preferential deposition of eolian sand in valleys. Large masses of dune sand then moved into blocking positions on the dry floors of streams that lacked the potential to generate flash floods.

Thick peat deposits beneath the floor of Swan Lake (Fig. 16) indicate that a slow, steady rise of the water level in the basin began in early Holocene time and continued for over 6000 years. Gyttja above the peat indicates that this slow rise was followed by a more rapid rise that, at about 3700 years ago, created the open-water conditions present today (Fig. 18). Rather than interpreting the 17-m rise of the wetland surface during Holocene time as a result of a long-term trend toward a wetter regional climate (Wright and others, 1985), our working hypothesis for the history of the Swan Lake basin calls upon locally controlled changes in the rate of groundwater flow and requires two arid episodes during which dunes blocked dry stream courses.

Marshes initially formed in a dune-blocked (western) arm of the paleovalley system prior to about 12,000 $^{14}$C yr ago and thereafter steadily aggraded for more than 6000 years. Ingram (1982) showed that, in areas of much higher precipitation, peat accumulation impedes the drainage of rain water, resulting in the growth of groundwater mounds and domed mires. We suggest that in the Sand Hills, peat accumulation did not passively keep pace with the rise of the water table, but rather that the deposition of this impermeable material actively contributed to the rise of the water table by progressively impeding the down-gradient movement of water through the valley-blocking dune sand. About 6000 $^{14}$C yr BP a second blockage event took place down flow from Swan Lake and led to emplacement of the dune dams southeast of the present Blue and Crescent Lakes. Crescent and Blue Lakes formed about 4000 years ago (Figs. 18 and 19) and Swan Lake about 1000 years later. The lag in lake sedimentation reflects the time needed for the regional water table to rise about 11 m between the Blue and Swan Lake basins.

After 11,000 years of sedimentation and water table rise, Swan Lake is now nearly full of impermeable sediment. The difference in elevation between its surface and the water table beneath the dunes to the north and south (transverse to the paleovalley) is very small. This situation, combined with the low elevation of the now-abandoned spill point on the east side of Crescent Lake (Fig. 15), indicates that the water table rise cannot be sustained very far into the future. The present extent of wetlands in the study area is therefore near both the maximum for Holocene time and the maximum possible.

We postulate that the paleovalleys control the wide variation of lake water chemistry in the western Sand Hills: the fresh-water lakes at the southern margin of the sand sea, where the gradient of the groundwater table is steep (Winter, 1976), are flow-through lakes that lose salts to the springs at the head of Blue Creek. Only short segments of thick, sand-filled paleovalleys lie between these lakes and the discharge point. We interpret the alkaline, saline lakes to the north, however, as discharge points for closed, local groundwater flow systems (Toth, 1962; Gosselin and others, 1994): these lakes cannot lose salts to the regional aquifer because they occupy an area with a low hydraulic gradient and because groundwater flow through the thinner valley fills in the north is impeded by the thick, impermeable mud deposited in lakes to the south.

The capillary fringe in sand is thin and sedimentary structures show that upland vegetation was very sparse when dunes were active; the diminished stream flow that allowed blockage cannot be explained by an increase in evapotranspiration. The blockages of valleys by dune sand—like the giant bedforms themselves—testify to long periods during which precipitation was much less than at present.

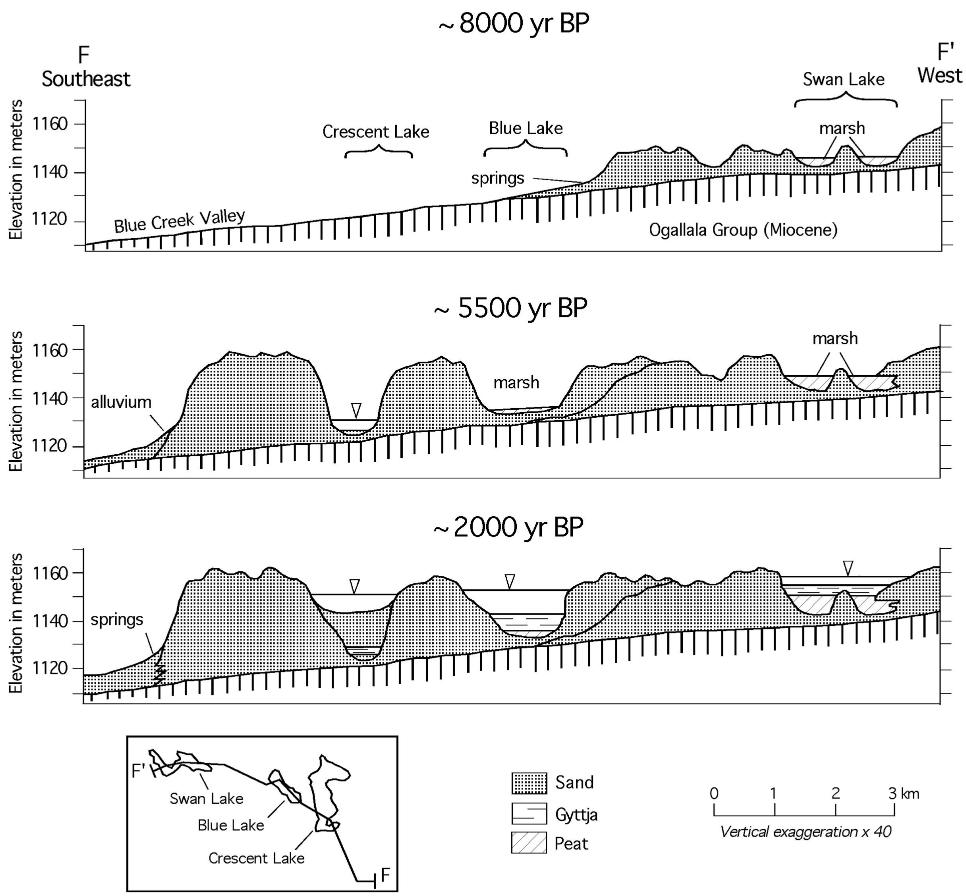


Figure 19. Schematic cross sections (F-F') along the west axis of Blue Creek Valley from Swan Lake through Crescent Lake at three times in the Holocene. Dune sand blocked the valley between the present Swan Lake and Blue Lakes sometime during 10,500 to 12,000 years ago. (8000 yr BP)—The regional water table was high enough for about 2 m of peat to accumulate in two separate areas of Swan Lake basin. Downstream from the dune dam, Blue Creek Valley is unblocked. A second episode of blockage probably occurred prior to 5500 years ago and created Blue and Crescent basins. Some dune sand advanced into parts of Swan Lake basin but was stopped from reaching the central parts of the basin by riparian vegetation. (5500 yr BP)—Between 5000 and 6000 years ago less than two m of peat accumulated in marshes in Blue basin while lacustrine sediments accumulated in the deeper of the two Crescent Lake basins. Lacustrine sedimentation of gyttja began about 4000 years ago in Blue and Crescent basins followed by the partial collapse of the deep, southern Crescent basin about 1000 years later. Open water conditions lagged almost 1000 years behind the two lakes down gradient. (2000 yr BP)—All three lake basins reach their present configuration (after Mason and others, 1997).

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