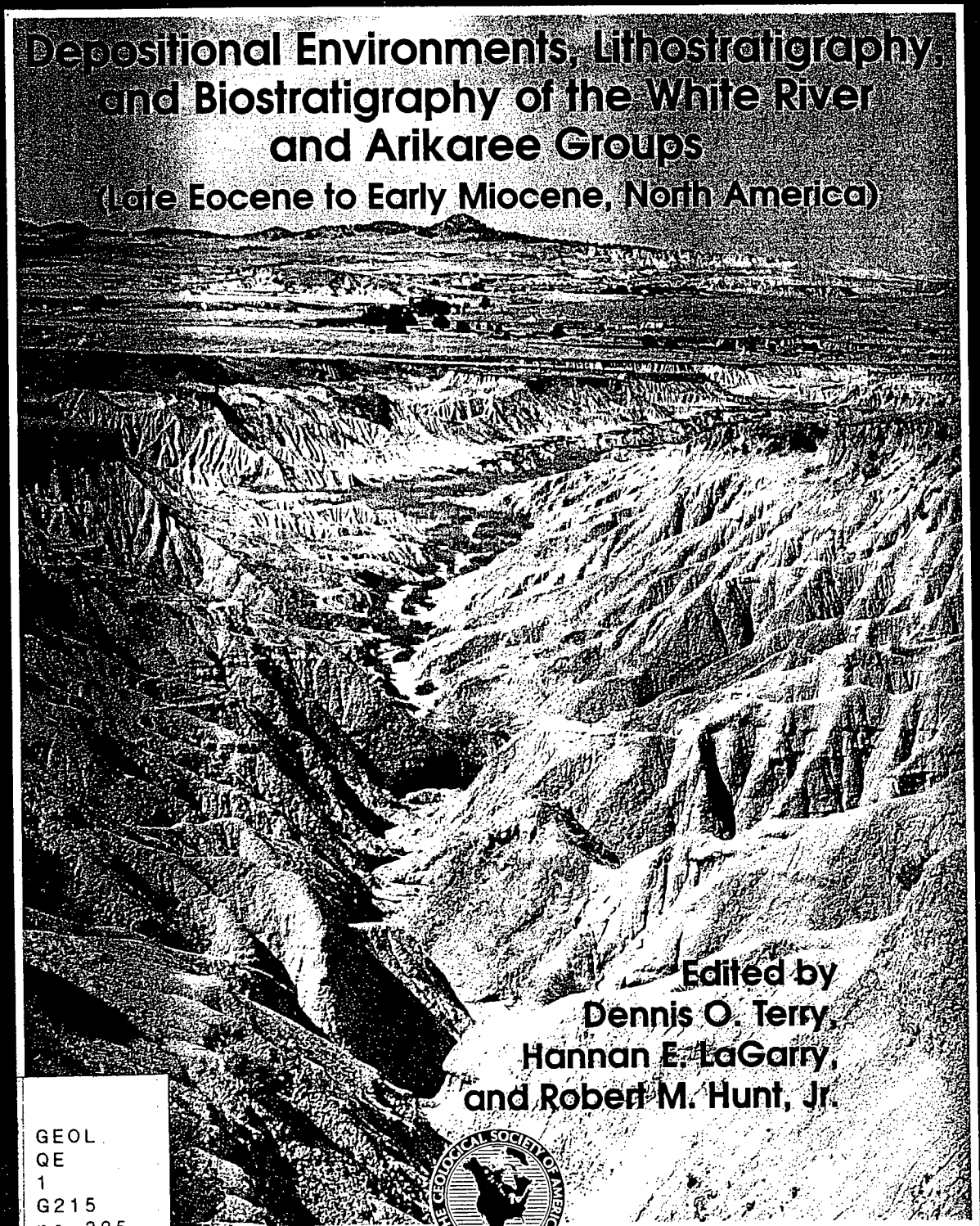
Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups

(Late Eocene to Early Miocene, North America)



Edited by
Dennis O. Terry,
Hannan E. LaGarry,
and Robert M. Hunt, Jr.

THE GEOLOGICAL SOCIETY OF AMERICA · 1888 ·

SPECIAL PAPER

Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America)

Edited by

Dennis O. Terry
Department of Geology
University of Nebraska–Lincoln
Lincoln, Nebraska 68588-0340

Hannan E. LaGarry
Conservation and Survey Division (IANR)
and State Museum
University of Nebraska–Lincoln
Lincoln, Nebraska 68588-0514

and

Robert M. Hunt, Jr.
Division of Vertebrate Paleontology
University of Nebraska–Lincoln
Lincoln, Nebraska 68588

THE GEOLOGICAL SOCIETY OF AMERICA · 1888 ·

SPECIAL PAPER

325

1998

Published by The Geological Society of America, Inc.
3300 Penrose Place, P.O. Box 9140, Boulder, Colorado 80301

Printed in U.S.A.

GSA Books Science Editor Abhijit Basu

Library of Congress Cataloging-in-Publication Data
Depositional environments, lithostratigraphy, and biostratigraphy of the White River and Arikaree Groups : (Late Eocene to early Miocene, North America) / edited by Dennis O. Terry, Hannan E. LaGarry and Robert M. Hunt Jr.
p. cm. -- (Special paper ; 325)
Includes bibliographical references and index.
ISBN 0-8137-2325-6
1. Geology, Stratigraphic--Tertiary. 2. Paleomagnetism. 3. Volcanism. 4. Sedimentation and deposition--North America. 5. Sedimentation and deposition--Rocky Mountain Region. 6. White River Group. I. Terry, Dennis O., 1965- . II. LaGarry, Hannan E., 1962- . III. Hunt, Robert M., Jr. IV. Series: Special papers (Geological Society of America) ; 325.
QE691.D45 1998
551.7'8'097--dc21 98-26926
CIP

Cover: Fine-grained volcaniclastic rocks of the White River and Arikaree Groups initially studied by N. H. Darton (1899) at the Pine Ridge escarpment in northwestern Nebraska (Roundtop locality).

10 9 8 7 6 5 4 3 2 1

Contents

Preface

The White River and Arikaree Groups comprise an enormous volume of chiefly fine-grained volcaniclastic sediments that accumulated in the North American midcontinent and adjacent Rocky Mountain basins from the late Eocene to the early Miocene (~37 to 19 Ma). White River–Arikaree rocks occur primarily in Nebraska, South and North Dakota, Wyoming, Montana, and Colorado. They have been the focus of geologic and paleontologic investigations since the mid-19th century because of the rich fossil mammalian record preserved in these largely diagenetically unaltered sediments.

The succession of mammal faunas contained within the White River–Arikaree sequence has become a nonmarine biochronologic standard for North America for the late Eocene to early Miocene interval: the Chadronian, Orellan, Whitneyan, and Arikareean land-mammal "ages" (Wood et al., 1941) are based on faunal aggregates from the White River and Arikaree units of the central and northern Great Plains. Nonmarine rocks of comparable age elsewhere in North America are routinely correlated to this established chronology (Tedford et al., 1987). In addition to fossil mammals, White River–Arikaree rocks contain numerous vitric tuffs datable by radiometric methods. These fine-grained ash-rich sediments have also been amenable to paleomagnetic sampling, allowing correlation to the Geomagnetic Polarity Time Scale (Cande and Kent, 1992; Berggren et al., 1995).

Because the abundant fossil mammals were often the focus of earlier studies, initial classification of these rocks emphasized only the most evident lithic distinctions. The rocks were subdivided using the more common mammals (e.g., *Titanotherium* beds) characteristic of various stratigraphic levels. Detailed lithic descriptions, the identification of bounding unconformities and useful marker horizons, and an appreciation of the varied depositional environments lagged behind the steady acquisition of well-preserved fossil mammals. To some degree this was due to a perceived lithic homogeneity of the White River and Arikaree subunits, seemingly monotonous in their fine-grained texture, uniform color, apparent lack of bedding, and primary sedimentary structures.

Lithologic uniformity and the blanketing regional geometry indicate an eolian origin for most White River and Arikaree rocks in the Great Plains and Rocky Mountains. An abundance of volcanic glass, especially prominent in the White River but also common in the Arikaree, together with a smaller fraction of pyrogenic crystals, documents a pyroclastic origin from volcanic centers to the west.

Deposition of the White River and Arikaree Groups across the North American midcontinent resulted from the interaction of climate, sediment input, and regional tectonics. These sediments are best approached using the tools of a variety of disciplines: lithostratigraphy, sedimentology, paleopedology, tephrochronology, biostratigraphy, and magnetostratigraphy. New data and interpretations are presented in this volume, and correlative rocks and faunas of the same age elsewhere in North America are evaluated and reviewed.

Larson and Evanoff interpret the source area for the great volumes of ash that characterize the White River Group. They provide revised $^{40}Ar/^{39}Ar$ single-crystal laser-fusion dates for key marker tuffs in the Douglas and Flagstaff Rim areas of Wyoming, propose tephrostratigraphic correlations to other key outcrops on the Great Plains, provide an in-depth analysis of potential source areas for these volcaniclastic sediments based on geochemistry of mineral phases within the tuffs, and correlate tuff ages within the White River Group to volcanic centers to the west. Their results suggest that the majority of tuffs were derived from explosive volcanism in Utah and Nevada.

Evans and Welzenbach present the first detailed descriptions of lacustrine limestones within the Chadron Formation of the White River Group. While the vertebrate fossils of the White River have received much attention over the last century, inverte-

brates and plants from the limestones are of interest in environmental interpretation but remain poorly known. Their work is a step forward in understanding this aspect of the biota. Although volumetrically minor when compared with the whole of the White River Group, these limestones may also contribute to an understanding of regional tectonic processes. They are essentially devoid of siliciclastics, which is intriguing since they are intimately interstratified with flood-plain sediments. According to the authors, these limestones are the result of carbonate-rich springs that formed in response to the uplift of the Black Hills, and subsequent development of an eastward-sloping regional groundwater gradient.

Prothero and Whittlesey evaluate the nature of litho- and biostratigraphic completeness of the White River Group across the Great Plains at selected localities. They propose a preliminary biostratigraphic zonation for part of the White River Group in South Dakota and Nebraska, testable by biostratigraphers employing more detailed and exacting stratigraphic sampling in these classic White River strata.

The chapters by Terry and LaGarry revise and redescribe the White River Group in northwestern Nebraska and provide a new model of lithostratigraphic classification in the central Great Plains. They examine the relationship of prominent paleosol horizons at the base of the White River Group, discuss the distribution and lithic features of the Chamberlain Pass Formation in Nebraska and South Dakota, and explore the regional differences in lithology and stratigraphic relations of the various members of the Chadron Formation (Ahearn, Crazy Johnson, Peanut Peak, and Big Cottonwood Creek Members) between South Dakota and Nebraska.

Terry and LaGarry argue that the newly recognized Big Cottonwood Creek Member in Nebraska represents a period of time not represented by sediments in the Big Badlands of South Dakota. The contact of the Chadron and Brule Formations in the Big Badlands, specifically in the Sage Creek–Pinnacles area of Badlands National Park, is marked by a significant unconformity. This contact is recognized by a lithologic change and marked by erosional relief at the boundary; it has also been tied to magnetostratigraphic and biostratigraphic markers. Prothero and Whittlesey (this volume) suggest an unconformity of at least 400,000 yr exists at the Chadron-Brule contact in the Big Badlands of South Dakota. The Big Cottonwood Creek Member in northwestern Nebraska is believed to represent at least a portion of this missing interval of time. These revisions have resulted in the repositioning of the lithostratigraphic boundary between the Chadron and Brule Formations in northwestern Nebraska, and a similar adjustment of intraformational boundaries within the Brule Formation.

MacFadden and Hunt present the results of paleomagnetic sampling of the upper Arikaree Group in the type area of N. H. Darton (1899) in northwestern Nebraska. The study reveals as the principal paleomagnetic datum in the upper Arikaree a long, normal magnetozone, correlated to Chron C6n (Cande and Kent, 1992), that characterizes the type Harrison Formation of Hatcher (1902). Resting unconformably on the Harrison Formation are the Upper Harrison beds (Peterson, 1907, 1909), whose lower part is also usually normally magnetized and which is calibrated by a tuff dated at 19.2 ± 0.5 Ma (FT-zircon) (Hunt et al., 1983). Paleomagnetic sampling was extended down-section to near the base of the Arikaree Group in northwestern Nebraska. The lower Arikaree Group is characterized by a long reversed interval overlain by a long normal magnetozone, but the lack of datable tuffs in the local section prevented accurate calibration. Correlation to a lithically similar sequence in the southern panhandle of Nebraska suggests, however, that these rocks probably correspond to Chron 9r and 9n, dating this interval to ~28.2 to ~27 Ma.

Sampling was also extended up-section into the lowest unit of the Hemingford Group in the vicinity of Agate National Monument (Runningwater Formation, Cook, 1965). A prominent normal magnetozone comprises an extended interval believed to correlate to Chron C5En, suggesting the lowest part of the Hemingford Group dates to ~18 to ~18.8 Ma. The Arikareean-Hemingfordian boundary is placed according to this evidence at ~18.8 Ma.

Faunas of Arikareean age can be identified across North America and are correlated to the central Great Plains Arikareean succession by their diverse fossil mammals. Albright reports an integrated synthesis of the Arikareean faunas of the Gulf Coastal Plain, chiefly from Florida and Texas-Louisiana; these coastal plain faunas show evident communication with the better known mammalian assemblages of Nebraska, South Dakota, and Wyoming during this late Oligocene–early Miocene interval. However, they retain endemic species that indicate the presence of a unique coastal biotic province at this time in the mid-Cenozoic.

Hoganson, Murphy, and Forsman review the unique geologic setting of North Dakota White River and Arikaree rocks in light of new information on biochronologic age. These typically fine-grained volcaniclastic sediments cap numerous isolated buttes at the northern geographic limit of the White River–Arikaree in the Great Plains. Despite a lack of lateral continuity of outcrop between buttes, a comprehensive picture of the relative age and lithologic character throughout the region has been developed by careful and methodical work.

The tectonically complex Cenozoic basins of southern California contain biogeographically important Arikareean mammals in varied depositional settings. Although fossils are not common, Woodburne reviews and brings new information to bear on the mammalian succession in the Mojave Desert Province, comparing the Arikareean and the succeeding Hemingfordian faunas with those of the midcontinent. New geochronologic information on the stratigraphic sequence in the northern Cady Mountains and the Hector area of the Mojave province, particularly the identification and revised dating of a widespread ignimbrite (Peach Springs Tuff), provides new insight into the timing of this important faunal succession.

Much useful work remains to be accomplished in the study of White River and Arikaree sediments and their correlatives in North America. The opportunity to gain new insights into cli-

mate, tectonics, and biogeographic patterns in North America during the mid-Cenozoic is evident to current researchers. We believe that a comprehensive regional synthesis of this interval is at hand in the years ahead.

REFERENCES CITED

Berggren, W. A., Kent, D. V., Swisher, C. C., III, and Aubry, M. P., 1995, A revised Cenozoic geochronology and chronostratigraphy, *in* Berggren, W. A., Kent, D. V., Aubry, M. P., and Hardenbol, J., eds., Geochronology, time scales and global stratigraphic correlation: Tulsa, Society of Economic Paleontologists and Mineralogists Special Publication 54, p. 129–212.

Cande, S. C., and Kent, D. V., 1992, A new geomagnetic polarity time scale for the late Cretaceous and Cenozoic: Journal of Geophysical Research, v. 97, p. 13917–13951.

Cook, H. J., 1965, Runningwater Formation, middle Miocene of Nebraska: American Museum Novitates, no. 2227, p. 1–8.

Darton, N. H., 1899, Preliminary report on the geology and water resources of Nebraska west of the 103rd meridian: U.S. Geological Survey 19th Annual Report 1897–1898; Pt. 4, Hydrography: U.S. Geological Survey, p. 719–785.

Hatcher, J. B., 1902, Origin of the Oligocene and Miocene deposits of the Great Plains: Proceedings of the American Philosophical Society, v. 41, p. 113–131.

Hunt, R. M., Xue, X. X., and Kaufman, J., 1983, Miocene burrows of extinct beardogs: indication of early denning behavior of large mammalian carnivores: Science, v. 221, p. 364–366.

Peterson, O. A., 1907, The Miocene beds of western Nebraska and eastern Wyoming and their vertebrate faunae: Carnegie Museum Annals, v. 4, no. 1, p. 21–72.

Peterson, O. A., 1909, A revision of the Entelodontidae: Carnegie Museum Memoir 4, no. 3, p. 41–158.

Tedford, R. H. (chairman), and Galusha, T., Skinner, M. F., Taylor, B. E., Fields, R. W., MacDonald, J. R., Rensberger, J. M., Webb, S. D., and Whistler, D. P., 1987, Faunal succession and biochronology of the Arikareean through Hemphillian interval (late Oligocene through earliest Pliocene epochs) in North America, in Woodburne, M. O., ed., Cenozoic Mammals of North America: Berkeley, University of California Press, p. 153–210.

Wood, H. E., II (chairman), Chaney, R. W., Clark, J., Colbert, E. H., Jepson, G. L., Reeside, J. B., Jr., and Stock, C., 1941, Nomenclature and correlation of the North American continental Tertiary: Geological Society of America Bulletin, v. 52, p. 1–48.

Printed in U.S.A.

Geological Society of America
Special Paper 325
1998

Tephrostratigraphy and source of the tuffs of the White River sequence

Edwin E. Larson
Department of Geological Sciences, Campus Box 250, University of Colorado, Boulder, Colorado 80309-0250
Emmett Evanoff
University of Colorado Museum, Campus Box 315, Boulder, Colorado 80309-0315

ABSTRACT

The abundant mudrocks of the upper Eocene to lower Oligocene White River sequence (Great Plains and central Rocky Mountains) are largely composed (~60%) of volcaniclasts. The White River also contains numerous, thin, widespread tuff beds. Because the late Eocene and early Oligocene was a time of widespread volcanism throughout western North America, a localized source for the large volume of tuffaceous material in the sequence has remained unclear. The intercalated tuffs are not only excellent local and regional chronostratigraphic markers, but their mineralogy, geochemistry, and radiometric ages help constrain plausible sources of the abundant volcanic component in the White River sequence.

We have correlated many of the White River tuffs by means of characteristic mineral suites; geochemistry of hornblende, pyroxenes, biotite, feldspar, and hemo-ilmenite; grain-size analysis of heavy minerals; the paleomagnetic remanence directions; single-crystal laser-fusion $^{40}Ar/^{39}Ar$ ages; and their bio- and magnetostratigraphic relations.

The tuffs, which include rhyolitic, rhyodacitic, dacitic, and andesitic compositions, range in age from ~35.5 to 30 Ma. Tuffs older than 31 Ma are mostly rhyolitic to rhyodacitic; those younger than 31 Ma are predominantly dacitic. Crystal sizes in individual tuffs decrease toward the north and east, indicating transport directions from the southwest. These data, when combined with radiometric ages, and after comparison of tuff compositions with those of possible source areas, strongly suggest that the vast majority of tuffs were derived from explosive volcanism in Utah and Nevada. We have identified only one tuff (in northeast Colorado) that appears to have had a Colorado source.

INTRODUCTION

The Oligocene White River sequence of the Great Plains and central Rocky Mountains is an unconformity-bounded succession of terrestrial rocks that is primarily composed of tuffaceous mudstones and siltstones. The sequence also contains minor amounts of fluvial, coarse sandstone, lacustrine limestone and gypsum, and tuff beds. The sequence includes a lower fluvial section (the Chadron Formation of Nebraska and its equivalents) and an upper fluvial-aeolian part (the Brule Formation of Nebraska and its equivalents). One of the most remarkable features of the White River is its vast content of volcaniclastic sediment (Sato and Denson, 1967; Swinehart et al., 1985). The proportion of volcaniclastic sediment increases from east to west and from the base to the top (Wood, 1947; Emry et al., 1987), with the percentage of volcanic fragments becoming >80% in the upper Brule Formation (Swinehart and Rebone, 1984).

The area originally covered by the White River sequence was ~400,000 km^2 (Fig. 1) and its thickness presently ranges from a feather edge to >300 m. In Nebraska, where the White

Larson, E. E., and Evanoff, E., 1998, Tephrostratigraphy and source of the tuffs of the White River sequence, *in* Terry, D.O., Jr., LaGarry, H. E., and Hunt, R. M., Jr., eds., Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America): Boulder, Colorado, Geological Society of America Special Paper 325.

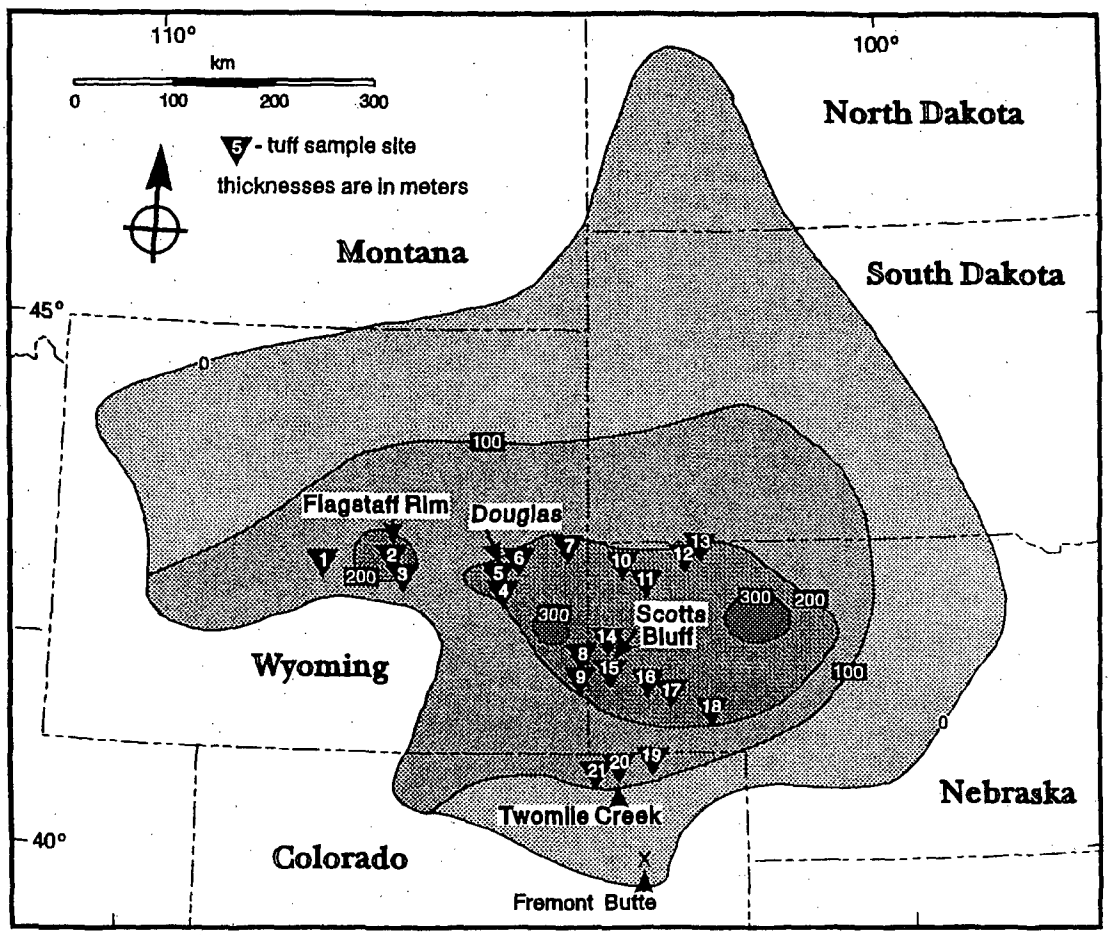


Figure 1. Original distribution and thickness of the White River sequence (including exposures of the White River Formation and Group) as determined from distribution of exposures. Thicknesses are from Love (1960), Denson and Gill (1965), Swinehart et al. (1985), Swinehart and Diffendal (1990), Lillegraven (1993), and Murphy et al. (1995). The distribution in Nebraska includes unpublished data of Harold DeGraw and James Swinehart (written communication, 1995). Numbers refer to localities that are given in Table 1.

River is thickest, volcaniclasts (glass and volcanic minerals) make up ~60% of the volume (Swinehart et al., 1985). Extrapolation of this average value throughout the outcrop area of the White River provides a minimum estimate of the volcaniclastic component of ~25,000 km^3.

A long-standing problem has concerned the source(s) of this immense volume of volcanically derived sediment. Possible Oligocene source areas include Mexico, west Texas, New Mexico, Colorado, the Great Basin, and the Pacific Northwest. To solve this problem, we have carefully examined the tuff beds intercalated within the White River sequence, obtaining mineralogic, chemical, radiometric, grain size, and bio- and magnetostratigraphic data. Much of the data enabled us to correlate the individual ash beds regionally. Many enabled us to constrain potential volcanic source areas.

TUFFS OF THE WHITE RIVER SEQUENCE

Tuffs are lithified volcanic ash beds that were deposited from an ash cloud, downwind from a volcanic eruption (Cas and Wright, 1988). The definition of tuff used herein is equivalent to ash-fall tuff (as contrasted with ash-flow tuff) and includes "fallout tuffs" of volcanologists (Fisher and Schmincke, 1984), and "air-fall" tuffs of sedimentologists (Tucker, 1981). We do not use the broader definition of tuffs that includes all reworked tuffaceous material (fluvial tuffs and aeolian tuffs of Fisher and Schmincke, 1984) because

almost all of the White River mudrocks would be classified as "tuff" by application of that definition.

The tuffs we sampled (Table 1) are volumetrically small, but stratigraphically significant units. Those that can be traced decrease in number, thickness, and lateral persistence from central Wyoming into South Dakota. Inasmuch as volcanic ash is deposited over a few days to a few weeks (Schmincke and van den Bogaard, 1991), tuffs essentially represent chronostratigraphic marker beds. As such, they have been used for detailed stratigraphic positioning of fossil localities, magnetic polarity zones, and depositional features within limited areas (Darton, 1899, 1905; Wanless, 1923; Schultz and Stout, 1955; Emry, 1973; Prothero, 1982, 1985; Swinehart et al., 1985; Evanoff et al., 1992). Some of the tuffs in Nebraska and South Dakota are sufficiently prominent and widespread to deserve informal names, including the upper purplish-white lower Whitney, upper Whitney, and Nonpareil ashes of Nebraska (Schultz and Stout, 1955; Swinehart et al., 1985) and the Rockyford ash of South Dakota (Nicknish and Macdonald, 1962). Their mineralogy, geochemistry, and stratigraphic position allow individual tuffs and tuff sequences to be recognized over wide areas.

Recognition of tuffs in the field

Tuff beds are not always obvious features in the field. The most useful criteria for their recognition is their white to gray color, widespread distribution, distinct bottom and diffuse top contacts, and presence of euhedral heavy minerals and glass shards. The White River tuff beds are white to gray siltstones or claystones that occur either as widespread sheets or local lenticular beds. Most are white to light gray, forming light bands in White River outcrops (Fig. 2A), although the more mafic tuffs are medium gray (Munsell color N6), forming dark gray bands in outcrops (Fig. 2B). In fluvial intervals of the White River, the tuffs are typically very distinct beds (Fig. 2A,B). Tuffs in the aeolian part of the White River are distinct from a distance but are difficult to recognize at close distance (Fig. 2C). Tuff beds are typically structureless, but they can have laminations and bedding in their upper parts. All tuffs show some degree of bioturbation from roots and invertebrates. Biotite is the most common and largest heavy mineral in White River tuffs, and the presence of euhedral crystals of biotite is the best criterion for recognizing a tuff in the field. Crystal-poor tuffs are especially

TABLE 1. LOCALITY LIST OF THE WHITE RIVER TUFFS SAMPLED IN THIS STUDY

Map Location	Locality	County	Legal Locations(s)	Tuffs Sampled	LMA(s)*
Wyoming					
1	Cameron Springs	Fremont	Sect1;T32N,R89W	1	C
2	Flagstaff Rim	Natrona	Sects22,23,34;T31N,R83W	14	C
3	Ledge Creek	Natrona	Sect22;T29N,R82W	5	C
4	South Douglas	Converse	Sects28-33;T31N,R70W Sect24;T31N,R71W Sect5;T30N,R70W	11	C, O
5	North Douglas	Converse	Sects31,32;T32N,R70W Sects27,34-36;T32N,R71W	5	O
6	Shawnee	Converse	Sect7;T32N,R68W	4	C, O
7	Seaman Hills	Niobrara	Sect32;T35N,R61W	2	O
8	Table Mountain	Goshen	Sect15;T22N,R60W	2	C, O
9	Three Tubs	Goshen	Sect5;T20N,R60W	3	C, O
Nebraska					
10	Toadstool Park Roundtop	Sioux	Sects5,8;T33N,R53W Sects16,34;T33N,R53W	6	C; O, W, A
11	Crawford	Dawes	Sect33;T32N,R51W	1	W
12	Sheridan Gates	Sheridan	Sect31;T34N,R46W	2	W
13	Beaver Wall	Sheridan	Sect10;T33N,R46W	1	A
14	Erdman Ranch	Sioux	Sect22;T24N,R56W	1	W
15	Scottsbluff	Scottsbluff	Sects27,33;T22N,R55W	4	O, W
16	Chimney Rock	Morrill	Sect20;T20N,R52W	1	W
17	Courthouse Rock	Morrill	Sect17;T19N,R50W	2	W
18	Coumbe Bluff	Garden	Sect26;T17N,R45W	2	A
Colorado					
19	West Chimney	Logan	Sect3;T11N,R54W	2	O? W
20	Twomile Creek	Weld	Sect31;T11N,R56W	1	O?
21	Pawnee Buttes	Weld	Sect28;T10N,R59W	2	O

*LMA = land-mammal age(s): C = Chadronian; O = Orellan; W = Whitneyan; A = Arikareean.

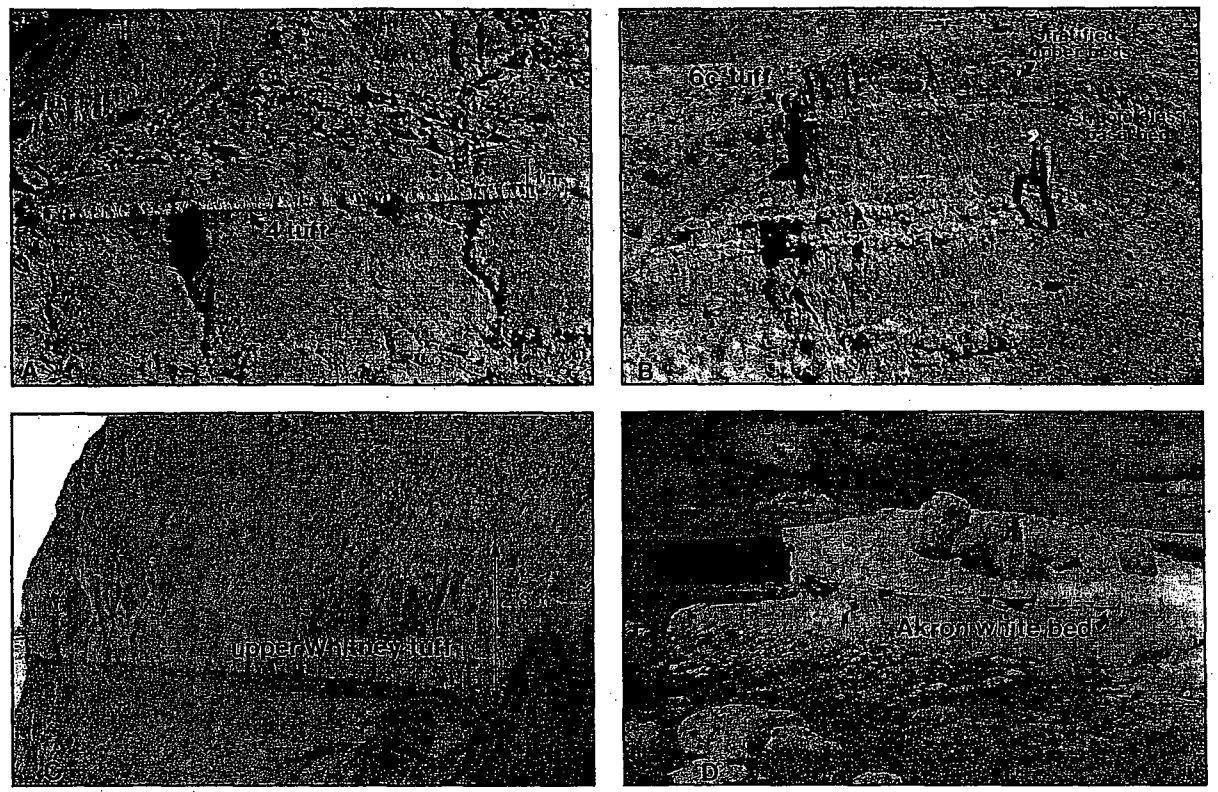


Figure 2. A variety of outcrops of White River tuffs and a tuff-like fluvial unit. A, The 4 tuff in the Douglas South area makes a white resistant ledge 0.4 m thick. The tuff is interbedded with fluvially deposited mudstones. B, The 6c black tuff in the Douglas South area forms a dark resistant band within a sequence of fluvially deposited mudstones. C, The upper Whitney tuff on the east side of Scotts Bluff, Scotts Bluff National Monument. The upper Whitney tuff was deposited in a sequence of aeolian siltstones and has a very indistinct top contact (approximately at the level of the dotted line). D, A white bed that resembles a tuff near Akron, Washington County, Colorado. Despite this bed's white color, widespread distribution, and distinct bottom contact, it is not a tuff; it contains a mineral suite that is inappropriate for that origin.

hard to distinguish in the field and require additional laboratory analysis.

Some fluvially reworked White River deposits can mimic tuffs, but they are usually distinguished by their local distribution, distinct bedding, and lack of euhedral heavy minerals. Biotite grains can be abundant in these beds, but these grains will have very few or no unweathered crystal faces. In a few cases, fluvially reworked beds can resemble a crystal-poor tuff, with widespread distribution, a distinct bottom contact, and light color, as, for example, in the Fremont-Butte area, northwest of Akron, Colorado (Figs. 1, 2D). This bed, however, contains little volcanically derived minerals, no glass shards, and no euhedral dark minerals. Moreover, microcline was a significant component of the feldspar fraction, indicating a crystalline Precambrian source component.

Recognition of tuffs in the laboratory

Laboratory studies are useful for the verification of tuffs. A simple procedure consists of breaking down the sediment in water and observing the residue petrographically. An abundance of glass shards, the conspicuous presence of euhedral biotite and feldspar, and few rounded grains of quartz and feldspar are indicative of a tuff.

Further analysis entails disaggregating 200 to 300 g of the suspected tuff. After drying, the disaggregated material is subjected to heavy-liquid separation. If the sample came from a tuff, the heavy fraction should contain a volcanic mineral assemblage, including some of the following: biotite, green-brown hornblende, clinopyroxene, orthopyroxene (hypersthene), zircon, apatite, sphene, allanite, and, rarely, monazite and chevkinite. Many of the grains, especially those of biotite, zircon, apatite,

allanite, and sphene, are euhedral or fragments of euhedral crystals (Fig. 3); some have vestiges of glass coatings. In some cases, hornblende and/or pyroxene grains are also euhedral, but commonly these exhibit serrations or needle-like terminations because of postburial etching. Rounded grains suggestive of a much older (probably Precambrian) crystalline source, such as green hornblende, tourmaline, garnet, sphene, and dark-rimmed (radiation damaged) zircon, are rare or absent. In the light fraction, there is a minimum of rounded grains of quartz and feldspar. Microcline, obviously derived from a pre-Tertiary source, is rare to nonexistent. Unless the sample has been extensively altered, glass shards should be abundant.

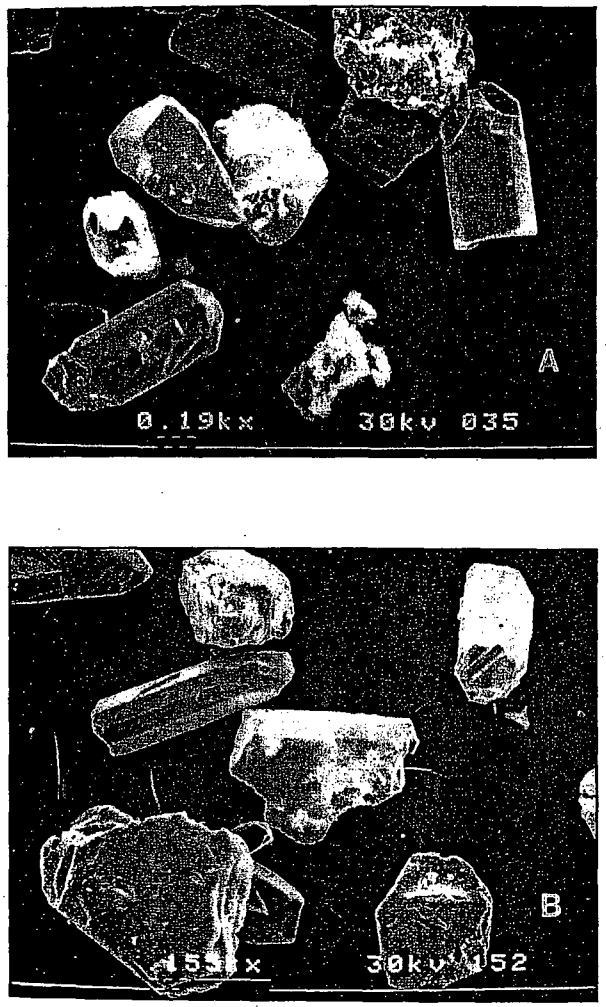


Figure 3. Scanning electron micrographs of some heavy minerals from a White River tuff, the 2 tuff from Flagstaff Rim. A, Zircon (doubly terminated tetragonal crystals); apatite (two partial crystals in upper right quadrant exhibiting top-side larger flat faces); sphene (wedge-shaped crystal in upper left quadrant); and hornblende fragment (etched grain without terminations between the apatite crystals). Offset dashed bar is 50 μ long. B, Two crystals of biotite in lower part of photograph, and four crystals of zircons (three in the upper part and one protruding from beneath the larger biotite). Offset solid bar is 100 μ long.

If microprobe analysis is carried out on some of the mineral grains, there should be a limited amount of chemical variation for each mineral species. Large ranges of variation imply significant contamination.

In most cases, the above procedures will enable ash-fall tuffs to be differentiated from nontuff units. There will be a few units, however, that defy definitive evaluation. This is particularly true for very fine grained and/or crystal-poor units that contain a great deal of clay matrix.

Fingerprinting and criteria for correlation

Correlation of an individual tuff bed across a region requires gaining as much information from a sample site as possible, so as to provide a "fingerprint" of the tuff. Most of the techniques we use have been previously discussed (Izett et al., 1972; Izett, 1981; Westgate and Gorton, 1981; and Best et al., 1989). The methods outlined below represent those that were most useful in providing characterizations of the White River tuffs.

In general, the types of data we have used in developing a tephrostratigraphy include mineralogic/geochemical, radiometric, grain-size, biostratigraphic, and paleomagnetic (Table 2). Some of these criteria can be used to match tuffs with their source region. Other criteria can be used only to match tuffs in widely separated localities. Some of the criteria are better correlation indicators than others.

Mineral suites and glass. One of the most easily obtained, yet powerful, pieces of data comes from the suite of minerals present in a tuff (e.g., Best et al., 1989). For the White River tuffs, the mineral suite nearly always comprises biotite, hornblende, plagioclase, quartz, zircon, apatite, and allanite. Sanidine occurs in many tuffs, whereas sphene is present in only ~10%. Clinopyroxene, present in a few of the tuffs, and the even rarer hypersthene, are only minor constituents. Monazite is a rare occurrence. An unu-

TABLE 2. CORRELATION CRITERIA FOR WHITE RIVER TUFFS

Criteria	Within Region	To Source
Mineralogy geochemistry		
Mineral suites and mineral abundances	X	X
Geochemical signatures (microprobe)		
Hemo-ilmenite	X	
Hornblende	X	X
Biotite	X	X
Feldspar ratios (K_2O to Na_2O)	X	X
$^{40}Ar/^{39}Ar$ Single-crystal laser-fusion dating	X	X
Biostratigraphic position	X	
Magnetic remanence directions	X	X

sual mineral in the suite provides a particularly strong positive or negative correlation criterion (Westgate and Gorton, 1981).

Several processes can alter the original mineral composition of an ash. It has been demonstrated (see Fisher and Schmincke, 1984) that some differences in the mineral suite may result from aeolian differentiation, as the heavier grains progressively drop out as the eruption cloud travels downwind. In the White River sequence, we have not found this to be a significant problem. Some changes in the mineral suite could also result from post-burial groundwater dissolution. In the White River tuffs, hornblende, hypersthene, and magnetite are most generally affected by dissolution. In some tuffs, locally, magnetite has been nearly completely dissolved.

Glass shards should be examined for maximum size, shape and textural characteristics (Izett, 1981; Westgate and Gorton, 1981; Fisher and Schmincke, 1984), inasmuch as there can be large differences in these three parameters among different tuffs. In White River tuffs, glass shards are typically poorly preserved and are, therefore, of limited use.

Microprobe analysis of mineral grains and glass. We have found hemo-ilmenite, hornblende, biotite, and alkali feldspar to be particularly useful in regional correlation. These minerals exhibit relatively large variations in both major- and minor-oxide components, thereby providing data that are potentially useful in correlation.

Hemo-ilmenite, a solid solution of hematite (Ht) and ilmenite (Il), is a common constituent in ash-fall tuffs. It is a phase quenched during rapid cooling such as that associated with explosive eruption.

For the White River tuffs, hemo-ilmenite compositions range from $Ht_{48}Il_{52}$ to Ht_5Il_{95}. Of the major-oxide components in hemo-ilmenite, there is considerable latitude for variation in the proportions of FeO, Fe_2O_3, and TiO_2. Two- and three-valence minor-oxide components, such as MnO, MgO, Cr_2O_3, Al_2O_3, and V_2O_3, also display significant ranges of variation. Because of the wide latitude in oxide variation, hemo-ilmenite provides an excellent means of tuff-to-tuff correlation. It is essentially impossible, however, to use hemo-ilmenite for correlation of a tuff with a suspected source because near-vent pyroclastic flows that cool relatively slowly after emplacement rarely contain the phase. Rather, during slow cooling, hemo-ilmenite undergoes exsolution into two separate mineral phases, hematite and ilmenite.

In the White River sequence, both hornblende and biotite can exhibit large variations between individual tuffs in content of FeO, MgO, Al_2O_3, and SiO_2, providing the potential for correlation regionally and with source regions.

For tuffs containing alkalic feldspars, we have used variation in Na/K ratio in alkali feldspars for corroboration of correlation by other means.

In estimating the initial magma compositions of White River tuffs, we have used several criteria: mean glass compositions as determined by microprobe analysis, proportion of ferromagnesian silicate minerals, presence or lack of quartz, ratio of sanidine to plagioclase, and composition of the average plagioclase.

$^{40}Ar/^{39}Ar$ ages. In the last 10 years, $^{40}Ar/^{39}Ar$ dating techniques have significantly improved. One of the biggest steps forward has come with the ability to obtain a date from a single crystal by means of laser fusion (Drake et al., 1988; Deino, 1989). In cases in which only a few stratigraphically widely spaced tuffs are involved, correlation by single-crystal laser-fusion (SCLF) dating alone may prove sufficient in regional correlation. But in sections in which many tuffs occur, a common occurrence in the White River sequence, the radiometric dates, by themselves, may not be capable of providing absolute correlation of individual tuff units from section to section. In these cases, however, the dates generally narrowed the number of tuff beds that could be correlative.

The best dates are generally obtained from alkali feldspars, with biotite a less reliable second choice. For White River tuffs, therefore, SCLF dating is particularly useful for alkali-feldspar–bearing tuffs of rhyolitic, rhyodacitic, and latitic composition, but less so for those of dacitic and andesitic composition. The method also requires crystals of a minimum size (~0.2 mm). Chadronian tuffs that are east of Douglas, Wyoming, have generally proven very difficult to date because of their fine grain size, as have most Orellan tuffs that we have sampled.

SCLF dating of the White River tuff sequence (Table 3) has provided a good estimate of the total time interval of volcanic activity. These data are of additional value in constraining the potential source areas from which the tuffs were derived.

Grain-size data. Grain sizes of zircon, biotite, hornblende, and, rarely, apatite, sphene, and glass shards were compared among tuff sites in the White River sequence, providing additional correlation data. The grain-size range for any one mineral at a sample site primarily is related to the initial size range in the Plinian cloud and the distance traveled to the site of deposition (Fisher and Schmincke, 1984).

For the White River tuffs, we have compared grain-size variation in several Chadronian tuffs mostly along an east-west transect between Flagstaff Rim and Douglas, Wyoming, ~100 km to the east. Grain sizes for two of the Orellan tuffs has also been compared, one from Douglas, Wyoming, to Scottsbluff, Nebraska, ~100 km eastward; the other, over a very large area in Wyoming, Nebraska, and Montana. We also have grain-size data for two Whitneyan tuffs from a large area in northeastern Colorado and the panhandle of Nebraska.

On the assumption that a consistent, progressive change in grain size in a tuff regionally is suggestive of fining downwind, away from the eruption source, we conclude that essentially all grain-size data strongly suggest a source region toward the west. The more areally distributed tuffs—one Orellan and two Whitneyan—indicate a southern component as well.

Biostratigraphy. Mammalian biostratigraphic relations within the White River sequence provide an essential, albeit generalized, means of correlation. For regional correlation, mammals indicating the Chadronian, Orellan, Whitneyan, and Arikareean land-mammal ages provide the best biostratigraphic resolution in the White River sequence (Emry et al., 1987). The

TABLE 3. PUBLISHED $^{40}AR/^{39}AR$ AGES FOR WHITE RIVER TUFFS

	Swisher and Prothero 1990				Obradovich et al. 1995		
	Mean	SD	SE	Mineral	Mean	SD	Mineral
Nonpareil ash	30.05	0.19	0.09	Biotite	n.d.		
Upper Whitney	30.58	0.61	0.18	Biotite	n.d.		
Lower Whitney	31.85	0.02	0.01	Biotite	n.d.		
	31.81	0.05	0.03	Anorthoclase			
	31.67	0.22	0.16	Plagioclase			
5 Tuff, Douglas	33.91	0.13	0.06	Biotite	33.59	0.02	Sanidine
Ash J	34.48	0.23	0.08	Biotite	34.36	0.11	Sanidine
	34.72	0.13	0.06	Anorthoclase			
Ash I	35.38	0.29	0.10	Biotite	n.d.		
Ash G	35.57	0.17	0.06	Biotite	n.d.		
	35.72	0.07	0.03	Anorthoclase			
Ash F	35.72	0.38	0.11	Biotite	n.d.		
	35.81	0.09	0.04	Anorthoclase			
Ash B	35.92	0.34	0.10	Biotite	35.41	0.14	Sanidine
	35.97	0.45	0.22	Anorthoclase			

Note: SD = standard deviation; SE = standard error; n.d. = not determined.

boundaries between these stages have been shown to be essentially isochronous within the resolution of magnetic polarity boundaries (Prothero, 1982). Therefore, tuffs associated with one land-mammal age fauna preclude correlation with another tuff associated with a fauna of a different land-mammal stage. Mammalian biostratigraphic relations are especially useful for isolated tuff outcrops of the White River, such as those in the mountain valleys of Wyoming and Colorado. Determining the exact identity of a tuff in a land-mammal stage, of course, requires additional correlation techniques.

Paleomagnetic remanence directions. Remanence directions of White River tuffs were of additional correlative value between the sections at Flagstaff Rim and Douglas, Wyoming. For individual tuffs in these two sections that were previously correlated by means of mineral-suite, grain-size, and radiometric data, remanence directions were consistently in agreement, both in polarity and in angular position. The method potentially has applicability for tuffs throughout the area of exposure of the White River sequence.

GEOGRAPHIC AND STRATIGRAPHIC VARIATION OF THE WHITE RIVER TUFFS

From analysis of ~70 separate ash-fall tuff sites in south-central and eastern Wyoming, northeastern Colorado, western Nebraska, and southern Montana (Fig. 1; Table 1), we have been able to establish a sequence of 25 separate tuff units that range in age from ~36 to 30 Ma. The tuffs represent, therefore, most of the Chadronian, Orellan, and Whitneyan land-mammal stages (Fig. 4). All felsic White River tuffs are characterized by a ferromagnesian suite dominated by biotite and hornblende; pyroxenes are either absent or of minor abundance.

Chadronian tuffs are primarily rhyolitic, and, to a lesser extent, rhyodacitic. Dacite compositions are entirely lacking in the Chadronian tuffs (Fig. 4). At Flagstaff Rim, Wyoming, there is one andesitic unit that is not present in the section near Douglas, ~100 km to the east.

Overall, tuffs of the Orellan tend to be finer grained than those of either the underlying Chadronian or the overlying Whitneyan, possibly indicating smaller volume eruptions or greater distance from the sources. Orellan tuffs are chiefly rhyolitic to rhyodacitic, except for a widespread andesitic tuff that has been identified at Douglas, Wyoming, Scottsbluff, Nebraska, and possibly, in equivalent rocks in southwestern Montana (Table 4). A dacitic, very coarse-grained tuff that is undated but may be Orellan, has been identified at Twomile Creek, northeastern Colorado.

The three Whitneyan tuffs, exposed principally in the Nebraska Panhandle and northeastern Colorado, are coarse grained and, locally, as much as 3 m thick, apparently as a result of aeolian processes subsequent to initial ash deposition. The lower one is rhyodacitic and the upper two are dacitic (Fig. 4).

In summary, the composition of the tuffs of the White River sequence remains predominantly rhyolitic or rhyodacitic for the Chadronian and Orellan. In moderate contrast, the three voluminous Whitneyan tuffs seem to record a change toward more mafic compositions.

VOLCANIC SOURCE AREAS

The source regions of volcaniclasts in the White River sequence have long been poorly defined. The tuffs were recognized a century ago by Darton (1898, 1899), who was also the first to recognize the large amount of volcaniclasts in the White River

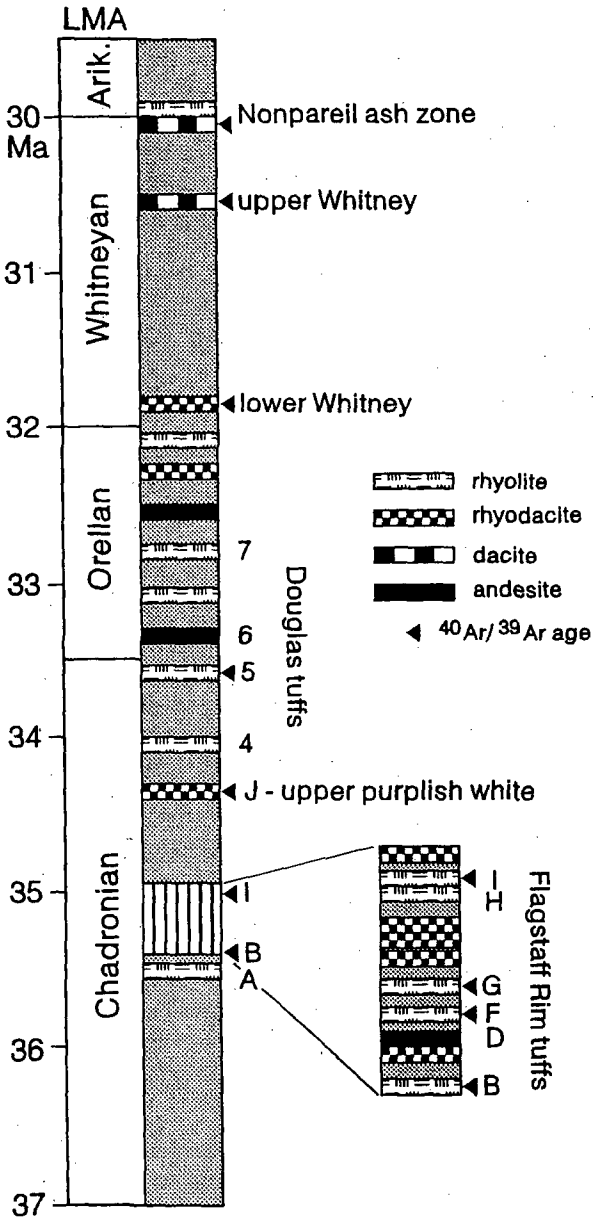


Figure 4. Chronostratigraphic distribution and composition of White River tuffs sampled in this study. Nondated tuffs were positioned in the sequence by extrapolation and interpolation. Tuff ages and land-mammal-age (LMA) boundaries are from Swisher and Prothero (1990), Prothero and Swisher (1992), and Obradovich et al. (1995). See Table 3 for details of $^{40}Ar/^{39}Ar$ ages. Arik. = Arikareean LMA.

mudrocks (Darton, 1901). He (Darton, 1912) designated Tertiary plutons in the Black Hills, South Dakota, as possible sources of the White River volcaniclasts. Wanless (1923) argued that the Black Hills plutons were too small to have produced the large volume of volcanic material. He favored a source related to an early episode of volcanism in the Yellowstone region. An Absaroka Mountains source was supported by Love (1939), on the basis of mineralogy and large bones from a possible brontothere from the Wiggins Formation, a stratigraphic unit he correlated with the White River of central Wyoming. Unfortunately, these bones were destroyed before the specimens could be identified in detail.

An Absaroka source for the White River was accepted for several decades until radiometric ages indicated that the Wiggins and associated formations were too old—middle to early late Eocene (Smedes and Prostka, 1972; Love et al., 1978; Bown, 1982). Moreover, fossil mammals associated with the Absaroka volcanic section indicate a Bridgerian to Uintan age, that is, middle to early late Eocene (McKenna, 1972; Love et al., 1978; Bown, 1982; Eaton, 1982). White River rocks occur in the Absaroka Mountains (Love et al., 1976) but are limited to paleovalley fills within the older volcanic rocks.

Recent workers have suggested source regions in Colorado (Swinehart and Rebone, 1984; Swinehart et al., 1985) or the Great Basin (Lillegraven, 1993), with little evidence except for general trends in tuff thicknesses and timing of volcanic fields in the western United States.

In trying to establish potential source areas for the White River ash-fall tuffs, three criteria are important. One is that explosive volcanism (primarily caldera-related) in a potential source region must have been active from ~36 to 30 Ma, the span of accumulation of most of the volcaniclasts in the White River sequence. A second is that the chemical compositions and mineral suites of the source-area felsic pyroclastic flows must match those of the White River tuffs. A third consideration is that the source direction should be appropriate for the trend in grain-size changes in correlated White River tuffs that decreases toward the east and northeast.

Presently, there are sufficient radiometric, petrologic, and geochemical data available to enable delineation of several fields of explosive volcanism in the western U.S. that may have served as sources for White River tuffs (Fig. 5). They include the San Juan and Thirtynine-Mile Field regions, Colorado; the Mogollon–Datil–Trans-Pecos region, New Mexico and Texas; the Sierra Madre Occidental, Mexico; the Cascades area, Oregon; and the Great Basin, Nevada and Utah.

For the most part, the tuffs in the White River did not come from volcanic activity in Colorado. The San Juan Mountains can be eliminated because their large-scale explosive volcanism (Fig. 6) did not begin until 29.5 Ma (Lipman, 1984), and their ferromagnesian suites are inappropriate. Most San Juan pyroclastic flows contain abundant pyroxenes but little or no hornblende (Lipman et al., 1970, 1978), unlike the case for White River ashes, in which the proportions are reversed.

Although eruptions from the Thirtynine-Mile Field are

TABLE 4. CORRELATIVE WHITE RIVER TUFFS IN WYOMING, NEBRASKA, AND COLORADO

LMA	Wyoming						Nebraska				Colorado	
	Cameron Springs	Flagstaff Rim	Douglas Area	Shawnee	Seaman Hills	Table Mountain	Toadstool Park	Scottsbluff	Courthouse Rock	Coumbe Bluff	West Chimney	Pawnee Buttes
Arikareean							Nonpareil			Nonpareil		
Whitneyan							Upper Whitney Lower Whitney	Upper Whitney Lower Whitney	Upper Whitney Lower Whitney		Vista tuff	
Orellan			7 tuff White tuff Black tuffs	Black tuffs	Black tuffs			Blue ash Black tuff				White tuff
Chadronian	White tuff	J ash G+73 ash G+36 ash F ash B+6C ash B tuff	5 tuff 4 tuff 3c tuff 3b tuff 3a tuff 2 tuff 1 tuff 0 tuff	White tuff White tuff White tuff White tuff	Upper Purple White	White tuff White tuff	?Upper Purple White					

Note: LMA = land-mammal ages.

appropriate in age for the Chadronian White-River tuffs, they are too few in number. Only three to four pyroclastic eruptions have been documented in the Thirtynine-Mile Field in the span from 36 to 29 Ma (McIntosh et al., 1992a,b). Moreover, the grain sizes of the Wyoming White River tuffs are generally too small for derivation from this nearby source area (see Izett, 1987). We consider only one tuff, the coarse grained dacitic one that occurs at Twomile Creek, northeastern Colorado, to be coarse grained enough to have been derived from the Thirty-nine-Mile Field.

The Mogollon–Datil–Trans-Pecos source region (Fig. 5) was explosively active during the appropriate time span. We have excluded this general region from consideration, however, because the chemistry of the felsic pyroclastic flows in that area is inappropriate, and the grain-size changes in the White River tuffs are inappropriate for a due-south source region. In the Mogollon-Datil field, felsic ash-flow units are principally low-silica rhyolites ranging in age from 36 to 31.3 Ma (McIntosh et al., 1990, 1991); no dacites have been reported. Biotite is a common constituent in most tuffs, and clinopyroxene in some of them; hornblende, however, is generally rare to nonexistent (Ratte and others, 1984; Elston et al., 1989).

Activity in the Trans-Pecos field began ~38 Ma and lasted to ~31 Ma (Barker, 1979; James and Henry, 1991; Henry et al., 1994). Of the 10 to 12 known calderas, however, about 50% formed before 36 Ma, thereby predating recognized ashes in the White River sequence. Many of the ash-flows are alkaline, and none contains hornblende. One final piece of negative evidence concerning the Mogollon–Datil–Trans-Pecos area is related to grain-size variation in the White River tuffs. If the White River Group had been derived from calderas to the south, there should have been minimal variation in grain size in the tuffs from west to east, which is contrary to the data.

Enough is known about the Sierra Madre Occidental volcanic field (Fig. 5), which was the source of as many as 350 major pyroclastic eruptions, to indicate that it was also active from ~36 to 27 Ma, during deposition of the White River sequence (McDowell and Clabaugh, 1979; Cameron et al., 1980; Swanson and McDowell, 1984; Wark et al., 1990). Had this been the source of the White River tuffs, however, we would have expected a far greater number of tuff units than have been identified thus far. And because of the distance of 1,500 to 3,000 km, many of the tuffs should have been very thin and very fine grained, which is contrary to the case for the Chadronian and Whitneyan tuffs. Also, grain-size variations from west to east in the White River tuffs would have been minimal if this southern source was operative, unlike the actual situation. Finally, in the reported Sierra Madre ash-flow units, clinopyroxene and orthopyroxene are the most abundant ferromagnesian grains; biotite and hornblende are only minor in abundance.

We have dismissed the western Cascades in the Pacific Northwest (Fig. 5) as a source region because the chemistry of the temporally appropriate volcanic activity was too mafic. The bulk of the lavas were basalts and andesites erupted from strato-

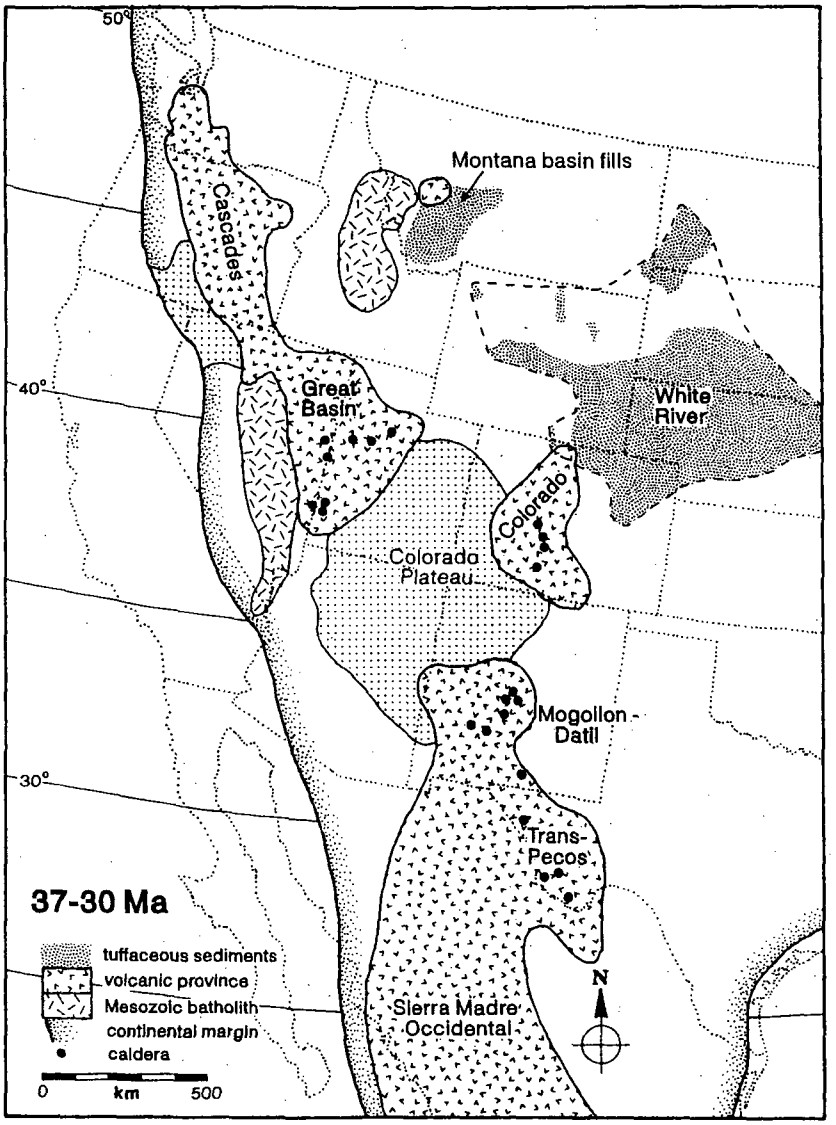


Figure 5. Palinspastic reconstruction of the central portion of North America, showing the active volcanic provinces between 37 and 30 Ma, during the deposition of the White River sequence. Most of the White River tuffs were derived from the Great Basin, indicating west-southwest prevailing winds during the latest Eocene and early Oligocene. The western continental margin position is from Heller et al. (1987) and Wells and Heller (1988). The distribution of volcanic provinces is from Stewart (1980), Chadwick (1985), and Mutschler et al. (1987). Only some of the larger known calderas are indicated.

volcanoes (Curless, 1992; Hladky et al., 1992; Hammond and Hooper, 1989; Smith, 1989).

The John Day Formation of central Oregon, which is a predominantly volcaniclastic unit (with a few ash-flow layers), primarily represents the eastward, downwind accumulation of pyroclastic material as a result of small-scale pyroclastic venting from western Cascade stratovolcanoes. Almost none of the John Day is Chadronian; most is Orellan to Arikareean (Bestland et al., 1994; Robinson et al., 1990). Moreover, most of the tuffaceous material is dacitic in composition. A few of the ash-flow units in the John Day were derived from local sources just to the west of the basin. These

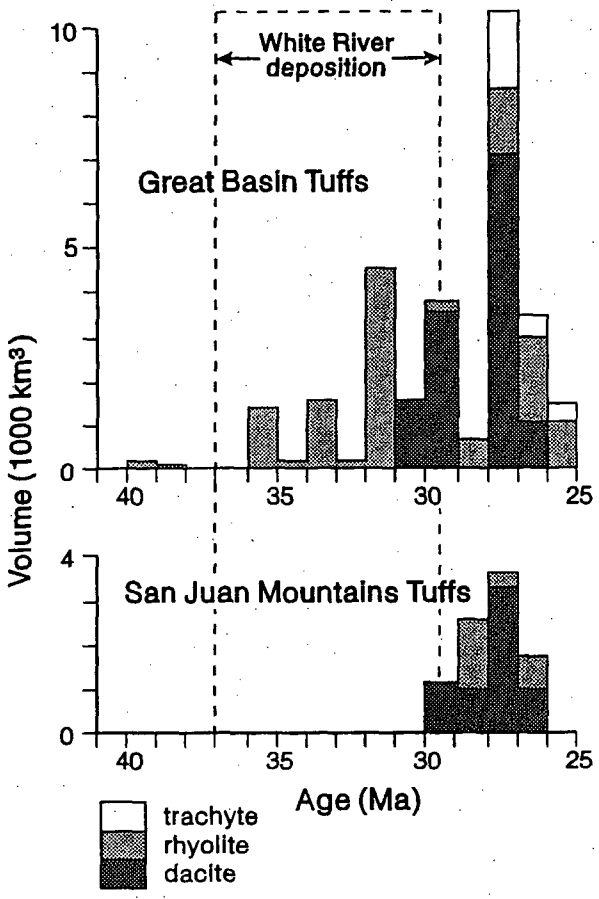


Figure 6. Histogram of the timing, frequency, and composition of late Paleogene volcanism in the San Juan and Great Basin volcanic provinces (modified from Best et al., 1989). As shown, the compositions and ages of tuffs from the San Juan Mountains are inappropriate for that area to have been the principal source of the White River ashes. In contrast, activity in the Great Basin from ~36 to 30 Ma corresponds closely with the timing of accumulation and tuff compositional variation in the White River sequence

units possess essentially no ferromagnesian minerals and are of extremely small volume, each 10 to 25 km³. The accumulation of ash-fall material related to these pyroclastic flows, hundreds of kilometers downwind, in Wyoming and Nebraska, would have been vanishingly minimal.

The remaining source region, a part of the Great Basin in eastern Nevada and western Utah (Fig. 5), satisfies all of the required criteria. It is, therefore, our principal choice for the likely source region. Explosive volcanism was active from ~36 to 29 Ma (Best et al., 1989). Over this time span, at least 30 major, felsic pyroclastic eruptions have been recognized, most related to caldera formation. As demonstrated by Best et al. (1989), the volcanism initially was rhyolitic and, to a lesser extent, rhyodacitic, before 31 Ma, but became increasingly dacitic after 31 Ma (Fig. 6). This matches the compositional characteristics of the White River ash section very well. Moreover, hornblende and biotite are very common (essentially ubiquitous) ferromagnesian minerals in the Great Basin ash-flow units; pyroxenes are generally much less abundant or lacking (Best et al., 1989; Gans et al., 1989).

A western to west-southwestern source also fits well with the decreases in grain size in many of the correlated White River ashes. On the basis of a study by Izett (1987), the maximum grain sizes of the White River tuffs, and their rate of decrease northeastward, suggest derivation from small- to large-volume eruptions at distances of ~500 to 800 km. These are appropriate distances for Great Basin sources ~36 to 30 Ma, prior to the significant amount of extension that has since affected the Great Basin.

Although a small volume of felsic pyroclastic activity in the Great Basin began as early as ~40 Ma, the majority of activity began ~36 Ma, reaching a peak at ~28 Ma (Best et al., 1989). The base of the volcaniclastic White River sequence, including that at Flagstaff Rim, Wyoming, where the Chadronian tuff sequence is most completely preserved, is also ~36 Ma, a coincidence that is probably not fortuitous. The White River sequence, which is composed of ~ 60% glass shards, therefore, ostensibly owes its existence to the increasing felsic pyroclastic activity upwind in the Great Basin.

The presence of Great Basin ash-fall material in southeastern Wyoming, northwestern Nebraska, and northeastern Colorado suggests high-level wind transport toward the east-northeast (Fig. 5). The lack of ash material in east-central and southeastern Colorado, and northeastern New Mexico, is consistent with narrow, east-northeasterly–directed ash plumes driven by prevalent winds. Alternatively, the plumes could have been much broader. The paucity of ashes in the above areas may only reflect subsequent erosion of previously deposited White River–type material, a dearth of critical exposures, or a lack of suitable basins of deposition.

DISCUSSION

The tuffs of the White River have the greatest potential to provide high-resolution correlation regionally as well as locally (Table 4). White River mammal fossils could also provide detailed biostratigraphic partitioning of the sequence, but detailed stratigraphic ranges have only been determined for the Flagstaff Rim area (Emry, 1992) and regionally for the Brule Formation (Prothero and Whittlesey, this volume). Magnetic polarity zones (Prothero, 1982, 1985, 1996; Prothero and Swisher, 1992) currently provide the most detailed correlations across the White River sequence. Magnetic polarity zones, however, are not in themselves uniquely diagnostic, and without the aid of additional stratigraphic indicators, correlation of magnetic zones in terrestrial sequences is problematical. For example, differences in depositional rates, and/or diagenetic histories between two areas, or the presence of subtle unconformities, can

Verlag, 472 p.

Gans, P. B., Mahood, G. A., and Schermer, E., 1989, Synextensional magmatism in the Basin and Range province: A case study from the eastern Great Basin: Geological Society of America Special Paper 233, 55 p.

Hammond, P. E., and Hooper, P. R., 1989, Progress report on mapping Fifes Peaks volcano, Washington, a stratovolcano model for the Western Cascades [abs.]: Northwest Regional American Geophysical Union Meeting, Eos (Transactions, American Geophysical Union), v. 71, p. 1069.

Heller, P. L., Tabor, R. W., and Suczek, C. A., 1987, Paleogeographic evolution of the United States Pacific Northwest during Paleogene time: Canadian Journal of Earth Sciences, v. 24, p. 1652–1667.

Henry, C. D., Kunk, M. J., McIntosh, W. C., 1994, $^{40}Ar/^{39}Ar$ chronology and volcanology of Trans-Pecos silicic volcanism in the Davis Mountains, Texas: Geological Society of America Bulletin, v. 106, p. 1359–1376.

Hladky, F. R., Wiley, T. J., and Duncan, R. A., 1992, Volcanic stratigraphy and new whole rock K-Ar ages of Western Cascades rocks in southern Oregon: Geological Society of America Abstracts with Programs, v. 24, no. 5, p. A33.

Izett, G. A., 1981, Volcanic ash beds: Recorders of upper Cenozoic silicic pyroclastic volcanism in the western United States: Journal of Geophysical Research, v. 86, p. 10200–10222.

Izett, G. A., 1987, Size of crystals and glass shards in Lava Creek B ash and of shock-metamorphosed minerals in K-T boundary rocks provide constraints on K-T boundary impact site: Geological Society of America Abstracts with Programs, v. 19, no. 7, p. A713.

Izett, G. A., Wilcox, R. E., and Borchardt, G. A., 1972, Correlation of a volcanic ash bed in Pleistocene deposits near Mount Blance, Texas, with the Guaje Pumice Bed of the Jemez Mountains, New Mexico: Quaternary Research, v. 2, p. 554–578.

James, E. W., and Henry, C. D., 1991, Compositional changes in Trans-Pecos Texas magmatism coincident with Cenozoic stress realignment: Journal of Geophysical Research, v. 96, p. 13561–13575.

Lillegraven, J. A., 1993, Correlation of Paleogene strata across Wyoming—a user's guide, *in* Snoke, A. W., Steidtmann, J. R., and Roberts, S. M., eds., Geology of Wyoming: Geological Survey of Wyoming Memoir 5, p. 414–477.

Lipman, P. W., 1984, The roots of ashflow calderas in western North America: Windows into the tops of granitic batholiths: Journal of Geophysical Research, v. 89, p. 8801–8841.

Lipman, P. W., Steven, T. A., and Mehnert, H. H., 1970, Volcanic history of the San Juan Mountains, Colorado, as indicated by potassium-argon dating: Geological Society of America Bulletin, v. 81, p. 2329–2352.

Lipman, P. W., Doe, B. R., Hedge, C. E., and Steven, T. A., 1978, Petrologic evolution of the San Juan volcanic field, southwestern Colorado: Pb and Sr isotope evidence: Geological Society of America Bulletin, v. 89, p. 59–82.

Love, J. D., 1939, Geology along the southern margin of the Absaroka Range, Wyoming: Geological Society of America Special Paper 20, 134 p.

Love, J. D., 1960, Crustal sedimentation and crustal movement in Wyoming: American Journal of Science, v. 258a, p. 204–214.

Love, J. D., McKenna, M. C., and Dawson, M. R., 1976, Eocene, Oligocene, and Miocene rocks and vertebrate fossils at the Emerald Lake locality, 3 miles south of Yellowstone National Park, Wyoming: U.S. Geological Survey Professional Paper 932-A, 28 p.

Love, J. D., Leopold, E. B., and Love, D. W., 1978, Eocene rocks, fossils, and geologic history, Teton Range, northwestern Wyoming: U.S. Geological Survey Professional Paper 932-B, 40 p.

McDowell, F. W., and Clabaugh, S. E., 1979, Ignimbrites of the Sierra Madre Occidental and their relation to the tectonic history of western Mexico: Geological Society of America Special Paper 180, p. 113–124.

McIntosh, W. C., Kedzie, L. L., and Sutter, J. F., 1991, Paleomagnetism and $^{40}Ar/^{39}Ar$ ages of ignimbrites, Mogollon-Datil volcanic field, southwestern New Mexico; New Mexico Bureau of Mines and Mineral Resources Bulletin 135, 79 p.

McIntosh, W. C., Sutter, J. F., Chapin, C. E., and Kedzie, L. L., 1990, High-precision $^{40}Ar/^{39}Ar$ sanidine geochronology of ignimbrites in the Mogollon-Datil volcanic field, southwestern New Mexico: Bulletin of Volcanology, v. 52, p. 584–601.

McIntosh, W. C., Geissman, J. W., Chapin, C. E., Kunk, M. J., and Henry, C. D, 1992a, Calibration of the latest Eocene-Oligocene geomagnetic polarity time scale using $^{40}Ar/^{39}Ar$ dated ignimbrites: Geology, v. 20, p. 459–463.

McIntosh, W. C., Swisher, C. C., III, and Chapin, C. E., 1992b, Single-crystal laser fusion $^{40}Ar/^{39}Ar$ sanidine ages of ignimbrites in the Thirtynine Mile volcanic field [abs.]: Eos (Transactions, American Geophysical Union), v. 73, no. 14, p. 339.

McKenna, M. C., 1972, Vertebrate paleontology of the Togwotee Pass area, northwestern Wyoming, *in* West, R. M., coordinator, Field conference on Tertiary biostratigraphy of southern and western Wyoming, Aug. 5–10, 1972, Guidebook: Society of Vertebrate Paleontologists, p. 80–101.

Murphy, E. C., Hoganson, J. W., and Forsman, N. F., 1995, The Chadron, Brule, and Arikaree Formations in North Dakota: North Dakota Geological Survey Report of Investigation 96, 144 p.

Mutschler, F. E., Larson, E. E., and Bruce, R. M., 1987, Laramide and younger magmatism in Colorado—new petrologic and tectonic variations on old themes: Colorado School of Mines Quarterly, v. 82, p. 1–47.

Nicknish, J. M., and Macdonald, J. R., 1962, Basal Miocene ash in White River badlands, South Dakota: American Association of Petroleum Geologists Bulletin, v. 46, p. 685–711.

Obradovich, J. D., Evanoff, E., and Larson, E. E., 1995, Revised single-crystal laser-fusion $^{40}Ar/^{39}Ar$ ages of Chadronian tuffs in the White River Formation of Wyoming: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A77.

Prothero, D. R., 1982, How isochronous are mammalian biostratigraphic events?: *in* Proceedings, Third North American Paleontological Convention, v. 2, p. 405–409.

Prothero, D. R., 1985, Chadronian (early Oligocene) magnetostratigraphy of eastern Wyoming: implications for the age of the Eocene-Oligocene boundary: Geology, v. 13, p. 555–565.

Prothero, D. R., 1996, Magnetic stratigraphy of the White River Group in the High Plains, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, U.K., Cambridge University Press, p. 262–277.

Prothero, D. R., and Swisher, C. C., III, 1992, Magnetostratigraphy and geochronology of the terrestrial Eocene-Oligocene transition in North America, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 46–73.

Ratte, J. C., Marvin, R. F., and Naeser, C. W., 1984, Calderas and ash-flow tuffs of the Mogollon Mountains, southwestern New Mexico: Journal of Geophysical Research, v. 89, p. 8713–8732.

Robinson, P. T., Walker, G. W., and McKee, E. H., 1990, Eocene(?), Oligocene, and lower Miocene rocks of the Blue Mountains region, *in* Geology of the Blue Mountains region of Oregon, Idaho, and Washington: U.S. Geological Survey Professional Paper 1437, p. 29–62.

Sato, Y., and Denson, N. M., 1967, Volcanism and tectonism as reflected by the distribution of nonopaque heavy minerals in some Tertiary rocks of Wyoming and adjacent states: U.S. Geological Survey Professional Paper 575-C, p. C42–C54.

Schmincke, H.-U., and van den Bogaard, P., 1991, Tephra layers and tephra events, *in* Einsele, G., Ricken, W., and Seilacher, A., eds., Cycles and events in stratigraphy: Berlin, Springer-Verlag, p. 392–429.

Schultz, C. B., and Stout, T. M., 1955, Classification of Oligocene sediments in Nebraska: University of Nebraska State Museum Bulletin, v. 4, p. 17–52.

Smedes, H. W., and Protska, J. H., 1972, Stratigraphic framework of the Absaroka volcanic supergroup in the Yellowstone National Park region: U.S. Geological Survey Professional Paper 729-C, 33 p.

Smith, J. G., 1989, Western Cascades, southern Oregon and northern California, *in* Chapin, C. E., and Zidek, J., eds., Field excursions to volcanic terranes in the western United States, Volume II: Cascades and intermountain

result in an unrecognizable mismatch of polarity zones. Our study of White River tephrostratigraphy has shown that polarity zones can be completely missing in one local section. In addition, tuffs are commonly more numerous than polarity zones, thereby providing greater stratigraphic resolution. By means of regional correlations of tuffs, a composite, high-resolution stratigraphic sequence can be compiled that combines local tephro-, magneto-, and biostratigraphic markers.

ACKNOWLEDGMENTS

We extend our appreciation to the many people and institutions that have given us support over the many years that has been needed for this study. Our greatest thanks go to James Swinehart for his stratigraphic knowledge and support on sampling tuffs in western Nebraska and adjacent parts of Wyoming and Colorado. Robert Emry introduced us to the numerous tuffs at Flagstaff Rim and Seaman Hills. Donald Prothero provided the stratigraphic framework and many lively discussions. Myron Best, Eric Christiansen, John Obradovich, and William McIntosh all provided important discussions on the ages and potential source areas for the White River volcaniclastics. Dennis Terry and Hannan LaGarry gave us valuable information concerning the stratigraphy and tuffs of western Nebraska. Gregory Cunningham, Logan Ivy, Robert Lander, and Thomas Stotler all provided help in the field. Access to the microprobe and scanning electron microscope was provided by John Drexler and the Department of Geological Sciences, University of Colorado, Boulder. Charles Chapin and Debra Hanneman reviewed early editions of this work and provided valuable suggestions for its improvement; Debra Hanneman also provided us with tuff samples from southwest Montana. Additional thanks go to Dennis Terry for insisting on including our work in this volume. Finally, we extend our appreciation to the many landowners in the study region who allowed us access to the tuff outcrops on their lands. The project was funded in part by the Nebraska Conservation and Survey Division of the Institute of Agriculture and Natural Resources, Lincoln, Nebraska, and by the University of Colorado Museum, Boulder, Colorado.

REFERENCES CITED

Barker, D. S., 1979, Cenozoic magmatism in the Trans-Pecos magmatic province: Relation to the Rio Grande Rift, *in* Riecker, R. E., ed., Rio Grande rift: Tectonics and magmatism: Washington, D.C., American Geophysical Union, p. 382–392.

Best, M. G., Christiansen, E. H., Deino, A. L., Gromme, C. S., McKee, E. H, and Noble, D. C., 1989, Eocene through Miocene volcanism in the Great Basin of the western United States: New Mexico Bureau of Mines and Mineral Resources Memoir 47, p. 91–133.

Bestland, E. A., Retallack, G. J., and Fremd, T., 1994, Sequence stratigraphy of the Eocene-Oligocene transition: Examples from the non-marine volcanically influenced John Day Basin, *in* Swanson, D. A., and Haugerud, R. A., eds., Geologic field trips in the Pacific Northwest, Volume 1: Seattle, 1994 Geological Society of America Annual Meeting, Department of Geological Sciences, University of Washington, p. 1A1–1A19.

Bown, T. M., 1982, Geology, paleontology and correlation of Eocene volcaniclastic rocks, southeast Absaroka Range, Hot Springs County, Wyoming: U.S. Geological Survey Professional Paper 1201-A, 75 p.

Cameron, K. L., Cameron, M., Bagby, W. C., Moll, E. J., and Drake, R. E., 1980, Petrologic characteristics of mid-Tertiary volcanic suites, Chihuahua, Mexico: Geology, v. 8, p. 87–91.

Cas, R. A. F., and Wright, J. V., 1988, Volcanic successions: Modern and ancient: London, Chapman and Hall, 528 p.

Chadwick, R. A., 1985, Overview of Cenozoic volcanism in the west-central United States, *in* Flores, R. M., and Kaplan, S. S., eds., Cenozoic paleogeography of west-central United States: Denver, Colorado, Rocky Mountain Section of the Society of Economic Paleontologists and Mineralogists, p. 359–382.

Curless, J. M., 1992, Geology of the Rocky Top area, Cascades, Oregon: Geological Society of America Abstracts with Programs, v. 24, no. 5, p. A18.

Darton, N. H., 1898, On the Tertiary of South Dakota and Nebraska [abs.]: Science, v. 7, p. 359.

Darton, N. H., 1899, Preliminary report on the geology and water resources of Nebraska west of the one hundred and third meridian: U.S. Geological Survey Nineteenth Annual Report, part 4, p. 721–785.

Darton, N. H., 1901, Preliminary description of the geology and water resources of the southern half of the Black Hills and adjoining regions in South Dakota and Wyoming: U.S. Geological Survey Twenty-first Annual Report, part 4, p. 489–599.

Darton, N. H., 1905, Preliminary report on the geology and underground water resources of the central Great Plains: U.S. Geological Survey Professional Paper 32, 243 p.

Darton, N. H., 1912, Volcanic action in the Black Hills of South Dakota: Science, v. 36, p. 602–603.

Deino, A., 1989, Single-crystal $^{40}Ar/^{39}Ar$ dating as an aid in correlation of ash flows: Examples from the Chimney Spring/New Pass tuffs and the Nine Hill/Bates Mountain tuffs of California and Nevada [abs.]: New Mexico Bureau of Mines and Mineral Resources, Memoir 47, p. 70.

Denson, N. M., and Gill, J. R., 1965, Uranium-bearing lignite and carbonaceous shale in the southwestern part of the Williston Basin—a regional study: U.S. Geological Survey Professional Paper 463, 75 p.

Drake, R. E., Deino, A L., Curtis, G. H., Swisher, C. S., Turrin, B., McCrory, M., and Becker, T., 1988, Applications of $^{40}Ar/^{39}Ar$ dating of single crystals by laser fusion [abs.]: Eos (Transactions, American Geophysical Union), v. 69, p. 1502.

Eaton, J. G., 1982, Paleontology and correlation of Eocene volcanic rocks in the Carter Mountain area, Park County, southeastern Absaroka Range, Wyoming: University of Wyoming Contributions to Geology, v. 21, p. 153–194.

Elston, W. E., Chapin, C. E., and McIntosh, W. C., 1989, Eocene-Miocene Mogollon-Datil volcanic field, New Mexico, *in* Chapin, C. E., and Zidek, J., eds., Field excursions to volcanic terranes in the western United States, Volume I: Southern Rocky Mountain region: New Mexico Bureau of Mines and Mineral Resources Memoir 46, p. 43–59.

Emry, R. J., 1973, Stratigraphy and preliminary biostratigraphy of the Flagstaff Rim area, Natrona County, Wyoming: Smithsonian Contributions to Paleobiology, no. 25, 20 p.

Emry, R. J., 1992, Mammalian range zones in the Chadronian White River Formation at Flagstaff Rim, Wyoming, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 106–115.

Emry, R. J., Bjork, P. R., and Russell, L. S., 1987, The Chadronian, Orellan, and Whitneyan North American Land Mammal Ages, *in* Woodburne, M. O., ed., Cenozoic mammals of North America: Berkeley, University of California Press, p. 118–152.

Evanoff, E., Prothero, D. R., and Lander, R. H., 1992, Eocene-Oligocene climatic change in North America: The White River Formation near Douglas, east-central Wyoming, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 116–130.

Fisher, R. V., and Schmincke, H.-U., 1984, Pyroclastic rocks: Berlin, Springer-

West: New Mexico Bureau of Mines and Mineral Resources Memoir 47, p. 211–215.

Stewart, J. H., 1980, Geology of Nevada: a discussion to accompany the geologic map of Nevada: Nevada Bureau of Mines and Geology Special Publication 4, 136 p.

Swanson, E. R., and McDowell, F. W., 1984, Calderas of the Sierra Madre Occidental volcanic field, western Mexico: Journal of Geophysical Research, v. 89, p. 8787–8799.

Swinehart, J. B., and Diffendal, R. F., Jr., 1990, Geology of the pre-dune strata, *in* Bleed, A., and Flowerday, C., eds., An atlas of the Sand Hills, 2nd ed.: Conservation and Survey Division, Institute of Agriculture and Natural Resources, University of Nebraska, Lincoln, Resource Atlas 5a, p. 29–42.

Swinehart, J. B., and Rebone, M., 1984, Cenozoic geology of the southern Nebraska panhandle; *in* 1984 Fall Field Trip Guidebook: Nebraska Geological Society, 35 p.

Swinehart, J. B., Souders, V. L., Degraw, H. M., and Diffendal, R. F., Jr., 1985, Cenozoic paleogeography of western Nebraska, *in* Flores, R. M., and Kaplan, S. S., eds., Cenozoic paleogeography of west-central United States: Denver, Colorado, Rocky Mountain Section, Society of Economic Paleontologists and Mineralogists, p. 209–229.

Swisher, C. C., III, and Prothero, D. R., 1990, Single-crystal $^{40}Ar/^{39}Ar$ dating of the Eocene-Oligocene transition in North America: Science, v. 249, p. 760–762.

Tucker, M. E., 1981, Sedimentary petrology, an introduction: New York, Halsted Press, John Wiley & Sons, 252 p.

Wanless, H. R., 1923, The stratigraphy of the White River beds of South Dakota; *in* Proceedings of the American Philosophical Society, v. 62, p. 190–269.

Wark, D. A., Kempter, K. A., and McDowell, F. M., 1990, Evolution of waning, subduction-related magmatism, northern Sierra Madre Occidental, Mexico: Geological Society of America Bulletin, v. 102, p. 1555–1564.

Wells, R. E., and Heller, P. L., 1988, The relative contribution of accretion, shear, and extension of Cenozoic rotation the Pacific Northwest: Geological Society of America Bulletin, v. 100, p. 325–338.

Westgate, J. A., and Gorton, M. P., 1981, Correlation techniques in tephra studies, *in* Self, S., and Sparks, R. S. J., eds., Tephra studies: Dordrecht, Holland, D. Reidel Publishing Co., p. 73–94.

Wood, H. E., 2nd, 1947, Oligocene faunas, facies, and formations, *in* Longwell, C. R., chairman, Sedimentary facies in geologic history: Geological Society of America Memoir 39, p. 83–90.

MANUSCRIPT ACCEPTED BY THE SOCIETY DECEMBER 5, 1997

Printed in U.S.A.

Geological Society of America
Special Paper 325
1998

Lithostratigraphic revision and correlation of the lower part of the White River Group: South Dakota to Nebraska

Dennis O. Terry, Jr.
Department of Geology, University of Nebraska—Lincoln, Lincoln, Nebraska 68588-0340

ABSTRACT

Lithologic correlations between type areas of the White River Group in Nebraska and South Dakota have resulted in a revised lithostratigraphy for the lower part of the White River Group. The following pedostratigraphic and lithostratigraphic units, from oldest to youngest, are newly recognized in northwestern Nebraska and can be correlated with units in the Big Badlands of South Dakota: the Yellow Mounds Paleosol Equivalent, Interior and Weta Paleosol Equivalents, Chamberlain Pass Formation, and Peanut Peak Member of the Chadron Formation. The term "Interior Paleosol Complex," used for the brightly colored zone at the base of the White River Group in northwestern Nebraska, is abandoned in favor of a two-part division. The lower part is related to the Yellow Mounds Paleosol Series of South Dakota and represents the pedogenically modified Cretaceous Pierre Shale. The upper part is composed of the unconformably overlying, pedogenically modified overbank mudstone facies of the Chamberlain Pass Formation (which contains the Interior and Weta Paleosol Series in South Dakota). Greenish-white channel sandstones at the base of the Chadron Formation in Nebraska (previously correlated to the Ahearn Member of the Chadron Formation in South Dakota) herein are correlated to the channel sandstone facies of the Chamberlain Pass Formation in South Dakota. The Chamberlain Pass Formation is unconformably overlain by the Chadron Formation in South Dakota and Nebraska. The Chadron Formation in South Dakota is divided into the Ahearn, Crazy Johnson, and Peanut Peak Members in South Dakota. Of these three, only the bluish-green, hummocky weathering mudstones of the Peanut Peak Member are present in northwestern Nebraska. These mudstones were previously classified as the lower part of the "Chadron B" in Nebraska and correlated with the Crazy Johnson Member in South Dakota. The term Chadron B is abandoned and the correlation to the Crazy Johnson Member rejected. The remainder of the Chadron Formation in northwestern Nebraska (the Chadron B and "Chadron C"), along with the basal beds of the Brule Formation ("Orella A"), are included within the newly defined Big Cottonwood Creek Member. The Chadron C was initially used to define strata between the "upper" and "lower purplish-white layers" and was correlated to the Peanut Peak Member in South Dakota. The term Chadron C is abandoned and the correlation with the Peanut Peak Member of South Dakota is rejected. The contact between the Peanut Peak Member and Big Cottonwood Creek Member of the Chadron Formation in northwestern Nebraska is intertonguing, except where strata of the Big Cottonwood Creek Member fill depressions and minor valleys in the Peanut Peak Member. The Peanut Peak Member of northwestern Nebraska differs from the Big

Terry, D. O., Jr., 1998, Lithostratigraphic revision and correlation of the lower part of the White River Group: South Dakota to Nebraska, *in* Terry, D. O., Jr., LaGarry, H. E., and Hunt, R. M., Jr., eds., Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America): Boulder, Colorado, Geological Society of America Special Paper 325.

Cottonwood Creek Member in that the Peanut Peak Member is composed of smectite-rich mudstone and claystone, weathers into hummocky hills and slopes, is less variegated in color, and has less silt. The Big Cottonwood Creek Member is siltier, cliff forming, and contains the various purplish-white layers of Schultz and Stout (1955). These proposed revisions and correlations are applicable to other exposures of the White River Group across the Great Plains.

INTRODUCTION

The Eocene/Oligocene White River Group of the northern Great Plains is composed of a thick sequence of volcaniclastic fluvial, eolian, and lacustrine strata that extend from southwestern North Dakota to northeastern Colorado (Fig. 1). The White River Group has been the subject of numerous studies over the past 150 yr. Starting with the expeditions of Meek and Hayden (1857) and Darton (1899), research on vertebrate and invertebrate paleontology, sedimentology, stratigraphy, and paleopedology of the White River Group has provided crucial information on mammalian evolution and the nature of late Paleogene paleoclimates (Hatcher, 1893; Wanless, 1922, 1923; Ward, 1922; Clark et al., 1967; Prothero and Berggren, 1992). Recent research has refined the chronology, biostratigraphy, and lithostratigraphy of the White River Group. The Chadronian-Orellan boundary, now placed at 33.59 ± 0.02 Ma (Obradovich et al., 1995), restricts the Chadronian Land-Mammal Age to the Late Eocene (Prothero and Swisher, 1992); a revision of the defining characters of the Chadronian/Orellan North American Land Mammal Age has just been proposed (Prothero and Whittlesey, this volume); a new lithostratigraphic unit, the Chamberlain Pass Formation, has been recognized as the oldest unit of the White River Group in the Big Badlands of South Dakota (Evans and Terry, 1994; Terry and Evans, 1994), and the lithostratigraphy of the White River Group of southwestern North Dakota has been recently revised (Murphy et al., 1993).

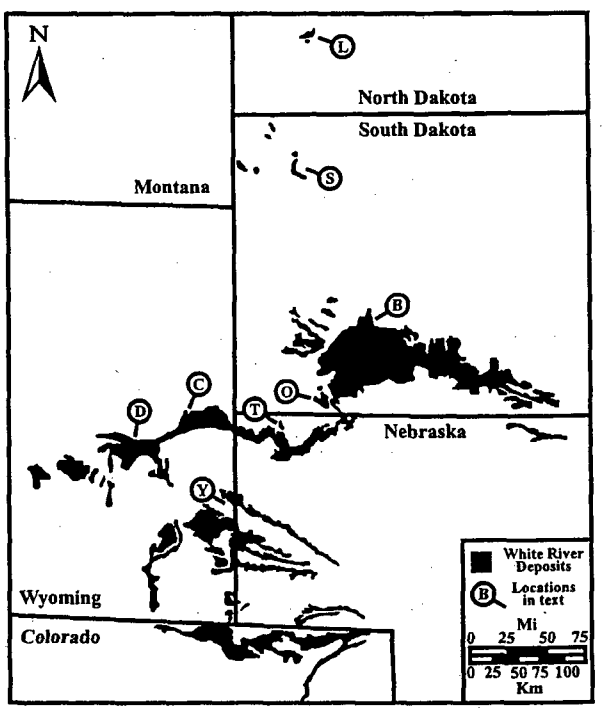


Figure 1. Generalized outcrop map of the White River Group/Formation across the northern Great Plains. Letters indicate particular research localities: B = Big Badlands/Badlands National Park, South Dakota; C = Lance Creek, Wyoming; D = Douglas, Wyoming; L = Little Badlands, North Dakota; O = Oelrichs, South Dakota; S = Slim Buttes, South Dakota; T = Toadstool Park, Nebraska; Y = Yoder, Wyoming. Modified from the American Association of Petroleum Geologists Geological Highway Map Series; Northern Great Plains (Map 12) (AAPG, 1972) and Northern Rocky Mountain Region (Map 5) (AAPG, 1984).

Based on my recent observations of the White River Group between the Big Badlands of South Dakota and the Toadstool Park area of northwestern Nebraska (Fig. 1), I propose a new lithostratigraphy for the lower part of the White River Group of northwestern Nebraska, namely, the Interior Paleosol Complex of Schultz and Stout (1955) and Chadron Formation, and provide herein a new lithostratigraphic correlation model to the Big Badlands of South Dakota. These lithostratigraphic revisions are based on criteria within the North American Stratigraphic Code (NACSN, 1983) for the recognition and definition of lithostratigraphic units, detailed measured sections, visual tracing of rock units, paleopedology, and use of the Big Badlands of South Dakota as a "standard section."

GEOGRAPHIC AND GEOLOGIC SETTING

The study area is located south of the Black Hills in the vicinity of Toadstool Park, Nebraska, a recreational area administered by the U.S. Forest Service (Fig. 1). Outcrops of the late Eocene–Oligocene White River Group in this area form the base of the Pine Ridge escarpment across Sioux and Dawes Counties. Isolated outliers occur north of the Pine Ridge escarpment and are either capped and protected by more resistant layers, such as limestone or sandstone, to form buttes and tables, or are eroded into steeply sloping and spired outcrops. Regional uplift during the Laramide Orogeny forced the retreat of the Cretaceous Interior Seaway, resulting in subaerial exposure and weathering of Cretaceous sediments in the study area. The Hartville, Laramie, and Black Hills uplifts provided sediment for late Eocene rivers that flowed east-southeast across the study area (Clark, 1975; Stanley and Benson, 1979). Uplift of the Black Hills is believed to have

ended in the Eocene prior to deposition of the White River Group (Lisenbee and DeWitt, 1993). The majority of the White River Group is composed of air-fall and fluvially reworked volcaniclastics that rest unconformably on the pedogenically modified Cretaceous Pierre Shale. The volcaniclastics that constitute a large part of the White River Group were probably derived from volcanic sources in Nevada and Utah (Larson and Evanoff, 1995; this volume). The White River Group in northwestern Nebraska is overlain by the Arikaree Group.

SOUTH DAKOTA

I use the Big Badlands of South Dakota as a lithostratigraphic standard with which to compare the lithostratigraphy of northwestern Nebraska. The majority of recent stratigraphic revision and interpretation related to the Chadron Formation and older units is based on measured sections and descriptions in the Big Badlands of South Dakota (Retallack, 1983, Evans and Terry, 1994; Terry and Evans, 1994). These recent changes, along with older concepts, are briefly reviewed so that comparisons with Nebraska sections can be made.

The 'Interior Weathered Zone' of South Dakota

Across the northern Great Plains, the White River Group rests on a zone of ancient pedogenic modification. Several names have been applied to this zone of pedogenic alteration, including the Interior Phase (Toepelman, 1922; Ward, 1922), Interior Formation (Wanless, 1922), Interior Paleosol Complex (Schultz and Stout, 1955), Eocene Paleosol (Pettyjohn, 1966), Interior Zone (Clark et al., 1967), and Interior Paleosol (Harksen and Macdonald, 1969a; Martin, 1987). Subsequently, Retallack (1983) determined that this "zone" in the Big Badlands of South Dakota was composed of two separate paleosols, the lower Yellow Mounds Paleosol Series that developed on the Pierre Shale following the retreat of the Cretaceous Seaway, and the upper Interior Paleosol Series that developed on distinct overlying fluvial deposits (Fig. 2A). The Yellow Mounds Paleosol Series in the Big Badlands is easily recognized as a bright yellow- and orange-colored zone, up to 26 m (85 ft) in some locations, which gradually gives way to unaltered Pierre Shale with increasing depth. The Interior Paleosol Series in the Big Badlands of South Dakota is easily recognized as the reddish band seen above the Yellow Mounds Paleosol Series at the top of the Interior Zone (Fig. 2A).

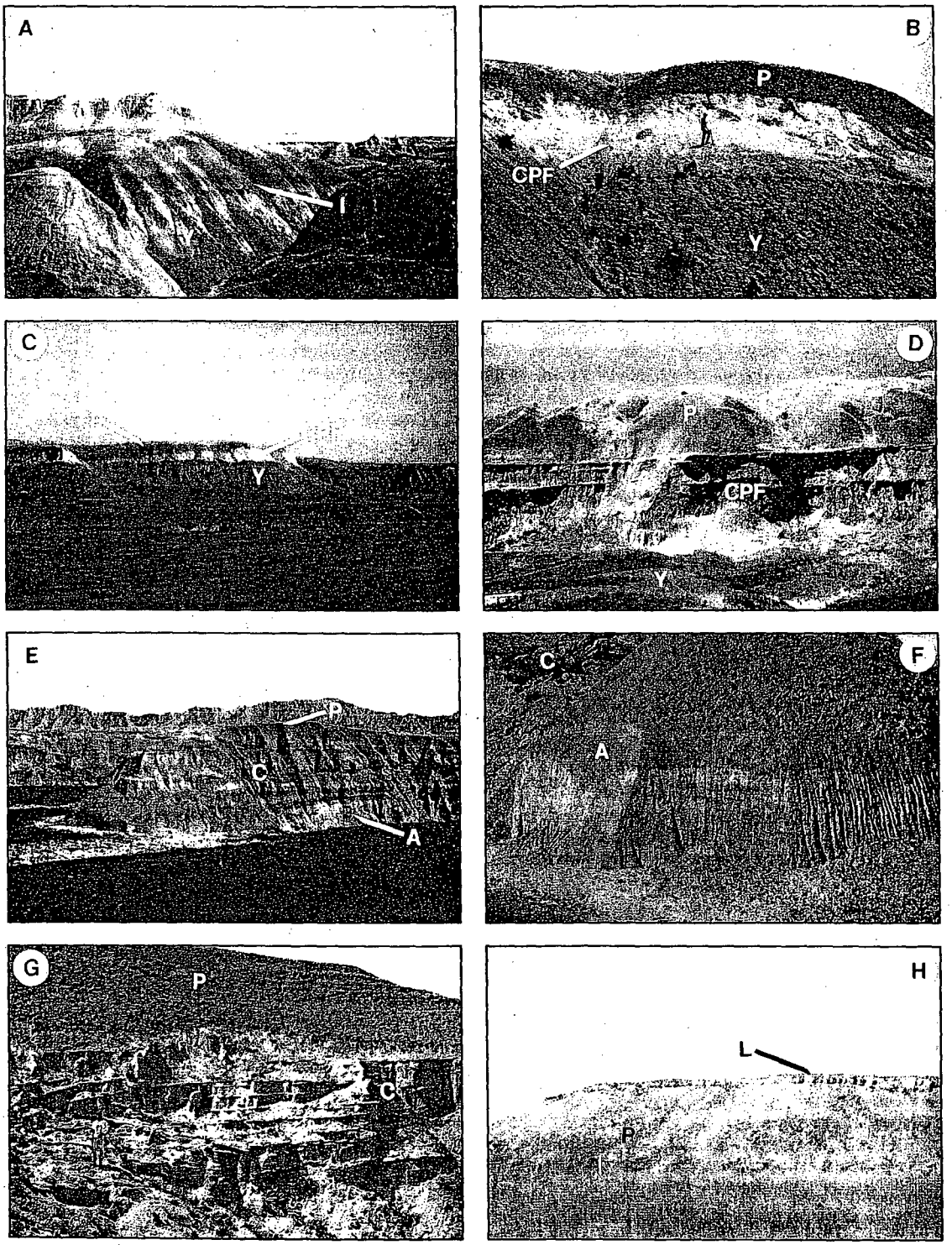
Terry (1991), Evans and Terry (1994), and Terry and Evans (1994) determined that the Interior Paleosol Series of Retallack (1983) represents the pedogenically modified distal overbank deposits of a distinct fluvial system predating the overlying Chadron Formation (Fig. 3A). Evans and Terry (1994) combined the Interior Paleosol Series, along with isolated, "blazing white" channel sandstone deposits considered by Clark et al. (1967) to be equivalent to the Slim Buttes Formation of northwestern South Dakota, into the Chamberlain Pass Formation in the Big Badlands of South Dakota (Fig. 2A–D). The Chamberlain Pass Formation represents the oldest preserved phase of Paleogene fluvial activity in the Big Badlands region following the final retreat of the Cretaceous Seaway. The Chamberlain Pass Formation of South Dakota is in all cases overlain by massive bluish-green and gray hummocky mudstones lithologically equivalent to the "undifferentiated Chadron Formation" (Fig. 2A–D) found outside of the Red River Valley of Clark (1937).

Chadron Formation of South Dakota

The intentions behind, and locations of, type section(s) for the Chadron Formation are not universally agreed on. The history of stratigraphic nomenclature for the Chadron Formation was reviewed by Harksen and Macdonald (1969a) and Singler and Picard (1980). I mention only those studies relevant to my stratigraphic revisions and correlations. Clark (1937) determined that the Chadron Formation was divisible into an eastern and western facies. According to Clark (1937), the eastern facies is composed of three members: (1) restricted basal white channel fills (Fig. 2C,D); (2) massive clays that contain intermittent limestone bands (Fig. 2C,D); and (3) laminated lenses of fine sand, silt, and limestone. The western facies is also composed of three members (Fig. 2E). The lowermost member consists of red and green sands and clays that fill the bottom of, and are confined to, a paleovalley (the Red River Valley) cut into and through the underlying Interior Zone (Fig. 2F). The base of the lower member is marked by a coarse gravel of quartz, quartzite, and granitic debris. The middle member is a succession of greenish clays and sands 12 to 15 m thick (40–50 ft) that contains fossils of brontotheres, rhinoceroses, *Mesohippus*, and *Archaeotherium* (Fig. 2G). The middle member is overlain by 6 to 9m (20–30 ft) of massive buff and green clay that contains numerous discontinuous limestone lenses and occasional, sharply restricted greenish-colored sandstone lenses (Fig. 2E,G). Clark recognized the upper member based on the decrease of sandstone and the disappearance of brontothere bones.

Based on an interfingering relationship of the eastern and western facies, Clark (1937) determined that: (1) the lower member of the western facies was restricted to the Red River Valley cut into the Interior Zone, but was somehow related to the basal white channel fills of the eastern facies; (2) the middle member of the western facies graded into the massive clays of the eastern facies; and (3) the upper member of the western facies correlates to the laminated lenses of the eastern facies. Clark (1954) eventually designated the lower, middle, and upper members of the western facies as the Ahearn, Crazy Johnson, and Peanut Peak Members, respectively (Fig. 2E).

Clark (1937) was uncertain of the exact age relationships of the eastern facies basal white channel fills and the lowermost member of the western facies (Ahearn Member; Fig. 2B,C,F). He recognized a difference in mineralogy and paleotopographic position, and suggested that the basal white channel fills of the eastern facies were tributaries that were left hanging by the downcutting of the ancient Red River (Clark, 1937). Clark et al.


A
I
B
P
CPF
Y
C
Y
D
P
CPF
Y
E
P
C
A
F
C
A
G
P
C
H
L
P

(1967) later suggested a correlation of the basal white channel fills of the eastern facies to the Eocene Slim Buttes Formation in Harding County, northwestern South Dakota. According to Clark et al. (1967), the Crazy Johnson and Peanut Peak Members are identifiable only where they overlie the Red River Valley (Figs. 2E–G, 3A). These two units blend indistinguishably into a uniform blanket of undifferentiated, bluish-green, hummocky mudstones outside the Red River paleovalley (Figs. 2A–D, 3A).

The stratigraphic succession of pedogenically altered Pierre Shale (Yellow Mounds Paleosol Series), Chamberlain Pass Formation (Interior and Weta Paleosol Series, and blazing white channel sandstone), and deposits of the Chadron Formation outside the Red River Valley produces a characteristic pattern of colors and lithologies that can be recognized in outcrops of the White River Group throughout the Big Badlands region (Figs. 2A–D, 3A). The entire Big Badlands region is underlain by pedogenically altered Pierre Shale, which displays mainly bright yellow colors, but also contains purple, lavender, and orange (Fig. 2A). The altered Pierre Shale is unconformably overlain by the Chamberlain Pass Formation. The Chamberlain Pass Formation is easily recognized as the reddish stripe at the top of the Interior Zone and white channel sandstone bodies (Fig. 2A–D). The Chamberlain Pass Formation is overlain by the bluish-green, hummocky, undifferentiated mudstone of the Chadron Formation deposited outside of the Red River Valley (Fig. 2A–D). The undifferentiated Chadron mudstone in the Big Badlands is overlain by the Brule Formation. The contact between the Chadron and Brule Formations is easily recognized by the change from the bluish-green, hummocky, popcorn-textured weathered surface of the Chadron Formation to the tread-and-riser topography of the silty, brown, beige, and red-striped Brule Formation (Fig. 2A). The upper part of the Peanut Peak Member occasionally contains several lacustrine limestone beds (Evans and Welzenbach, this volume). These limestone beds sometimes mark the contact of the Chadron and Brule Formations (D. O. Terry, 1995, unpublished data), and also occur as resistant caprocks on isolated buttes protecting the underlying Peanut Peak Member (Fig. 2H).

Figure 2. Photographs of the pedogenically modified Pierre Shale (Yellow Mounds Paleosol Series = Y), Interior Paleosol Series (I) within the Chamberlain Pass Formation (CPF), Peanut Peak Member (P) of the Chadron Formation, and Scenic Member (S) of the Brule Formation outside of the Red River Valley in the Big Badlands of South Dakota. A; Location D in Figure 3A; B; location A in Figure 3A; C; location E in Figure 3A; D; location G in Figure 3A; E–G; Ahearn (A), Crazy Johnson (C), and Peanut Peak Members of the Chadron Formation within the Red River Valley of Clark et al. (1967) in Figure 3A. Red River Valley outcrops located in the NW1/4 sec. 33, T. 43 N., R. 12 E., Heutmacher Table. Pick in bottom center of 2F, 90 cm long. Note person in center left of Figure 2G for scale. H: Lacustrine limestone (L), similar to the Bloom Basin limestone beds of Evans and Welzenbach (this volume), overlying the Peanut Peak Member of the Chadron Formation. Outcrop located in the S1/2NE1/4 sec. 12, T. 4 S., R. 18 E., Conata 7.5′ Quadrangle.

OBSERVATIONS IN NORTHWESTERN NEBRASKA

The same stratigraphic succession of lithology and colors of the pedogenically altered Pierre Shale, Chamberlain Pass Formation, and undifferentiated Chadron Formation in the Big Badlands of South Dakota is also present in the Toadstool Park region of northwestern Nebraska. Based on measured sections, paleopedology, and visual tracing of these units between the Big Badlands and Toadstool Park, the following pedostratigraphic and lithostratigraphic terms are extended from the Big Badlands of South Dakota and applied to exposures in the Toadstool Park area: *Yellow Mounds Paleosol Equivalent, Interior Paleosol Equivalent, Weta Paleosol Equivalent, Chamberlain Pass Formation,* and *Peanut Peak Member* of the Chadron Formation. This terminology is used throughout the following descriptions of paleosol profiles and measured sections in the Toadstool Park area. Comparisons with previous stratigraphic paradigms in the Toadstool Park area and concepts of regional correlation are discussed following descriptions of measured sections.

'Interior Paleosol Complex' of northwestern Nebraska

The same episodes of pedogenesis that formed the Yellow Mounds and Interior Paleosol Series of Retallack (1983) in South Dakota are herein recognized in Nebraska (Fig. 4) and serve to divide the Interior Paleosol Complex of Schultz and Stout (1955) into two distinct paleosol profiles. These episodes of pedogenesis qualify as *geosols*, ancient land surfaces of regional extent and stratigraphic utility that represent significant periods of pedogenesis (NACSN, 1983). Application of the terms Yellow Mounds, Interior, and Weta within the following paleosol descriptions does not imply that the same types of paleosols are preserved within the Toadstool Park region, only that they are similar in their general environment and parent materials on which they formed.

Within modern landscapes, the type of soil, and hence the classification of that soil, can change within distances of only a meter. This concept is valid for paleosols as well. Retallack (1994) has proposed the term *pedotype* for recognizing individual types of paleosol profiles. Each pedotype serves as a "type profile" to which other paleosols can be compared. Pedotypes are named for the area in which they occur, or for distinct characteristics of the paleosol. At this time, the terms Yellow Mounds, Interior, and Weta are valid only for the original sections in which they were described. The paleosols within the Interior Paleosol Complex of northwestern Nebraska, specifically within the Chamberlain Pass Formation, appear to differ sufficiently from the original descriptions of Retallack (1983) and Terry and Evans (1994) to warrant classification as new pedotypes. This is beyond the scope of this chapter. I therefore use the term "paleosol equivalent" to recognize a correlative *episode of pedogenesis* that created similar altered zones on top of the Pierre Shale (Yellow Mounds *Paleosol Equivalent*), and altered the Chamberlain Pass Formation (Interior *Paleosol*

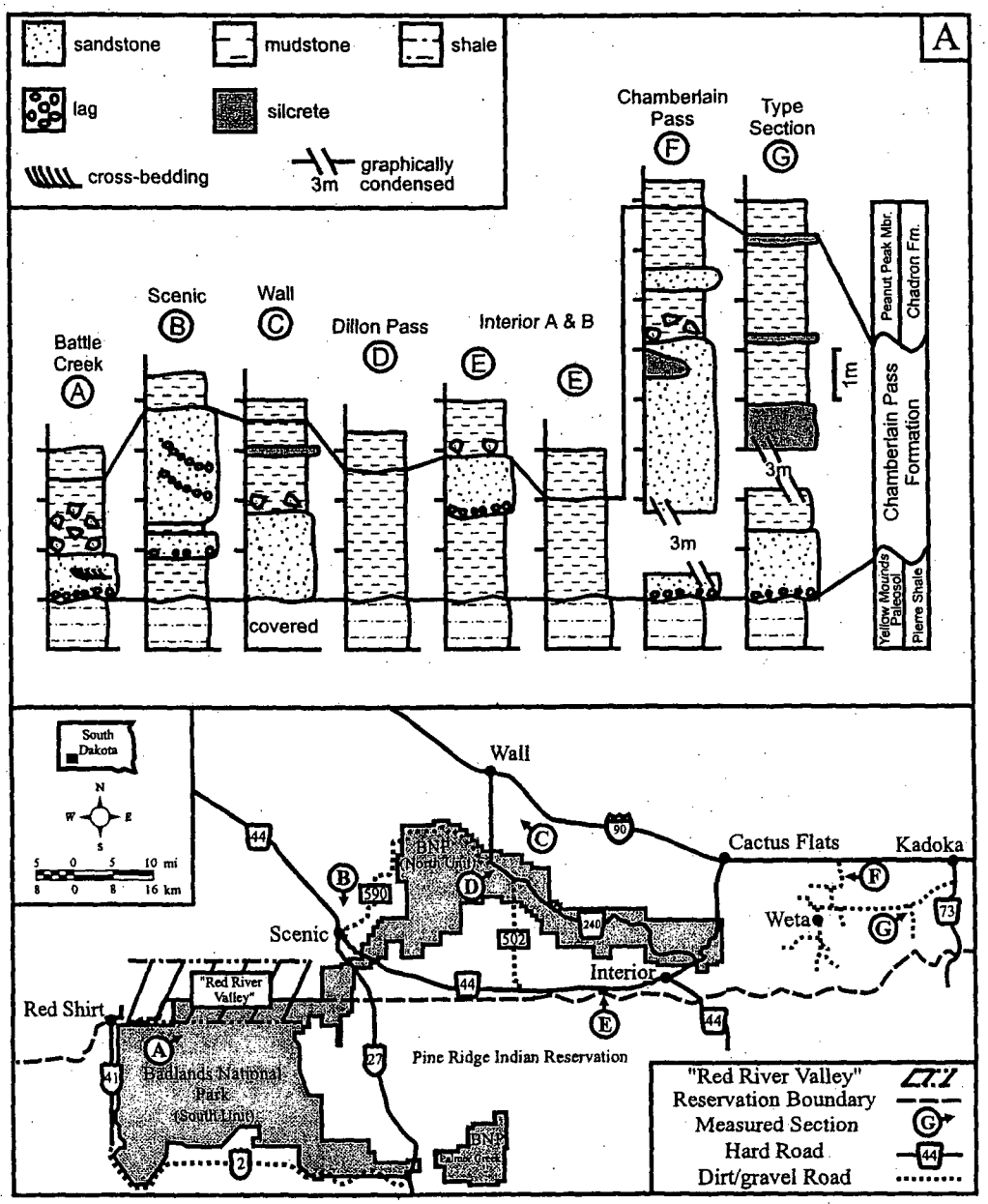


Figure 3A: Location map and measured sections of the Yellow Mounds Paleosol Equivalent, Chamberlain Pass Formation, and Peanut Peak Member of the Chadron Formation in the Badlands National Park area in relation to the "Red River Valley" of Clark et al. (1967) (modified from Evans and Terry, 1994). The section at Dillon Pass is based on data from Retallack (1983). Locations of measured sections as follows: A: SW1/4 sec. 12, T. 42 N., R. 46 W., Heutmacher Table 7.5′ Quadrangle. B: Center of sec. 11, T. 3 S., R. 13 E., Scenic 7.5′ Quadrangle. C: SW1/4SE1/4 sec. 26, T. 1 S., R. 16 E., Wall 7.5′ Quadrangle. D: NW1/4SW1/4 sec. 21, T. 2 S., R. 16 E., Wall SW 7.5′ Quadrangle. E: SE1/4SW1/4NW1/4 sec. 12, T. 4 S., R. 17 E., Conata 7.5′ Quadrangle. F: NE1/4NE1/4SE1/4NW1/4, sec. 2, T. 3 S., R. 20 E. G: E 1/2NE1/4, sec. 21, T. 3 S., R. 21 E., Kadoka Quadrangle.

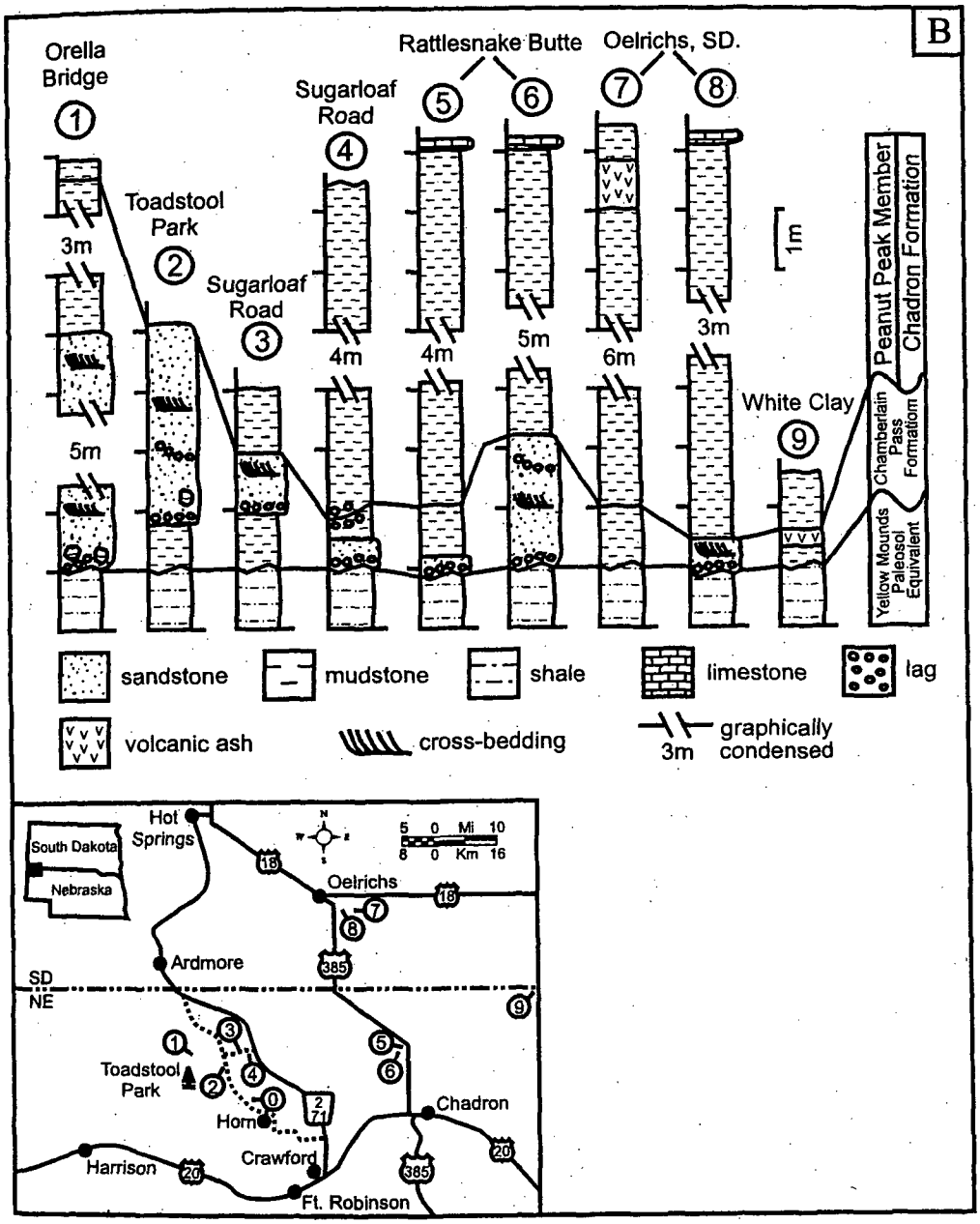


Figure 3B: Location map and measured sections of the Yellow Mounds Paleosol Equivalent, Chamberlain Pass Formation, and Peanut Peak Member of the Chadron Formation in northwestern Nebraska and southwestern South Dakota. Modified from Terry et al. (1995). Locations of measured sections are as follows: 1: NE1/4NE1/4SW1/4SW1/4SEC 36, T. 34 N., R. 54 W., Five Points 7.57, T. 34 N., R. 52 W., Wolf Butte 7.5′ Quadrangle. 4: NE1/4SE1/4SW1/4 sec. 18, T. 34 N., R. 52 W., Wolf Butte 7.5′ Quadrangle. 5, 6: NE1/2 sec. 9, T. 34 N., R. 49 W., Bohemian Creek 7.5′ Quadrangle. 7: SE1/4SE1/4SW1/4 sec. 16, T. 10 S., R. 8 E., Oelrichs 7.5′ Quadrangle. 8. SE1/4SE1/4 sec 17, T. 10 S., R. 8 E., Oelrichs 7.5′ Quadrangle. 9. NW1/4 sec. 24, T. 35 N., R. 45 W., Whiteclay 7.5′ Quadrangle.

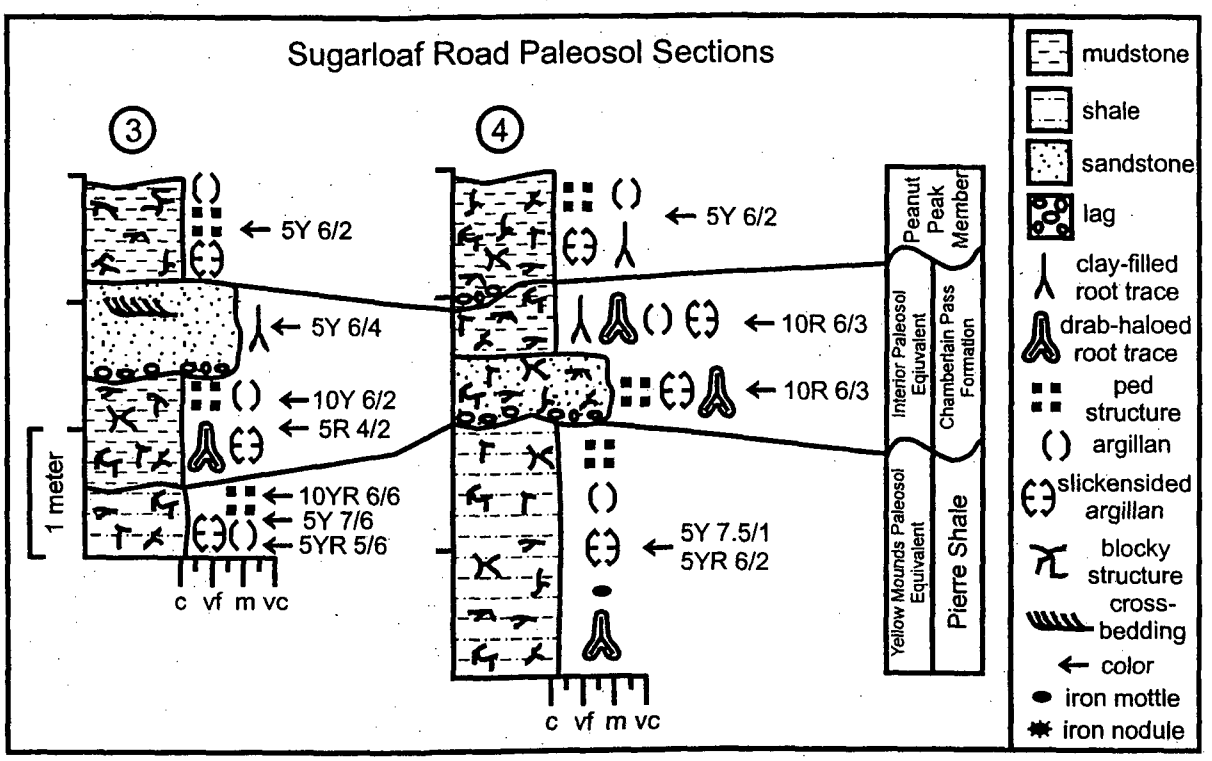


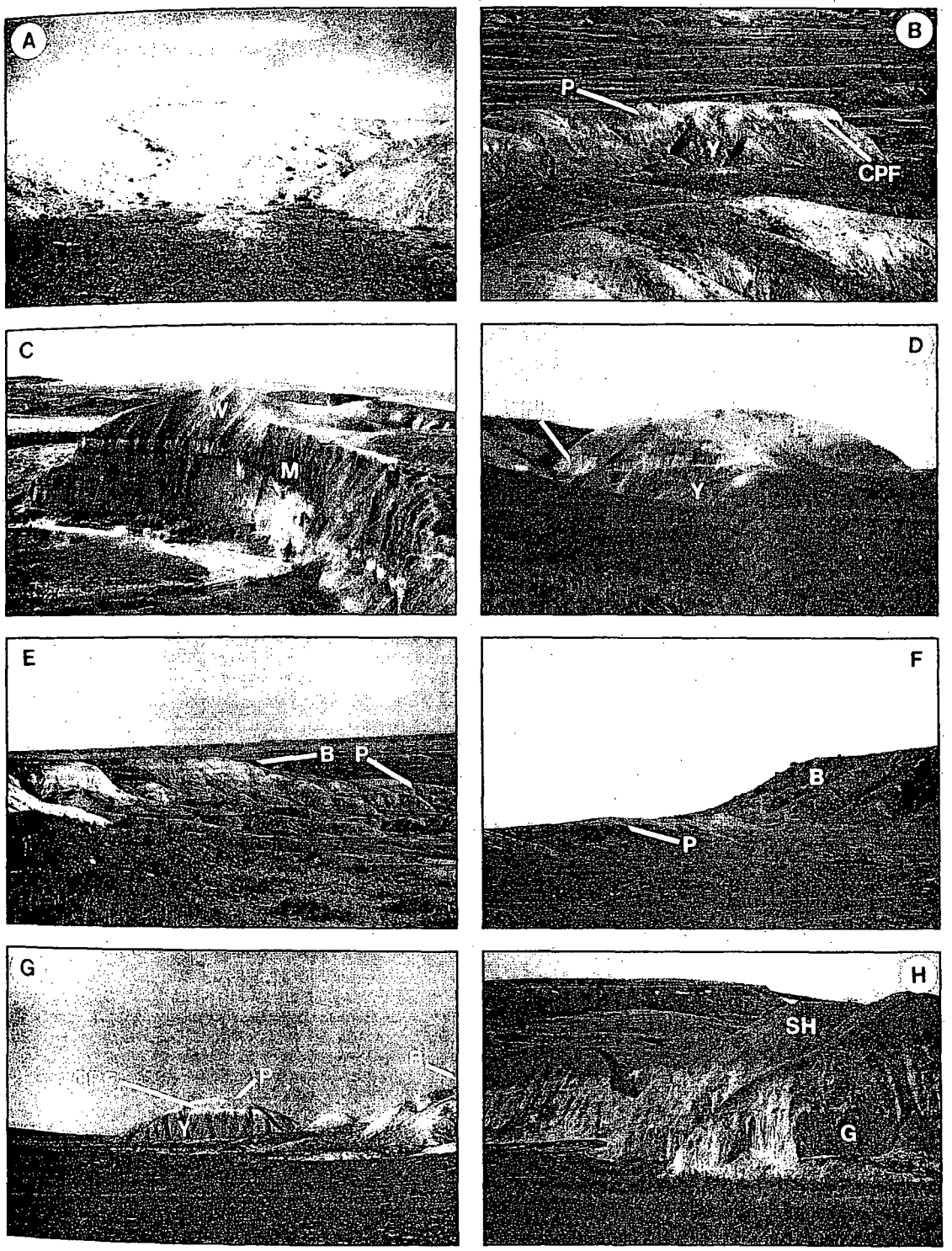
Figure 4. Measured sections of the Pierre Shale and Chamberlain Pass Formation (Interior Paleosol Complex of Schultz and Stout, 1955) in northwest Nebraska showing the pedogenic characteristics of the Yellow Mounds Paleosol Equivalent and Interior Paleosol Equivalent. Measured sections correspond to locations in Figures 3B and 5A,B. Color designations based on Munsell Soil Color Charts (1975).

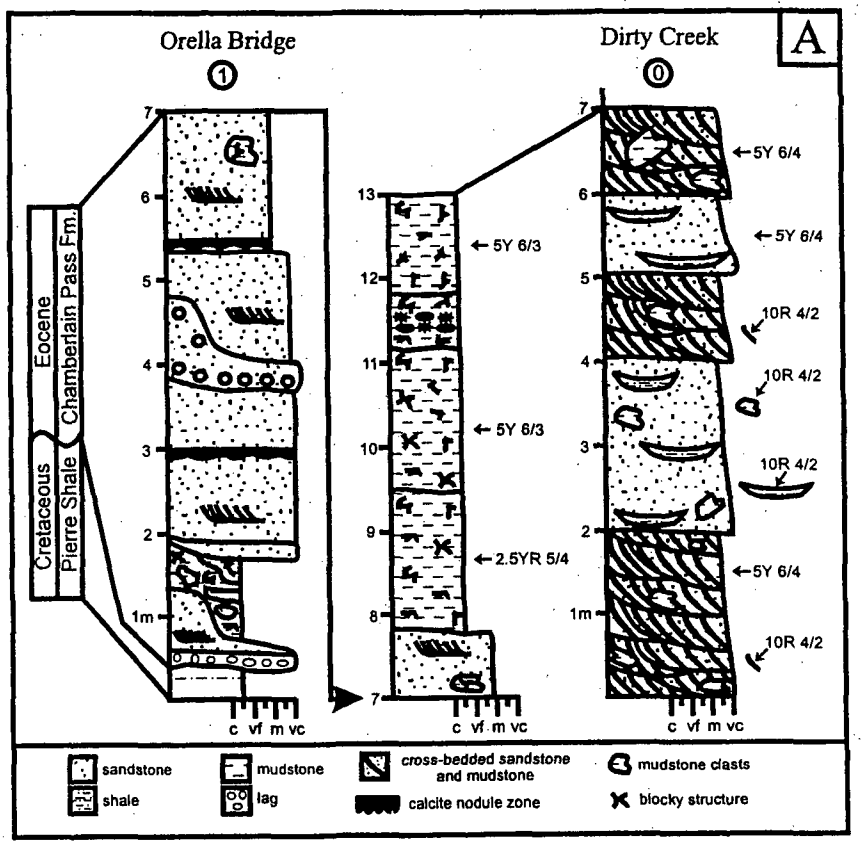
Equivalent and Weta *Paleosol Equivalent*), to establish correlations to South Dakota.

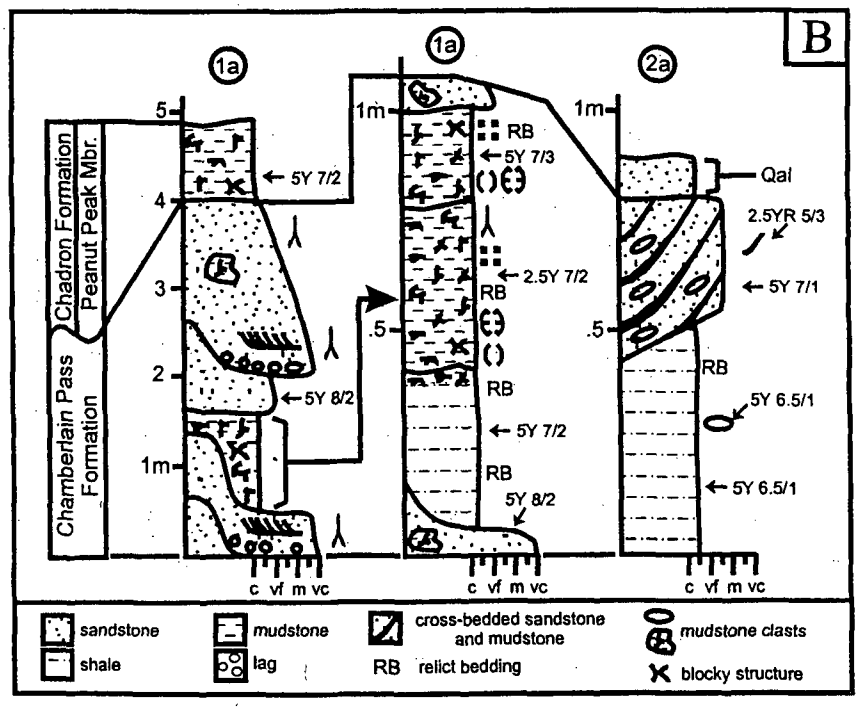
In northwestern Nebraska, the Interior Paleosol Complex (Schultz and Stout, 1955) is composed of two separate paleosols that developed on different parent materials (Fig. 4). The lower paleosol (Yellow Mounds Paleosol Equivalent) developed on the Cretaceous Pierre Shale and altered the normally black shale to bright yellow, purple, lavender, and orange (Fig. 5A). Pedogenic alteration gradually decreases downward and gives way to unaltered black shale. Pedogenic features are common, including soil horizons, root traces, argillans, concretions, and nodules. In places (Fig. 5C), the Yellow Mounds Paleosol Equivalent has been completely removed by rivers of a younger, overlying fluvial system, sometimes resulting in blocks of altered Pierre Shale being incorporated into the channel sandstone facies of this younger fluvial system.

The upper paleosol (Interior Paleosol Equivalent) developed on the channel sandstone and overbank mudstone facies of a distinct, post-Pierre/pre-Chadron fluvial unit, the Chamberlain Pass Formation (Figs. 5A–D, 6A). The lower contact of the Chamberlain Pass Formation with the Yellow Mounds Paleosol Equivalent is unconformable, and easily recognized by a change in color and lithology from the underlying black or bright yellow Pierre Shale

Figure 5. Photographs of the Yellow Mounds Paleosol Equivalent (Y), Interior Paleosol Equivalent (I), Weta Paleosol Equivalent (W), Chamberlain Pass Formation (CPF), and the Peanut Peak (P) and Big Cottonwood Creek Members (B) of the Chadron Formation in eastern Wyoming, northwestern Nebraska, southwestern South Dakota and southwestern North Dakota. Yellow Mounds Paleosol Equivalent and Interior Paleosol Equivalent = the Interior Paleosol Complex of Schultz and Stout (1955). Peanut Peak Member = the lower part of the Chadron B of Schultz and Stout (1955). Big Cottonwood Creek Member = the upper potion of the Chadron B, Chadron C, and Orella A of Schultz and Stout (1955). A; Location 3 in Figure 3B. Rock pick in center of photograph is 90 cm long. B: Location 4 in Fig. 3B. C: Loc. 1 of Figs. 3B and 6A. The channel sandstone facies of the Chamberlain Pass Formation (Chadron A of Schultz and Stout, 1955) rests on the Pierre Shale (covered). The channel sandstone is overlain by 5YR 6/4 (light reddish-brown) mudstone (M) that formed a clay plug within an abandoned meander loop. D; Location 6 of Figures 3B and 6C. E; Location 4 of Figure 3B. F; Locations 7 and 8 of Figure 3B. G; Exposures at Location C in Figure 1 near Lance Creek, Wyoming. Outcrop is located at the intersection of Highways 270 and 272, ~4.8 km (3 mi) northeast of Lance Creek. H; Exposures at Location L in Figure 1 near Dickinson, North Dakota. Outcrop located in the E1/2 sec. 7, T. 137 N., R. 97 W., New England NW Quadrangle. G = Golden Valley Formation, SH = South Heart Member of the Chadron Formation (Murphy et al., 1993).


A
B
P
Y
CPF
C
W
M
D
Y
E
B
P
F
B
P
G
P
Y
H
SH
G


A
Orella Bridge
Dirty Creek
Eocene
Chamberlain Pass Fm.
Cretaceous
Pierre Shale
5Y 6/4
5Y 6/3
10R 4/2
2.5YR 5/4
c vf m vc
sandstone
shale
mudstone
lag
cross-bedded sandstone and mudstone
calcite nodule zone
mudstone clasts
blocky structure


B
Chadron Formation
Peanut Peak Mbr.
Chamberlain Pass Formation
5Y 7/2
5Y 8/2
RB
5Y 7/3
2.5Y 7/2
Qal
2.5YR 5/3
5Y 7/1
5Y 6.5/1
c vf m vc
sandstone
shale
mudstone
lag
cross-bedded sandstone and mudstone
RB relict bedding
mudstone clasts
blocky structure

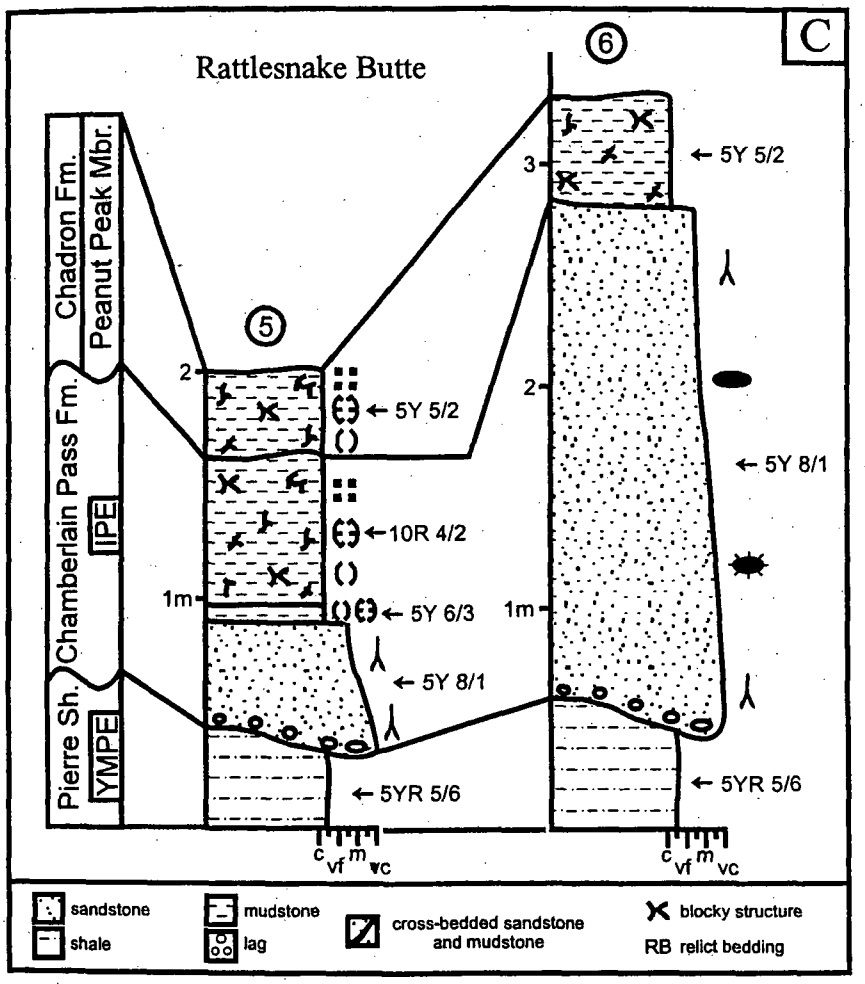


Figure 6 (on this and facing page). Detailed measured sections of various outcrops in the Toadstool Park area showing pedogenic structures and lithologic characteristics of stratigraphic units. A; Measured sections of the Orella Bridge exposure of Figures 3B and 5C and the Dirty Creek exposure. The Dirty Creek measured section is located in the NW1/4SW1/4NW1/4 sec. 13, T. 33 N., R. 53 W., Roundtop 7.5′ Quadrangle. The Dirty Creek Section is notable for its large amount of 10R 4/2 mudstone clasts, presumably derived from bank failure, and cross-bedded sands and 10R 4/2 muds. B; Section 1A is associated with the Orella Bridge exposures of Figures 3B and 5C and is located ~1.6 km to the west in the SW1/4SW1/4 sec. 35, T. 34 N., R. 54 W., Five Points 7.5′ Quadrangle. Section 2a is within the Big Cottonwood Creek Drainage, ~0.8 km to the northwest of Section 2 in Figure 3B, within the SW1/4NE1/4 sec. 4, T. 33 N., R. 52 W., Roundtop 7.5′ Quadrangle. C; Sections 5 and 6 of Figures 3B and 5D at Rattlesnake Butte. See Figure 4 for additional explanations of lithologic and pedogenic symbols.

to white channel sandstone and red overbank mudstone of the Chamberlain Pass Formation (Fig. 5A–D). The channel sandstone lithofacies of the Chamberlain Pass Formation is composed of as much as 8 m (26 ft) of yellowish, pale olive, and white, medium- to coarse-grained, trough and tabular cross-bedded sandstones commonly marked by a basal lag (Figs. 5B–D, 6A). Sandstone bodies are multistoried, showing rough fining-upward sequences. Although lateral accretion surfaces have not been identified, the ancient rivers could be considered "meandering" in the sense that they show evidence of lateral migration, including bank failure due to undercutting, general fining upward with eventual pedogenic modification of cohesive overbank deposits, and oxbow lakes that were filled to form clay plugs. Vertical accretion is suggested by the superposition of channel sandstone over mudstone and mudstone over sandstone (Fig. 3B). The channel sandstone facies occasionally contains thin, lenticular mudstone bodies that have been modified by pedogenesis (Fig. 6B).

The overbank mudstone lithofacies of the Chamberlain Pass Formation is composed predominantly of massive mudstone that has been overprinted by pedogenesis. Occasional silty yellowish-olive mudstone and siltstone units constitute a proximal overbank facies ranging as much as 3 m (9.8 ft) in thickness. The distal overbank mudstone lithofacies ranges in thickness from 0.8 to 1.8 m (2.6–6), and is easily recognized as a reddish stripe that overlies the yellow and orange Yellow Mounds Paleosol Equivalent (Figs. 4, 5A). Proximal overbank mudstones were subjected to different pedogenic processes than distal overbank areas, and sometimes contain Fe-oxide concretions and nodules resulting from alternating wet and dry soil conditions (Weta Paleosol Equivalent).

The upper contact of the Chamberlain Pass Formation/ Interior Paleosol Equivalent with the overlying Chadron Formation is unconformable and easily recognized by a change in color and lithology. The Chamberlain Pass Formation is composed of yellowish, pale olive, and white channel sandstone and the red and yellowish-olive mudstone. This sharply contrasts with the bluish-green and gray hummocky mudstone of the overlying Chadron Formation (Figs. 5B,D, 6B,C). Depending on the proximity to Chamberlain Pass Formation channels, the upper contact is sometimes very sandy, possibly as the result of incorporation of channel sands of the Chamberlain Pass Formation into the overlying Chadron Formation.

Measured sections of the "Interior Paleosol Complex" in northwestern Nebraska. Recognition and differentiation of both periods of pedogenic modification and their associated lithostratigraphic units are straightforward. Two measured sections (sections 3 and 4 of Fig. 3B) are described here (Fig. 4) to establish the validity of paleosol differentiation, lithostratigraphic divisions, and correlations to South Dakota. Colors are based on Munsell Soil Color Charts (1975). All colors are from fresh, excavated samples unless denoted otherwise.

Section 3 (Figs. 3B, 4, 5A) is marked at its base by modified Pierre Shale (Yellow Mounds Paleosol Equivalent). The Yellow Mounds Paleosol Equivalent is colored 10YR 6/6 (dark yellowish-orange), 5YR 5/6 (light brown), and 5Y 7/6 (moderate yellow). Pedogenic features include argillans coating peds and lining fractures within the soil, and slickensided argillans possibly formed by lithostatic compaction, or the expansion and contraction of the soil profile by wetting and drying. This zone of pedogenic alteration shows relict bedding from the original depositional fabric of the Pierre Shale. Progressing upward, the Yellow Mounds Paleosol Equivalent is replaced by the pedogenically altered overbank mudstone facies of the Chamberlain Pass Formation. The change is gradual over several centimeters and is recognized as reddish argillans (5R 4/2, grayish-red) from the overlying paleosol that coat, and eventually surround, fragments of the underlying Yellow Mounds Paleosol Equivalent.

The overlying paleosol (Interior Paleosol Equivalent) of the Chamberlain Pass Formation contains red (5R 4/2, grayish-red) platy peds with argillans and slickensided argillans of the same color, 0.5 to 2 mm in diameter mottles (5GY 7/4; moderate yellow green) mottles that are sometimes surrounded by 10YR 6/6 (dark yellowish-orange) haloes. These mottles likely represent drab-haloed root traces as defined by Retallack (1983). Some mottles are large, as much as 10 cm long by 1 to 3 cm wide. Some mottles are reversed in color, with the 10YR 6/6 (dark yellowish-orange) color surrounded by the 5GY 7/4 (moderate yellow-green). Progressing upward, peds become subangular blocky with the same argillans, slickensided argillans, and mottles seen lower in the Interior Paleosol Equivalent. The top of the 5R 4/2 (grayish-red) part of the Interior Paleosol Equivalent is marked by a transitional, but wavy, contact over 2 cm to a 13-cm-thick 10Y 6/2 (pale olive) claystone.

This overlying pale olive claystone also contains pedogenic features, including 1-to-3 mm 10YR 6/6 (dark yellowish-orange) mottles, platy peds, and slickensided argillans. Inclusions of the underlying 5R 4/2 (grayish-red) mudstone are common within this zone. The top of this paleosol is sharply truncated by a channel sandstone of the Chamberlain Pass Formation. The contact is marked by a lag dominated by quartz and quartzite pebbles and cobbles, and numerous claystone and mudstone clasts that give this basal lag a rusty iron color. The clay and mud chip lag is a combination of 5R 3/4 (dusky red), 10YR 6/6 (dark orangish-yellow), and 5YR 5/6 (light brown) fragments that represent erosion of the underlying Yellow Mounds Paleosol Equivalent and pedogenically modified mudstone of the Chamberlain Pass Formation. This basal lag is 7 cm thick and changes upward into a 5Y 7/2 (yellowish-gray) clay chip dominated zone before giving way to a 77-cm-thick, coarse- to fine-grained 5Y 6/4 (dusky yellow) channel sandstone. The sandstone is capped by 5Y 6/2 (light olive-gray) mudstone of the Chadron Formation. The contact is gradational, likely reflecting the incorporation of Chamberlain Pass Formation sediments into the base of the overlying Chadron Formation.

Section 4 is located ~1 km east of section 3 (Figs. 3B, 4, 5B). The base of section 4 is also marked by the Yellow Mounds Paleosol Equivalent (Figs. 4, 5B). Progressing upward, the color of the Yellow Mounds Paleosol Equivalent remains constant as a

mixture of 5Y 7.5/1 (light gray) and 5YR 6/2 (pinkish-gray) shale. Pedogenic features are numerous, including angular blocky, and platy peds, 5-by-1-cm iron mottles (7.5YR 5/6, strong brown) that parallel relict bedding, argillans and slightly slickensided argillans, and 10YR 6/6 (dark orangish-yellow) root traces as much as 5 cm wide that branch downward and are sometimes filled with gypsum. Closer to the contact, 10R 4/3 (weak red) slickensided argillans from the overlying paleosol (Interior Paleosol Equivalent) have formed within the Yellow Mounds Paleosol Equivalent. The contact with the overlying Chamberlain Pass Formation is sharp and erosive.

The base of the Chamberlain Pass Formation at this locality is marked by a band of large, cherty cobbles as much as 8 cm in diameter. Root traces from the underlying Yellow Mounds Paleosol Equivalent are truncated by this contact. Root traces from the overlying Interior Paleosol Equivalent extend downward into this band of cobbles, are deflected around the cobbles, and eventually penetrate into the underlying Yellow Mounds Paleosol Equivalent. The Chamberlain Pass Formation changes upward from a 47-cm-thick, 10R 6/3 (pale red) clay-rich sandstone to a 54-cm-thick, 10R 6/3 (pale red) silty claystone. The upper 7 cm of the sandstone is brecciated, likely due to pedo-brecciation. Pedogenic features are numerous, including platy and subangular peds, argillans and slickensided argillans, root traces identical in size and color to those from the Yellow Mounds Paleosol Equivalent, and 10R 4/2 (weak red) clay-filled root traces. A large amount of claystone breccia is also present in this interval, but decreases upward. The breccia ranges in size from 1 to 15 mm and is 2.5YR 6/2 (pale red) and 10R 4/3 (weak red) in color. The breccia is likely derived from the lateral migration and cutbank erosion of cohesive overbank sediments by rivers of the Chamberlain Pass Formation. The top contact of the Chamberlain Pass Formation is sharply truncated by an erosion surface and is marked by a thin band of fluvially transported quartz and quartzite cobbles. Root traces within Chamberlain Pass Formation are truncated at the upper contact. The Chamberlain Pass Formation is overlain by 5Y 6/2 (light olive-gray) mudstone of the Chadron Formation. Chadron Formation mudstone was also altered by pedogenesis, as indicated by the presence of argillans, slickensided argillans, platy and angular blocky peds, and clay-filled (10YR 5.5/1, gray) root traces.

Age. The age of the Yellow Mounds Paleosol Equivalent is constrained by the age of overlying units across the northern Great Plains. Pettyjohn (1966) described a regionally extensive zone of alteration (the "Eocene Paleosol") across the northern Great Plains that had formed on units as old as the Early Cretaceous Skull Creek Shale, and as young as the Slim Buttes Formation in northwestern South Dakota. The Eocene Paleosol is overlain by the White River Group across the study area. Although Pettyjohn (1966) acknowledged the likely time-transgressive nature of the Eocene Paleosol, he considered the period of soil formation to be predominantly of Eocene age. Retallack (1983) classified the Yellow Mounds Paleosol Series as an Ultisol. Modern soils of a similar nature are estimated to require at least 10,000 yr, to possibly several million years, to form (Cady and Daniels, 1968; Buol et al., 1973; Birkeland, 1974; Soil Survey Staff, 1975). No direct age determination for the duration of pedogenesis that formed the Yellow Mounds Paleosol Equivalent in the Toadstool Park area can be made, other than it is post–late Cretaceous to late Eocene in age.

No radiometric dates have been obtained from the Chamberlain Pass Formation in South Dakota or Nebraska, although exposures of the Chamberlain Pass Formation immediately west of White Clay, Nebraska are overlain by a white bentonite (location 9 of Fig. 3B). Careful searching may yield a pocket of datable material. The best estimates come from comparisons to fossils recovered from lithologic units of the same stratigraphic position and general lithology as the Chamberlain Pass Formation, and age constraints from overlying deposits. Vertebrate fossils from the Chamberlain Pass Formation are rare, but those that have been identified indicate a Chadronian "age" (Wood et al., 1941; Emry et al, 1987). Vertebrate fossils from the Chamberlain Pass Formation include a fragment of *Trigonias* (Clark et al., 1967) from the channel sandstone facies near Kadoka, South Dakota (loc. G of Fig. 3A), Chadronian vertebrates from beds of similar lithology and stratigraphic position in the East Short Pine Hills in southwestern Harding County, South Dakota (Lillegraven, 1970), and a brontothere jaw from the channel sandstone facies of the Chamberlain Pass Formation (loc. 1 of Figs. 3B, 5C) in northwestern Nebraska (Vondra, 1958b). This brontothere (UNSM field nos. 205-55/206-55) was recently located in the University of Nebraska State Museum collections and prepared (see LaGarry et al., 1996), and includes well-preserved right and left mandibles of a small individual and several isolated teeth. The jaws are unique in that the p2–m3 length of 29 cm is larger than that reported for *Duchesneodus* (see Lucas and Schoch, 1989), but smaller than that reported for typical medial Chadronian brontotheres, such as *Brontops* (see Mader, 1989). Although measurements and comparisons are preliminary, this brontothere may represent an early or medial Chadronian species.

This evidence, combined with the recently published age revision of the Chadronian North American Land-Mammal "age" (Swisher and Prothero, 1990), restricts the Chamberlain Pass Formation to the late Eocene. In addition to Vondra's (1958b) brontothere mandibles, LaGarry et al. (1996) have recovered a suite of vertebrate remains from the overbank and channel sandstone facies of the Chamberlain Pass Formation in northwestern Nebraska. The remains are fragmentary, but once identified may help to refine the age of the Chamberlain Pass Formation in northwestern Nebraska.

Chadron Formation of northwestern Nebraska

Lower member. The Chadron Formation of northwestern Nebraska is composed of two distinct members that are recognized on the basis of lithology, color, and erosional relief (Fig. 5E). The lower member, to which the name *Peanut Peak Member* is applied, is composed of as much as 8.65 m (28 ft) of

predominantly bluish-green and gray hummocky mudstone with occasional pockets of red, green, and yellow mudstone in upper parts (Figs. 3B, 4, 5A). This unit is smectite-rich, weathers into haystack-shaped hills and slopes with a popcorn-like surface, and overlies the Chamberlain Pass Formation.

The upper contact of the Peanut Peak Member with the overlying remainder of the Chadron Formation is variable. In places it is unconformable, such as along the east end of Sugarloaf Road (location 4 of Figs. 3B, 5E), where the remainder of the overlying Chadron Formation fills depressions/paleovalleys within the Peanut Peak Member (Terry and LaGarry, this volume). In contrast, at Toadstool Park, the contact of the Peanut Peak Member with the overlying remainder of the Chadron Formation is intertonguing, with lenses of the Peanut Peak Member visible in the overlying upper member and lenses of the overlying upper member visible within the Peanut Peak Member (Terry and LaGarry, this volume). The Peanut Peak Member differs from the overlying remainder of the Chadron Formation in that it is composed of smectite-rich mudstone and claystone, weathers into bluish-green and gray hummocky hills and slopes, is less variegated in color, has less silt, and contains none of the "purplish-white layers" of Schultz and Stout (1955) (Fig. 5E).

I extend the usage of the term Peanut Peak Member to include the undifferentiated, bluish-green and gray hummocky mudstone of the Chadron Formation outside of the Red River Valley, South Dakota, of Clark et al. (1967), and its lithologic equivalents in northwestern Nebraska, for several reasons: (1) It serves no purpose to introduce a new lithostratigraphic name for these strata. The term Peanut Peak Member has already been employed by Clark (1954) and Clark et al. (1967) for greenish, gray, tan, and orange hummocky mudstone within the upper part of the Red River Valley in the Big Badlands (Fig. 2E,G). (2) By visually tracing this unit from the Big Badlands of South Dakota into northwestern Nebraska, the Peanut Peak Member (as used herein) is no longer a geographically isolated unit within the Red River Valley. This sense of geographic restriction and dissimilarity with the majority of the Chadron Formation outside the Red River Valley (undifferentiated Chadron Formation) led Harksen and Macdonald (1969a,b) to reject the divisions of the Chadron Formation established by Clark (1937, 1954) and to suggest a reference section for the Chadron Formation located 4.83 km (3 mi) southeast of the type geologic locality for the "*Titanotherium* Zone" as established by Osborn (1929). Harksen and Macdonald (1969a) believed that this section better represented the general lithology of the Chadron Formation. (3) This package of undifferentiated bluish-green and gray hummocky mudstone is overlain in northwestern Nebraska by the remainder of the Chadron Formation (Terry and LaGarry, this volume), which is itself a distinct lithostratigraphic unit (Fig. 5E). Thus, this package of bluish-green and gray hummocky mudstone that overlies the Chamberlain Pass Formation in both the Big Badlands of South Dakota and northwestern Nebraska must have a formal lithostratigraphic designation (see NACSN, 1983) other than "undifferentiated Chadron Formation" (Figs. 2A–D, 5A–E).

Age. The Peanut Peak Member in northwestern Nebraska has not been directly dated by radiometric means. The best estimates are based on faunal content and magnetostratigraphic correlation. Based on its fossil content, the Peanut Peak Member is Chadronian, as are the underlying Chamberlain Pass Formation and overlying remainder of the Chadron Formation. Based on paleomagnetic data, the Peanut Peak Member of the Chadron Formation in northwestern Nebraska falls within Chron 15 (Prothero and Swisher, 1992). This evidence, combined with the recently published age revision of the Chadronian NALMA (Swisher and Prothero, 1990), restricts the Peanut Peak Member of the Chadron Formation in northwestern Nebraska to late Eocene.

Upper member. The overlying remainder (upper member) of the Chadron Formation in northwestern Nebraska is a siltier, cliff-forming, and variegated unit of pedogenically modified claystone, silty claystone, and siltstone with occasional isolated channel sandstone bodies (Terry and LaGarry, 1994, this volume; Terry, 1995; Terry et al., 1995) (Fig. 5E). Terry and LaGarry (this volume) redefine this remaining upper part of the Chadron Formation in northwestern Nebraska as the *Big Cottonwood Creek Member*. The Big Cottonwood Creek Member of the Chadron Formation contains the various purplish-white layers of Schultz and Stout (1955); lighter colored bands of rock composed of gypsum, volcanic ash, or limestone that stand out in contrast to the surrounding badlands. Some purplish-white layers represent paleosol profiles. Descriptions and usage of the various purplish-white layers in Nebraska are discussed by Terry and LaGarry elsewhere in this volume. The contact of the Big Cottonwood Creek Member of the Chadron Formation with the overlying Brule Formation in northwestern Nebraska is marked by a change from variegated silty mudstone and claystone of the Big Cottonwood Creek Member of the Chadron Formation to tan and brown clayey siltstone, siltstone, sheet sandstone, and channel sandstone of the Orella Member of the Brule Formation. This Chadron/Brule Formation boundary differs greatly from previous stratigraphic concepts for units in northwestern Nebraska, and is discussed by Terry and LaGarry (this volume) and LaGarry (this volume).

I have traced the Big Cottonwood Creek Member northeast into South Dakota to the town of Oelrichs with no change in stratigraphic relationships (Fig. 5F). At Oelrichs, the Big Cottonwood Creek Member interfingers with lacustrine limestone similar to those described by Welzenbach (1992), Welzenbach and Evans (1992), and Evans and Welzenbach (this volume). These limestones (Fig. 2H) are within the upper part of the Chadron Formation (= Peanut Peak Member as defined in this chapter) in the Big Badlands of South Dakota (Welzenbach, 1992; Welzenbach and Evans, 1992), and occasionally mark the boundary between the Chadron and Brule Formations (D. O. Terry, 1995, unpublished data). The siltier, variegated, and cliff-forming Big Cottonwood Creek Member of the Chadron Formation in northwestern Nebraska is absent in the Badlands National Park area

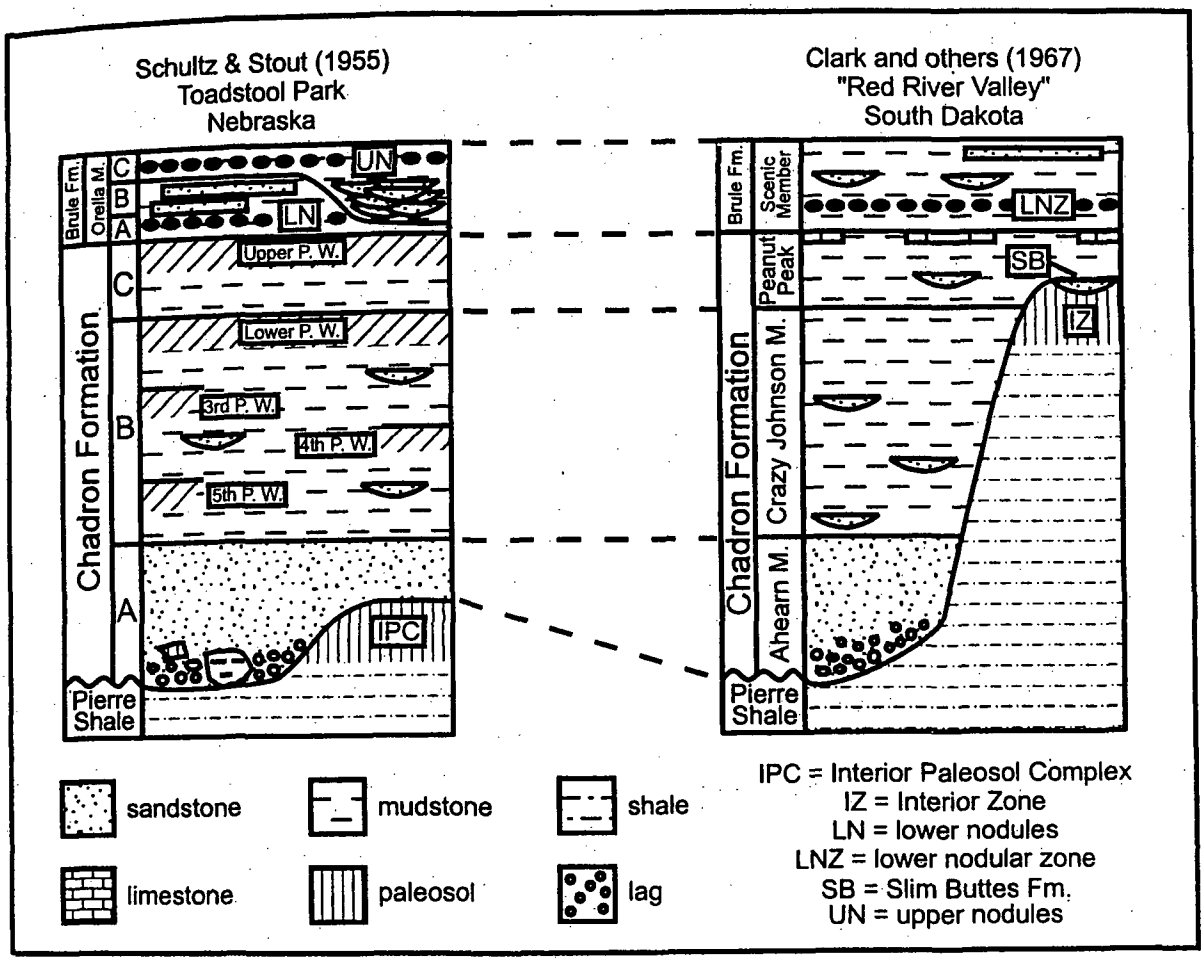


Figure 7. Stratigraphic terminology of Schultz and Stout (1955) for the Chadron Formation of northwestern Nebraska and their suggested correlations to members of the Chadron Formation within the Red River Valley of Clark et al. (1967) in the Big Badlands of South Dakota.

(Fig. 2A–E). Within Badlands National Park the Peanut Peak Member of the Chadron Formation is overlain by Bump's (1956) Scenic Member of the Brule Formation (Fig. 2A).

Comparisons with previous stratigraphic hierarchies and correlations

The stratigraphic terminology of Schultz and Stout (1955) for members of the Chadron Formation in northwestern Nebraska, although never formally defined in type sections or assigned names other than A, B, C, and/or "lower," "middle," "upper," has become entrenched in the literature (Fig. 7). This terminology and the suggested regional correlations associated with it have become commonly cited and accepted (Schultz et al., 1955; Leonard, 1957; Vondra, 1958a, 1960; Swenson, 1959; Harvey, 1960; Schultz and Falkenbach, 1968; Stone, 1972; Singler and Picard, 1979, 1980, 1981; Schultz and Stout, 1980; Retallack, 1983; Martin, 1985; Swinehart et al., 1985; Prothero and Swisher, 1992; Murphy et al., 1993).

Within the stratigraphic paradigm of Schultz and Stout (1955), the White River Group of northwestern Nebraska rests on the Interior Paleosol Complex (Fig. 7), a zone of pedogenic alteration that modified the Cretaceous Pierre Shale from its usual black color into a bright yellow zone capped by a reddish horizon. The first deposits of the White River Group are represented by the basal sands and conglomerates of the "Chadron A," which in turn are overlain by mudstones of the "Chadron B" and "Chadron C," respectively. The boundaries of the Chadron B and Chadron C are defined by purplish-white layers. The contact between the Chadron B and Chadron C is marked by the "lower" or "second purplish-white layer." The top of the Chadron Forma-

tion (top of the Chadron C) is marked by the "upper purplish-white layer," a regionally extensive volcanic ash. The Chadron Formation is overlain by the Orella Member of the Brule Formation. The lowest part of the Orella Member, which Schultz and Stout (1955) have defined as the "Orella A," directly overlies the upper purplish-white layer. These divisions were established, and the boundaries chosen, as a convenient guide for the stratigraphic documentation of collected fossils (Schultz and Stout, 1955).

Schultz and Stout (1955) proposed that their stratigraphic divisions of the Chadron Formation in northwestern Nebraska could be directly correlated to Clark's (1937, 1954) three-fold division of the Chadron Formation in the Big Badlands of South Dakota (Fig. 7). Namely, the Chadron A, B, and C of Nebraska correlated with the Ahearn, Crazy Johnson, and Peanut Peak Members of South Dakota, respectively. Schultz and Stout (1955) also considered the Chadron A to be equivalent to the "Yoder Beds" of Schlaikjer (1935) and Kihm (1987) near Torrington, Wyoming (Fig. 1), and the Interior Paleosol Complex of northwestern Nebraska to be equivalent to the Interior Zone of Ward (1922) near Interior, South Dakota.

Schultz and Stout's (1955) stratigraphic divisions of the Chadron Formation in northwestern Nebraska were based on event beds such as volcanic ashes and paleosols. Several new developments invalidate these divisions and the proposed correlations to units in South Dakota: (1) The use of event beds is in direct violation of the current North American Stratigraphic Code (NACSN, 1983) for recognition of lithostratigraphic units (e.g., groups, formations, and members). (2) Retallack (1983) recognized that the Interior Zone of South Dakota is composed of two separate paleosols, the Yellow Mounds and Interior Paleosol Series (Fig. 2A). (3) The Chamberlain Pass Formation of Evans and Terry (1994), Terry and Evans (1994), is now recognized in northwestern Nebraska and the Badlands of South Dakota (Figs. 2A–D, 3, 5A–D, 6). (4) The stratigraphic units of Schultz and Stout (1955) do not lithologically correlate to their suggested counterparts in the Big Badlands of South Dakota. (5) In addition, Schultz and Stout's (1955) units in Nebraska cannot be matched to rock units in South Dakota based on the principle of superposition, nor can the purplish-white layers be visually traced into the Big Badlands of South Dakota.

Using my newly proposed lithostratigraphic hierarchy (Figs. 3B, 8, 9), the Interior Paleosol Complex of Schultz and Stout (1955) is composed of the pedogenically modified Pierre Shale and the pedogenically modified overbank mudstone of the Chamberlain Pass Formation, which corresponds to the Yellow Mounds Paleosol Equivalent and Interior Paleosol Equivalent,

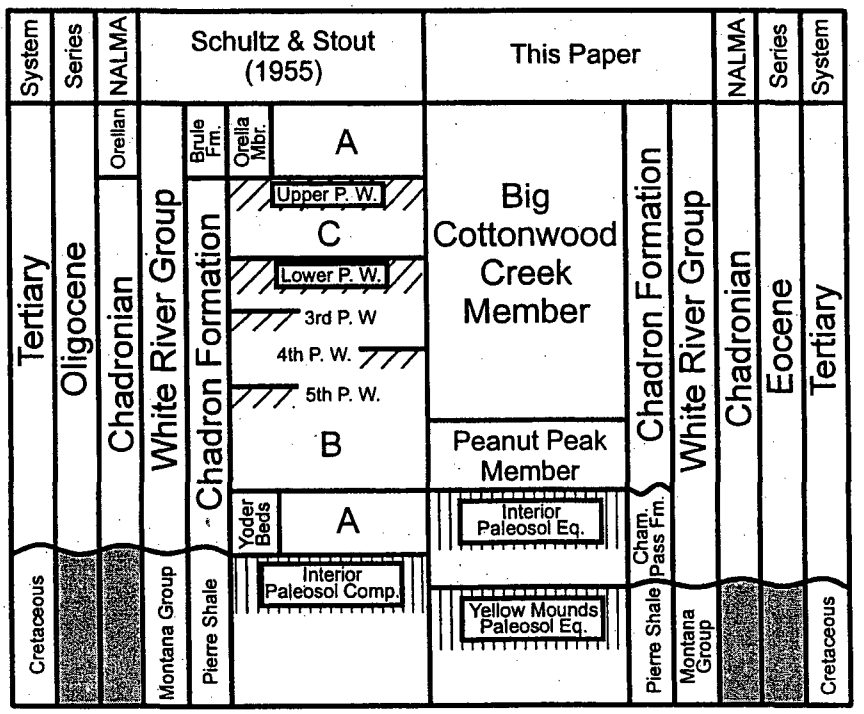


Figure 8. Proposed stratigraphic revisions for the lower part of the White River Group of northwestern Nebraska.

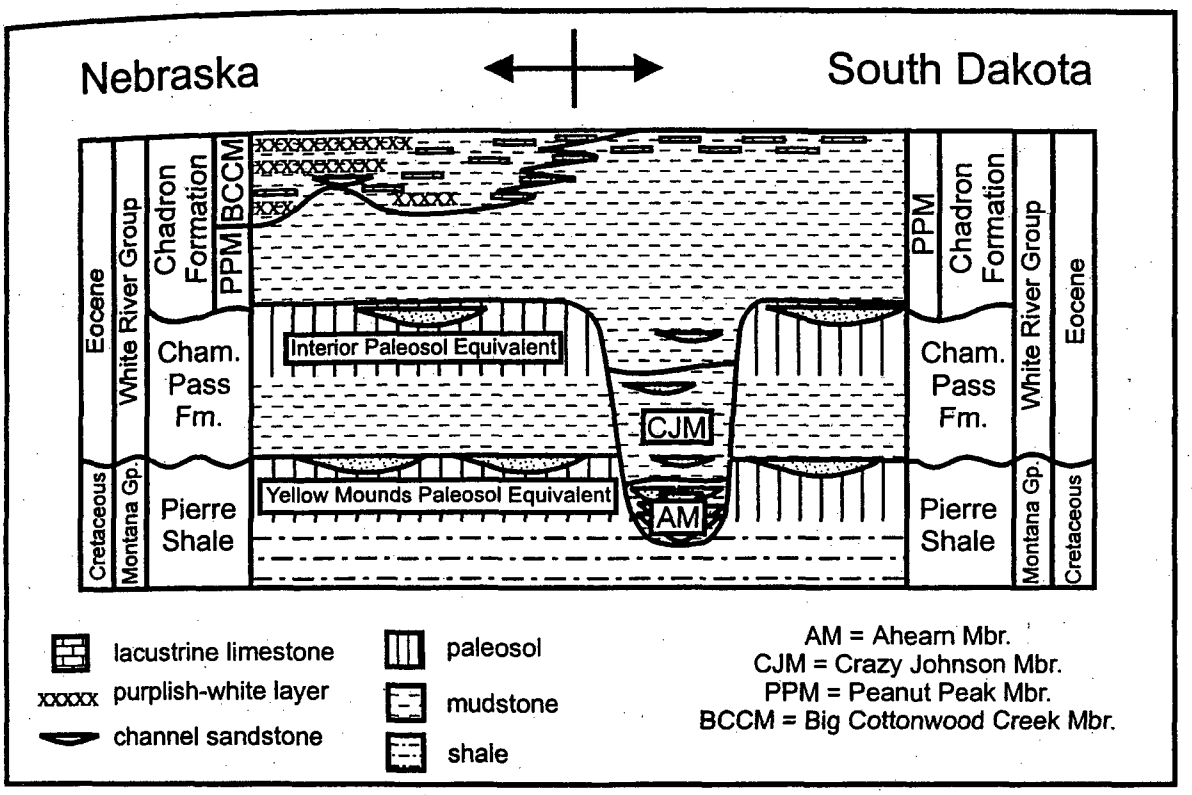


Figure 9. Diagram showing the proposed regional geographic relationships of the various members of the Chadron Formation between the Toadstool Park area of northwestern Nebraska and the Red River Valley of Clark et al. (1967) in the Big Badlands of South Dakota. Modified from Terry et al. (1995).

respectively (Figs. 2A, 5A,B). The Chadron A of Schultz and Stout (1955) corresponds to the channel sandstone facies of the Chamberlain Pass Formation (Figs. 2B–D, 5B–D). The lower part of the Chadron B, the bluish-green, smectite-rich hummocky mudstone, is equivalent to the Peanut Peak Member as defined in this chapter (Figs. 2A–G, 5A–E). The overlying siltier, variegated, and cliff-forming remainder of the Chadron B, Chadron C, and Orella A of the overlying Brule Formation is contained within a newly recognized, separate, and distinct lithostratigraphic unit that Terry and LaGarry (this volume) designate as the *Big Cottonwood Creek Member* of the Chadron Formation (Figs. 5E,F, 8, 9).

DISCUSSION

Regional correlations

Nebraska. The Yellow Mounds Paleosol Equivalent, Chamberlain Pass Formation, and Chadron Formation are not exposed south of the Pine Ridge Escarpment. Swinehart et al. (1985) were able to identify three subsurface depositional sequences within the White River Group based on a detailed analysis of ~11,600 electric logs of oil and gas wells and samples and logs from 500 others. The lowest of these three sequences encompassed the bottom of the Chadron Formation to the unconformity within the Orella Member of the Brule Formation. This "Chadron Sequence" rests upon an oxidized zone that is likely the same period of pedogenesis that produced the Interior Paleosol Complex (Yellow Mounds and Interior Paleosol Equivalents) in northwestern Nebraska and across the Great Plains. J. B. Swinehart (1994, personal communication) has observed cored sandstone in Sheridan County, Nebraska that is lithologically similar to the channel sandstone facies of the Chamberlain Pass Formation. This sandstone, along with the upper part of the subsurface oxidized zone, is likely equivalent to the Chamberlain Pass Formation.

Wyoming. Buttes near Lance Creek, Wyoming, display a yellowish zone, possibly the Yellow Mounds Paleosol Equivalent, overlain by a white sand body similar to the Chamberlain Pass Formation (Figs. 1, 5G, 10). This white sand body is in turn overlain by greenish-gray hummocky mudstones similar to the Peanut Peak Member. The remainder of the butte is composed of strata lithologically similar to the Big Cottonwood Creek Member of Terry and LaGarry (this volume) and Terry et al. (1995).

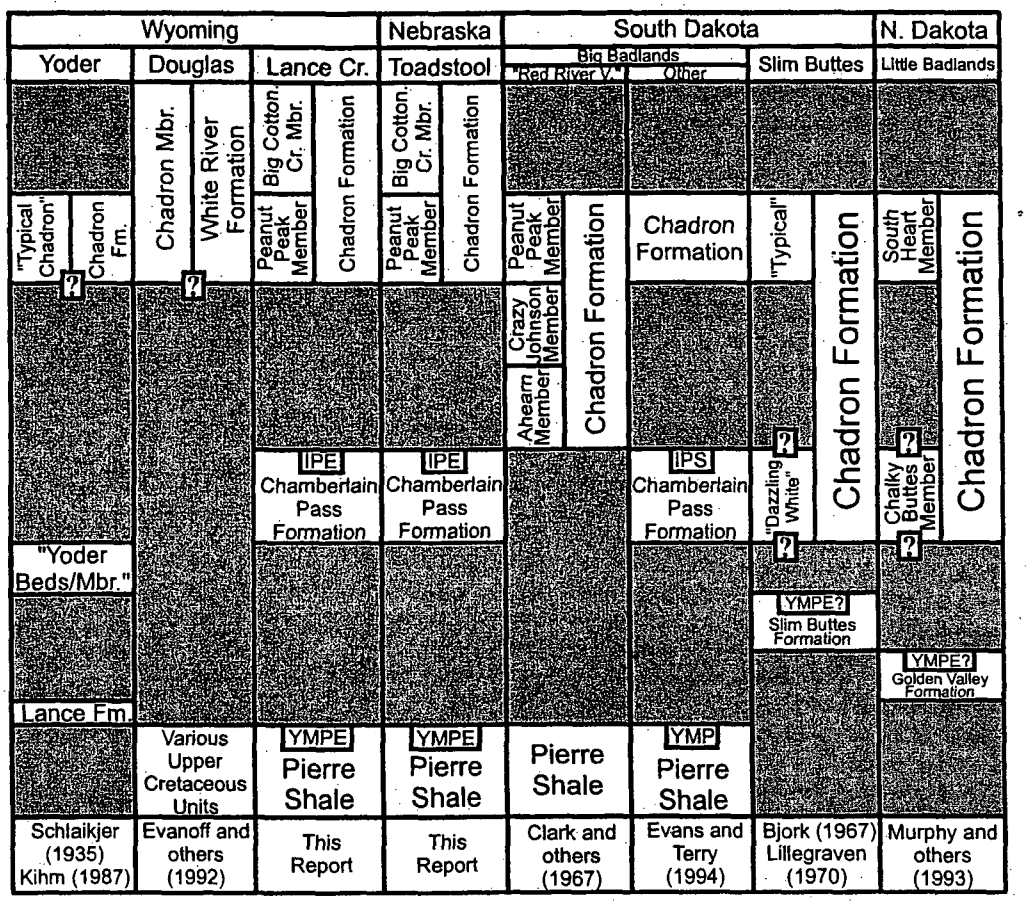


Figure 10. Proposed lithologic equivalency chart for the Chamberlain Pass Formation and the various members of the Chadron Formation in relation to underlying units and major periods of pedogenic modification. IPS and YMP = Interior and Yellow Mounds Paleosol Series of Retallack (1983), respectively; IPE = Interior Paleosol Equivalent; YMPE = Yellow Mounds Paleosol Equivalent.

These similarities were also noted by Schultz and Stout (1955) and Luebke (1964), but were classified and interpreted within the stratigraphic hierarchy developed for the Toadstool Park area by Schultz and Stout (1955).

At Yoder, Wyoming, I have observed beds lithologically similar to the Peanut Peak Member (Figs. 1, 10) overlying the Yoder Beds of Schlaikjer (1935) and Kihm (1987). I do not agree with the correlation of the Yoder Beds by Schultz and Stout (1955) to the Chadron A (channel sandstone facies of the Chamberlain Pass Formation) of northwestern Nebraska for several reasons. (1) First, the two deposits are composed of different lithologies. The Yoder beds are composed of finely bedded, coarse red sandstone, red and greenish sandy clays, and reddish-green clays with channels of intricately cross-bedded, wine-colored coarse gravels to fine sandstone (Schlaikjer, 1935). Channel sandstone of the Chamberlain Pass Formation is composed of blazing white to yellowish-white, coarse- to fine-grained, cross-bedded, and multistory sandstone bodies, commonly marked at the base by quartz-rich lag deposits. Reddish clay is present, but only as occasional streaks, or as a thick clay plug developed on top of the channel sandstone facies just north of Toadstool Park, Nebraska (Terry, 1991; Evans and Terry, 1994; Terry et al., 1995). (2) The Yoder Beds are areally restricted within a few miles of the Yoder and Veteran, Wyoming, area within a paleovalley formed in the underlying Cretaceous Lance Formation. Outside this area the Lance Formation is overlain by greenish-gray hummocky mudstone of the Chadron Formation (Kihm, 1987). The channel sandstone within the Chamberlain Pass Formation is part of an areally extensive and distinct fluvial lithostratigraphic unit. (3) The Yoder Beds are very fossiliferous, whereas the Chamberlain Pass Formation is almost devoid of fossils.

Correlations of the revised strata discussed herein to expo-

sures in the Douglas and Flagstaff Rim areas of central and east-central Wyoming have not yet been made (Fig. 1). The White River Group is labeled as a formation in Wyoming (Fig. 10) and contains the Chadron and Brule Members (Evanoff, 1990; Evanoff et al., 1992). The two members are separated by the widespread # 5 tuff (Evanoff et al., 1992). The Chadron Member is composed of green to brown clayey mudstone, typically nodular brown to tan sandy mudstone, thin sheet sandstone, and several thick ribbon sandstones (Evanoff et al., 1992). The overlying Brule Member is composed of nodular brown to tan sandy mudstone, massive tan sandy siltstone, rare ribbon sandstone, and rare conglomeratic sheet sandstone (Evanoff et al., 1992). The use of the names "Chadron" and "Brule" for these units is based on the slight lithologic difference between the two members and the similarities of these two units to the Chadron and Brule Formations in northwestern Nebraska (Evanoff et al., 1992). The ranking of these units as members is based on an occasional difficulty in distinguishing the two units in outcrop, and for their less distinct appearance as compared to Nebraska equivalents (Evanoff et al., 1992). Leonard (1957) measured numerous sections through the White River Group in the Douglas area and suggested correlations to Toadstool Park using the stratigraphic hierarchy of Schultz and Stout (1955). The validity of these correlations is unknown.

Northwestern South Dakota. Lithologically identical counterparts of the channel sandstone facies of the Chamberlain Pass Formation and the Peanut Peak Member of the Chadron Formation are visible in Harding County, South Dakota at Slim Buttes, 32 km (20 mi) east of Buffalo. Channel sandstone deposits are "dazzling white" in color, and rest on the golden brown, interbedded, and variegated sands, silt, and clays of the Eocene Slim Buttes Formation (Bjork, 1967). In contrast to Pettyjohn (1966), I am undecided as to the presence of the Yellow Mounds Paleosol Equivalent. I have observed iron oxide concretions in the same pedostratigraphic position as other Yellow Mounds Paleosol Equivalent outcrops, but I have not examined the Slim Buttes Formation for other evidence of pedogenic modification. The dazzling white sandstone is overlain by gray hummocky claystone, identical to the Peanut Peak Member, that Bjork (1967) lumped together with the white sandstone as the Chadron Formation (Fig. 10). Lillegraven (1970) used a different stratigraphic approach, dividing the Chadron Formation into the "golden brown," "dazzling white," and "typical Chadron" units (Fig. 10). I have not observed a reddish-colored overbank mudstone, similar to that within the Chamberlain Pass Formation, associated with the dazzling white sandstone at Slim Buttes. The contact between the dazzling white and typical Chadron is sometimes fractured, with gray clay filling the fractures (D. O. Terry, 1993, unpublished data). This is identical to the pedo-brecciation preserved at the contact of the Chamberlain Pass Formation and Peanut Peak Member of the Chadron Formation near Interior, South Dakota (Terry, 1991; Terry and Evans, 1994). Paleosols are preserved within the gray hummocky mudstone of the Chadron Formation at Slim Buttes (D. O. Terry, 1993, unpublished data). The Chadron Formation at Slim Buttes is overlain by the Brule Formation.

Southwestern North Dakota. The same stratigraphic succession at Slim Buttes, South Dakota is visible in the Little Badlands of southwestern North Dakota (Figs. 1, 5H). The Little Badlands near Dickinson, North Dakota display the general Eocene/Oligocene stratigraphy of this area (Fig. 10). The White River Group in the Little Badlands rests unconformably on the light brown to golden micaceous sandstone, siltstone, mudstone, and thin lignites of the Eocene Camels Butte Member of the Golden Valley Formation (Murphy et al., 1993). The Camels Butte Member was apparently subjected to the same period of pre–White River Group leaching that formed the Interior Paleosol Complex (Yellow Mounds Paleosol Equivalent) in the Big Badlands of South Dakota (Hickey, 1977) (Fig. 5H). This weathering changed the normally buff and yellow colors of the Camels Butte Member to bright orange, yellow, and white (Hickey, 1977).

The Camels Butte Member in this area is overlain by a chalky white, cross-bedded conglomeratic sandstone and sandy mudstone that Murphy et al. (1993) designate as the Chalky Buttes Member of the Chadron Formation. This unit, which ranges from 3 to 6 m (10–20 ft) in thickness in the Little Badlands, is similar stratigraphically and lithologically to the dazzling white sandstones at Slim Buttes and the channel sandstone facies of the Chamberlain Pass Formation in the Big Badlands of South Dakota. The Chalky Buttes Member is overlain by smectitic, gray to brown, hummocky mudstone and occasional limestone lenses that Murphy et al. (1993) designated as the "South Heart Member" of the Chadron Formation (Fig. 5H). The South Heart Member, which is 3 to 9 m (10–30 ft) thick in the Little Badlands, is identical in lithology and stratigraphic position to the "typical Chadron" at Slim Buttes and the Peanut Peak Member (as per this chapter) of the Chadron Formation in the Big Badlands of South Dakota (Fig. 10).

Great Plains geologic history

As initially envisioned by Evans and Terry (1994), the Chamberlain Pass Formation was contained within an asymmetric basin marked on its southern edge by the Pine Ridge fault zone and by the Sage Ridge fault on the northern edge. The Chamberlain Pass Formation was thought to represent the fluvial deposits of a drainage system headed in the Black Hills. As I have demonstrated herein, the Chamberlain Pass Formation extends beyond this basin. Its overall extent is unknown at this time, but I have seen similar deposits as far west as Lance Creek, Wyoming, east to White Clay, Nebraska, and north to Dickinson, North Dakota (Figs. 1, 3B, 5G,H). The Chamberlain Pass Formation may represent the deposits of a widespread Late Eocene fluvial system on the northern Great Plains.

The revision of the Chadron Formation is significant (see also Terry and LaGarry, this volume). Strata of the Peanut Peak Member are not geographically restricted within the Red River Valley of Clark et al. (1967), but occur instead as a distinctive lithologic unit that can be traced from Wyoming to North Dakota (Figs. 2, 5). As such, the Peanut Peak Member is a valuable litho-

stratigraphic marker for determining the superposition of various members and formations within the White River Group across the northern Great Plains (Fig. 10).

Along with the superposition of various lithologic units, the nature of the contacts between units may help to determine the geologic history of the region. The contact relationships of the Peanut Peak and Big Cottonwood Creek Members suggest that a period of low siliciclastic sedimentation developed across the southwestern South Dakota and northwestern Nebraska area following the deposition of the Peanut Peak Member. In the Big Badlands, the upper part of the Peanut Peak Member commonly contains one or more lacustrine limestones (Evans and Welzenbach, this volume). These limestone beds are widespread, and have been cited as the contact between the Chadron and Brule Formations in the Big Badlands of South Dakota (Wanless, 1923; Clark, 1937; Clark et al., 1967; Harksen and Macdonald, 1969a). Welzenbach (1992) has referred to these limestones near Wall, South Dakota, as the Bloom Basin limestone beds.

According to Welzenbach (1992) and Evans and Welzenbach (this volume), these limestones were deposited within perennial, stratified lakes. Based on the low amount of siliciclastics and the presence of tufa pillars, the lakes are believed to have been spring-fed. Evans and Welzenbach (this volume) suggest that the formation of these lacustrine limestones is due to the interaction of paleoclimate, equal rates of sediment supply and basin subsidence, and the development of a regional groundwater recharge system formed by the uplift of the Black Hills. These deposits may represent lacustrine environments on the terraces of an incised drainage system, and possibly indicate drainage disruption prior to the deposition of the overlying Brule Formation (Evans and Welzenbach, this volume). Similar limestone beds occur at the top of the Peanut Peak Member at Oelrichs, South Dakota and at Rattlesnake Butte, Nebraska (Fig. 3B).

In northwestern Nebraska, the contact of the Peanut Peak Member and overlying Big Cottonwood Creek Member is unconformable where the Big Cottonwood Creek Member fills shallow paleovalleys within the Peanut Peak Member (Terry and LaGarry, this volume). Paleovalley formation within the Peanut Peak Member may likely correspond to this regional period of reduced siliciclastic sedimentation. With renewed sedimentation, deposition of siliclastic sediments appears to have shifted to the northwestern Nebraska area. As deposition proceeded in Nebraska these paleovalleys were eventually filled and overtopped by 42 to 52 m (138–170 ft) of sediment to form the Big Cottonwood Creek Member. During this time, lacustrine deposition dominated in the Big Badlands area of South Dakota, but toward the southwest deposition was predominantly fluvial. Lacustrine deposition within the Big Cottonwood Creek Member progressively increases toward the northeast (Terry and LaGarry, this volume, Fig. 4), and is visible in measured sections between Toadstool Park in Nebraska and Limestone Butte near Oelrichs, South Dakota (Figs. 5F, 9).

The mechanism responsible for this shift in fluvial depocenters and formation of the Big Cottonwood Creek Member during the late Eocene is uncertain at this time, but may be at least partially influenced by renewed uplift of the Black Hills or Chadron Dome. According to Lisenbee and DeWitt (1993), the Black Hills was tectonically inactive during the deposition of the White River Group, although data collected by Evans (1996) challenges this hypothesis. The amount of time represented by deposition of the Big Cottonwood Creek Member in northwestern Nebraska may be equivalent to that of the "Bloom Basin limestone beds" of Welzenbach (1992) and other lacustrine limestones described by Evans and Welzenbach (this volume) near the top of the Peanut Peak Member in the Big Badlands (Figs. 2H, 9). Thus, these lacustrine limestones would represent a condensed section in the Big Badlands area that formed over a significant period of time. New data from Prothero and Whittlesey (this volume) indicate an unconformity spanning at least 400,000 yr between the Chadron and Brule Formations in the Sage Creek and Pinnacles area of the Big Badlands. The condensed section of the limestone beds near the Chadron/Brule contact in the Big Badlands and deposition of the Big Cottonwood Creek Member in northwestern Nebraska likely represents this interval of time.

CONCLUSIONS

Based on measured sections, the visual tracing of rock units between the Big Badlands of South Dakota and the Toadstool Park area of northwestern Nebraska, paleopedology, and the use of the Big Badlands of South Dakota as a stratigraphic standard, I propose the following lithostratigraphic changes within the lower part of the White River Group in northwestern Nebraska (Figs. 3B, 8, 9). These changes conform to guidelines of the North American Stratigraphic Code (NACSN, 1983) for recognizing lithostratigraphic units, and replace the informal, event boundary-based classification of Schultz and Stout (1955) (Fig. 7).

1. The term Interior Paleosol Complex of Schultz and Stout (1955) is abandoned. This zone is composed of two separate, pedogenically modified units; the Yellow Mounds Paleosol Equivalent that developed on the Cretaceous Pierre Shale, and the Interior Paleosol Series Equivalent that developed on the overbank deposits of the Chamberlain Pass Formation (Figs. 2A, 5A).

2. The Chamberlain Pass Formation of Evans and Terry (1994) is recognized in northwestern Nebraska. It is composed of the channel sandstone that Schultz and Stout (1955) called the "Chadron A" and red overbank mudstone originally included in the Interior Paleosol Complex (Figs. 2A–D, 5A–D). The term Chadron A is abandoned, and the correlation by Schultz and Stout (1955) of these channel sands (Fig. 2F) to the Ahearn Member of Clark (1954) and the Yoder Beds of Schlaikjer (1935) and Kihm (1987) is rejected.

3. The term "Chadron B" for strata above the Chadron A to the "second" or "lower purplish-white" is abandoned (Figs. 5B,D,E). The correlation of the Chadron B to the Crazy Johnson Member of Clark (1954) in South Dakota is rejected (Fig. 2G). The lower bluish-green and gray hummocky part of the Chadron B is instead lithologically correlated to the "undifferentiated

Chadron" deposited outside of the Red River Valley of Clark (1937) (Figs. 2A–D, 5B,D,E). The term "Peanut Peak Member" is expanded to include these undifferentiated strata outside of the Red River Valley in South Dakota and northwestern Nebraska.

4. The overlying siltier, variegated, and cliff-forming remainder of the Chadron B and overlying "Chadron C" is incorporated, along with the "Orella A" of the overlying Brule Formation, into a new lithostratigraphic unit (Fig. 5E,F), the *Big Cottonwood Creek Member* of the Chadron Formation (Terry and LaGarry, this volume). The term Chadron C is abandoned, and the correlation to the Peanut Peak Member of Clark (1954) within the Red River Valley in South Dakota is rejected (Fig. 2E,G).

This new lithostratigraphic interpretation provides a basis for the recognition of the regional superposition of members and formations within the White River Group across the northern Great Plains, and may help to refine the geologic history of this region, specifically, the interaction of tectonics and sedimentation, during a brief and discrete interval of the Paleogene.

ACKNOWLEDGMENTS

This manuscript was completed as a part of my Ph.D. dissertation while at the University of Nebraska—Lincoln. I thank Hannan LaGarry, David Loope, Jim Swinehart, and Jim Evans for discussions on the finer points of the Stratigraphic Code; Jim Evans for contributing ideas and insights on the Nebraska stratigraphy and for sharing unpublished data; and the various staffs at Badlands National Park and the Nebraska National Forest for contributing time and resources throughout the course of this study. Earlier versions of this manuscript were greatly improved by reviews from Bob Hunt, Jason Lillegraven, David Loope, Greg Retallack, and Hannan LaGarry. This research was funded by the Nebraska National Forest, a Shell Graduate Fellowship from the University of Nebraska—Lincoln Geology Department, the Nebraska Geological Society Yatkola-Edwards Research Grant, two grants from the University of Nebraska Lincoln Center for Great Plains Studies, and a Sigma Xi Grant-in-Aid.

REFERENCES CITED

AAPG (American Association of Petroleum Geologists), 1972, Northern Rocky Mountain region geological highway map: Geological Highway Map Series #12, scale, 1:1,900,800, 1 sheet.

AAPG (American Association of Petroleum Geologists), 1984, Northern Great Plains region geological highway map: Geological Highway Map Series #5, scale 1:1,900,800.

Birkeland, P. W., 1974, Pedology, weathering, and geomorphological research: New York, Oxford University Press, 285 p.

Bjork, P. R., 1967, Latest Eocene vertebrates from northwestern South Dakota: Journal of Paleontology, v. 41, p. 227–236.

Bump, J. D., 1956, Geographic manes for members of the Brule Formation of the Big Badlands of South Dakota: American Journal of Science, v. 254, p. 429–432.

Buol, S. W., Hole, F. D., and McCracken, P. J., 1973, Soil genesis and classification: Ames, Iowa University Press, 560 p.

Cady, J. G, and Daniels, R. B., 1968, Genesis of some very old soils—the paleudults: 9th International Soils Science Congress, Transactions, Adelaide, v. 4, p. 103–112.

Clark, J., 1937, The stratigraphy and paleontology of the Chadron Formation in the Big Badlands of South Dakota: Carnegie Museum Annals, v. 25, p. 261–350.

Clark, J., 1954, Geographic designation of the members of the Chadron Formation in South Dakota: Carnegie Museum Annals, v. 33, p. 197–198.

Clark, J., 1975, Controls of sedimentation and provenance of sediments in the Oligocene of the central Rocky Mountains, *in* Curtis, B. F., ed., Cenozoic history of the southern Rocky Mountains: Geological Society of America Memoir 144, p. 95–117.

Clark, J., Beerbower, J. R., and Kietzke, K. K., 1967, Oligocene sedimentation, stratigraphy, paleoecology and paleoclimatology in the Big Badlands of South Dakota: Field Museum of Natural History, Fieldiana Geology Memoirs, v. 5, 158 p.

Darton, N. H., 1899, Preliminary report on the geology and water resources of Nebraska west of the one hundred and third meridian: U.S. Geological Survey 19th Annual Report, Pt. 4, p. 719–814.

Emry, R. J., Bjork, P. R., and Russell, L. S., 1987, The Chadronian, Orellan, and Whitneyan North American land mammal ages, *in* Woodburne, M. O., ed., Cenozoic mammals of North America: Geochronology and biostratigraphy: Berkeley, University of California Press, p. 118–152.

Evanoff, E., 1990, Late Eocene and early Oligocene paleoclimates as indicated by the sedimentology and nonmarine gastropods of the White River Formation near Douglas, Wyoming [Ph.D. thesis]: Boulder, University of Colorado, 440 p.

Evanoff, E., Prothero, D. R., and Lander, R. H., 1992, Eocene-Oligocene climatic change in North America: the White River Formation near Douglas, east-central Wyoming, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 116–130.

Evans, J. E., 1996, Evidence for Late Eocene unroofing of the Black Hills Uplift: Geological Society of America Abstracts with Programs, v. 28, no. 7, p. A373.

Evans, J. E., and Terry, D. O., Jr., 1994, The significance of incision and fluvial sedimentation in the Basal White River Group (Eocene-Oligocene), Badlands of South Dakota, U.S.A.: Sedimentary Geology, v. 90, p. 137–152.

Harksen, J. C., and Macdonald, J. R., 1969a, Type sections for the Chadron and Brule Formations of the White River Oligocene in the Big Badlands of South Dakota: South Dakota Geological Survey Report of Investigations 99, 23 p.

Harksen, J. C., and Macdonald, J. R., 1969b, Guidebook to the major Cenozoic deposits of southwestern South Dakota: South Dakota Geological Survey Guidebook, no. 2, 103 p.

Harvey, C., 1960, Stratigraphy, sedimentation, and environment of the White River Group of the Oligocene of northern Sioux County, Nebraska [Ph. D. thesis]: Lincoln, University of Nebraska, 151 p.

Hatcher, J. B., 1893, The Titanotherium beds: American Naturalist, v. 27, p. 204–221.

Hickey, L. J., 1977, Stratigraphy and paleobotany of the Golden Valley Formation (Early Tertiary) of western North Dakota: Geological Society of America Memoir 150, 183 p.

Kihm, A. J., 1987, Mammalian paleontology and geology of the Yoder Member, Chadron Formation, east-central Wyoming, *in* Martin, J. E., and Ostrander, G. E., eds., Papers in vertebrate paleontology in honor of Morton Green: Dakoterra, v. 3, p. 28–45.

LaGarry, H. E., LaGarry, L. A., and Terry, D. O., 1996, New vertebrate fauna from the Chamberlain Pass FM. (Eocene), Sioux County, Nebraska; Proceedings, 106th Annual Nebraska Academy of Sciences, Earth Science Section, p. 45.

Larson, E. E., and Evanoff, E., 1995, Tephrostratigraphy of the White River sequence in Nebraska, Wyoming, and Colorado: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A67.

Leonard, B. F., 1957, Oligocene stratigraphy of the Douglas area, Converse County, Wyoming [M.S. thesis]: Lincoln, University of Nebraska, 48 p.

Lillegraven, J. A., 1970, Stratigraphy, structure, and vertebrate fossils of the Oligocene Brule Formation, Slim Buttes, northwestern South Dakota: Geological Society of America Bulletin, v. 81, p. 831–850.

Lisenbee, A. L., and DeWitt, E., 1993, Laramide evolution of the Black Hills uplift, *in* Snoke, A. W., Steidtmann, J. R., and Roberts, S. M., eds., Geology of Wyoming: Geological Survey of Wyoming Memoir, no. 5, p. 374–412.

Lucas, S. G., and Schoch, R. M., 1989, Taxonomy of Duchesneodus (Brontotheriidae) from the late Eocene of North America, *in* Prothero, D. R., and Schoch, R. M., eds., The evolution of perissodactyls: New York, Oxford University Press, p. 490–503.

Luebke, L. O., 1964, Oligocene stratigraphy of the Lance Creek area, Wyoming [M.S. thesis]: Lincoln, University of Nebraska, 78 p.

Mader, B. J., 1989, Brontotheriidae: a systematic revision and preliminary phylogeny of North American genera, *in* Prothero, D. R., and Schoch, R. M., eds., The evolutionof perissodactyls: New York, Oxford University Press, p. 458–484.

Martin, J. E., 1985, Introduction to the geology and paleontology of western South Dakota and northwestern Nebraska, *in* Martin, J. E., ed., Fossiliferous Cenozoic deposits of western South Dakota and northwestern Nebraska, Dakoterra, v. 2, pt. 2, p. 1–11.

Martin, J. E., 1987, The White River Badlands of South Dakota, *in* Beus, S. S., Centennial Field Guide, Boulder, Colorado, Geological Society of America, Rocky Mountain Section, v. 2, p. 233–236.

Meek, F. B., and Hayden, F. V., 1857, Descriptions of new species and genera of fossils, collected by Dr. F. V. Hayden in Nebraska Territory; with some remarks on the Tertiary and Cretaceous formations of the northwest, and the parallelism of the latter with those of other portions of the United States and Territories: Proceedings of the Academy of Natural Science of Philadelphia for 1857, p. 117–148.

Munsell Soil Color Charts; 1975: Baltimore, Munsell Color Company, 24 p.

Murphy, E. C., Hoganson, J. W., and Forsman, N. F., 1993, The Chadron, Brule, and Arikaree Formations in North Dakota: North Dakota Geological Survey Report of Investigations 96, 144 p.

NACSN (North American Commission on Stratigraphic Nomenclature), 1983, North American Stratigraphic Code: American Association of Petroleum Geologists Bulletin, v. 67, p. 841–875.

Obradovich, J. D., Evanoff, E., and Larson, E. E., 1995, Revised single-crystal laser-fusion $^{40}Ar/^{39}Ar$ ages of Chadronian tuffs in the White River Formation of Wyoming: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A77.

Osborn, H. F., 1929, Titanotheres of ancient Wyoming, Dakota, and Nebraska: U.S. Geological Survey Monograph 55, 953 p.

Pettyjohn, W. A., 1966, Eocene paleosol in the northern Great Plains: U.S. Geological Survey Professional Paper 550C, p. 61–65.

Prothero, D. R., and Berggren, W. A., 1992, Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, 568 p.

Prothero, D. R., and Swisher, C. C., III, 1992, Magnetostratigraphy and geochronology of the terrestrial Eocene-Oligocene transition in North America, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 46–73.

Retallack, G. J., 1983, Late Eocene and Oligocene paleosols from Badlands National Park, South Dakota: Geological Society of America Special Paper 193, 82 p.

Retallack, G. J., 1994, A pedotype approach to latest Cretaceous and earliest Tertiary paleosols in eastern Montana: Geological Society of America Bulletin, v. 106, p. 1377–1397.

Schlaikjer, E. M., 1935, Contributions to the stratigraphy and paleontology of the Goshen Hole area, Wyoming, Part III, A new basal Oligocene formation: Bulletin of the Museum of Comparative Zoology, v. 76, no. 3, p. 71–93.

Schultz, C. B., and Falkenbach, C. H., 1968, The phylogeny of the oreodonts. Parts 1–2: Bulletin of the American Museum of Natural History, v. 139, 498 p.

Schultz, C. B., and Stout, T. M., 1955, Classification of Oligocene sediments of Nebraska: Bulletin of the University of Nebraska State Museum, v. 4, p. 17–52.

Schultz, C. B., and Stout, T. M., 1980, Ancient soils and climatic changes in the central Great Plains: Nebraska Academy of Science Transactions, v. 8, p. 187–205.

Schultz, C. B., Tanner, L. G., and Harvey, C., 1955, Paleosols of the Oligocene of Nebraska: Bulletin of the University of Nebraska State Museum, v. 4, p. 1–16.

Singler, C. R., and Picard, M. D., 1979, Petrography of the White River Group (Oligocene) in northwest Nebraska and adjacent Wyoming: Contributions to Geology, University of Wyoming, v. 18, no. 1, p. 51–67.

Singler, C. R., and Picard, M. D., 1980, Stratigraphic review of Oligocene beds in northern Great Plains: Wyoming Geological Association Earth Science Bulletin, v. 13, no. 1, p. 1–18.

Singler, C. R., and Picard, M. D., 1981, Paleosols in the Oligocene of northwest Nebraska: Contributions to Geology, University of Wyoming, v. 20, no. 1, p. 57–68.

Soil Survey Staff, 1975, Soil taxonomy: U.S. Department of Agriculture Handbook 436, 754 p.

Stanley, K. O., and Benson, L. V., 1979, Early diagenesis of High Plains Tertiary vitric and arkosic sandstone, Wyoming and Nebraska, *in* Scholle, P. A., and Schluger, P. R., eds., Aspects of diagenesis: Society of Economic Geologists and Paleontologists Special Publication 26, p. 401–423.

Stone, W. J., 1972, Middle Cenozoic stratigraphy of North Dakota, *in* Ting, F. T. C, ed., Depositional environments of the lignite-bearing strata in western North Dakota: North Dakota Geological Survey Miscellaneous Series 50, p. 123–132.

Swenson, D. B., 1959, The heavy mineral assemblages of the Chadron and Brule Formations in northwestern Nebraska [M.S. thesis]: Lincoln, University of Nebraska—Lincoln, 66 p.

Swinehart, J. B., Souders, V. L., Degraw, H. M., and Diffendal, R. F., Jr., 1985, Cenozoic paleogeography of western Nebraska, *in* Flores, R. M., and Kaplan, S., eds., Cenozoic paleogeography of the west-central United States, Special Publication, Rocky Mountain Section: Tulsa, Society of Economic Paleontologists and Mineralogists, p. 209–229.

Swisher, C. C., III, and Prothero, D. R., 1990, Single-crystal $^{40}Ar/^{39}Ar$ dating of the Eocene-Oligocene transition in North America: Science, v. 249, p. 760–762.

Terry, D. O., Jr., 1991, The study and implications of comparative pedogenesis of sediments from the base of the White River Group, South Dakota [M.S. thesis]: Bowling Green, Ohio, Bowling Green State University, 184 p.

Terry, D. O., 1995, Paleopedology of the Big Cottonwood Creek Member of the Eocene Chadron Formation, White River Group, Badlands of Northwestern Nebraska: Proceedings, 105th Annual Nebraska Academy of Sciences, Earth Science Section, p. 57.

Terry, D. O., Jr., and Evans, J. E., 1994, Pedogenesis and paleoclimatic implications of the Chamberlain Pass Formation, Basal White River Group, Badlands of South Dakota: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 110, p. 197–215.

Terry, D. O., Jr., and LaGarry, H. E., 1994, Lithostratigraphic correlation and provisional redefinition of the Eocene-Oligocene White River Group, Nebraska and South Dakota: Geological Society of America Abstracts with Programs, v. 26, no. 7, p. A249.

Terry, D. O., Jr., LaGarry, H. E., and Wells, W. B., 1995, The White River Group revisited: Vertebrate trackways, ecosystems, and lithostratigraphic revision, redefinition, and redescription: Guidebook, Geological Society of America, North-Central/South-Central Meeting, p. 43–57.

Toepelman, W. C., 1922, The paleontology of the [Badlands] area, *in* Ward, F., ed., The geology of a portion of the Badlands: South Dakota Geological and Natural History Survey Bulletin, v. 11, p. 61–73.

Vondra, C. F., 1958a, The stratigraphy of the Chadron Formation in northwestern Nebraska [M.S. thesis]: Lincoln, University of Nebraska, 138 p.

Vondra, C. F., 1958b, Depositional history of the Chadron Formation in northwestern Nebraska: Proceedings, 68th Annual Meeting, Nebraska Academy of Sciences Abstracts with Programs, p. 16.

Vondra, C. F., 1960, Stratigraphy of the Chadron Formation in northwestern

Nebraska: Compass, v. 37, no. 2, p. 73–90.

Wanless, H. R., 1922, Lithology of the White River sediments: American Philosophical Society Proceedings, v. 62, p. 663–669.

Wanless, H. R., 1923, The stratigraphy of the White River beds of South Dakota: American Philosophical Society Proceedings, v. 62, p. 191–267.

Ward, F., 1922, The geology of a portion of the Badlands: South Dakota Geological and Natural History Survey Bulletin, v. 11, 80 p.

Welzenbach, L. C., 1992, Limestones in the Lower White River Group (Eocene-Oligocene), Badlands of South Dakota: depositional environment and paleoclimatic implications [M.S. thesis]: Bowling Green, Ohio, Bowling Green State University, 131 p.

Welzenbach, L. C., and Evans, J. E., 1992, Sedimentology and paleolimnology of an extensive lacustrine unit, White River Group (Eocene-Oligocene), South Dakota: Geological Society of America Abstracts with Programs, v. 24, no. 7, p. A54.

Wood, H. E., Chaney, R. W., Clark, J., Colbert, E. H., Jepsen, G. L., Reeside, J. B., Jr., and Stock, C., 1941, Nomenclature and correlation of the North American continental Tertiary: Geological Society of America Bulletin, v. 52, p. 1–48.

Manuscript Accepted by the Society December 5, 1997

Printed in U.S.A.

Geological Society of America
Special Paper 325
1998

Magnetic stratigraphy and biostratigraphy of the Orellan and Whitneyan land-mammal "ages" in the White River Group

Donald R. Prothero
Department of Geology, Occidental College, Los Angeles, California 90041
Karen E. Whittlesey
Department of Geological Sciences, University of Southern California, Los Angeles, California 90089

ABSTRACT

Detailed magnetostratigraphic studies conducted over the last 20 yr permit much better chronostratigraphic correlation of the White River Group. In addition, recent systematic revisions of long-unstudied taxa and fine-scale biostratigraphic data from recently studied fossil collections allow us to subdivide the Orellan and Whitneyan land-mammal "ages." Instead of the hybrid of lithostratigraphy and biochronology created by the Wood Committee (Wood et al., 1941), it is now possible to designate biostratigraphic zones and true chronostratigraphic stages and geochronologic ages.

The Chadronian-Orellan boundary (previously based on both the Chadron-Brule formational contact, and on the last occurrence of brontotheres) is here redefined by the first appearance datum (FAD) of *Hypertragulus calcaratus* (latest Chron C13r, 33.7 Ma). This FAD also marks the beginning of the earliest Orellan *H. calcaratus* Interval Zone. The late early Orellan (*Miniochoerus affinis* Interval Zone) is defined by the FAD of *Miniochoerus affinis, Eumys elegans,* and the last appearance datum (LAD) of *Ischyromys parvidens* (late Chron C13n, 33.3 Ma). The early late Orellan (*Miniochoerus gracilis* Interval Zone) is recognized by the FAD of *Miniochoerus gracilis* and *Mesohippus barbouri* (earliest Chron C12r, about 32.8 Ma). The latest Orellan (*Merycoidodon bullatus* Interval Zone) is defined by the FAD of *Miniochoerus starkensis, Palaeolagus burkei,* and *Merycoidodon bullatus* (early Chron C12r, about 32.5 Ma). The Orellan-Whitneyan boundary is marked by the FAD of *Leptauchenia major,* and abundant *L. decora* (mid-Chron C12r, about 32.0 Ma), so the early Whitneyan has been designated the *L. major* Interval Zone. The late Whitneyan (*Merycoidodon major* Interval Zone) can be recognized by the FAD of *M. major, Miohippus intermedius, M. gidleyi,* and the LAD of *Miniochoerus* and *Hyaenodon horridus* (late Chron C12r, about 31.3 Ma). The Whitneyan-Arikareean boundary is marked by the FAD of *Nanotragulus loomisi, Palaeocastor nebrascensis, Leidymys blacki,* and *Mesoreodon minor* (Chron C11n, about 30.0 Ma).

INTRODUCTION

When mammal fossils were discovered in the Big Badlands in 1846, they were among the first fossil vertebrates reported from west of the Mississippi River. Soon thereafter, collectors flocked to the Badlands and built up large collections of fossil mammals (see historical reviews by Harksen and Macdonald, 1969b; Emry et al., 1987). In the 150 yr since the first White River fossils were described, these collections have grown enormously. White River fossils are sold commercially worldwide

Prothero, D. R., and Whittlesey, K. E., 1998, Magnetic stratigraphy and biostratigraphy of the Orellan and Whitneyan land-mammal "ages" in the White River Group, *in* Terry, D. O., Jr., LaGarry, H. E., and Hunt, R. M., Jr., eds., Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America): Boulder, Colorado, Geological Society of America Special Paper 325.

and have been studied by many scientists over the past century. Despite this long history of study, however, many aspects of White River paleontology are still poorly known.

The stratigraphic terminology of the White River Group was initially biostratigraphic in character. Hatcher (1893) and Wortman (1893) recognized three units: the "*Titanotherium* beds" (now the Chadron Formation of Darton, 1899); the "Turtle-*Oreodon* beds" (now the Scenic or Orella Members of the Brule Formation); and the "*Protoceras* beds" (now the Poleslide or Whitney Members of the Brule Formation). Early in this century, William Diller Matthew and Henry Fairfield Osborn (Matthew, 1899; Osborn and Matthew, 1909; Osborn, 1907, 1910, 1929) continued this fundamentally biostratigraphic tradition by recognizing the "*Titanotherium* beds," "lower," "middle," and "upper *Oreodon* beds," "*Protoceras* beds," and "*Leptauchenia* beds." Although *Titanotherium* and *Oreodon* are no longer valid taxa, Matthew and Osborn laid the foundation for a range-zone biostratigraphy of the White River Group comparable to that practiced by biostratigraphers of marine invertebrate fossils.

When the Wood Committee (Wood et al., 1941) set up the "North American provincial ages" (now known as North American land-mammal "ages"), they abandoned this biostratigraphic tradition. Instead, they used a mixture of lithostratigraphy and biochronology (in the sense of Williams, 1901) to define their "provincial ages." For example, the Chadronian was defined both on the lithostratigraphic Chadron Formation, and also on the overlapping biochronologic ranges of *Mesohippus* and brontotheres. The Wood Committee apparently assumed that the last appearance of brontotheres would always coincide with the top of the Chadron Formation.

Fossils collected by Morris Skinner and other field crews of the Frick Laboratory in the 1950s made the Wood Committee's definition of the Chadronian obsolete. On September 6, 1953, Morris Skinner (unpublished section books, vol. 5, p. 9–11, *in* the Department of Vertebrate Paleontology Archives, American Museum of Natural History) collected titanothere bones from a channel-fill deposit cut down from ~6.5 m (20 ft) above his "Persistent White Layer," or "PWL." This is presumably the same marker bed that was used to recognize the Chadron-Brule contact in Nebraska (although called the "Upper Purplish White," or "UPW," layer by Schultz and Stout, 1955, 1961; see the revisions of Terry and LaGarry, this volume). These brontothere specimens came from the Seaman Hills, near Lusk, Niobrara County, Wyoming, in a locality known as the Thompson Ranch Anthill (after its conical shape), located in NE sec. 28, T. 35 N., R. 61 W., Shepherds Point 7.5′ Quadrangle, Niobrara County, Wyoming. In his 1960 summary of the Seaman Hills stratigraphy, Skinner again indicated "Chadronian age at least to here" at a level 6.5 m (20 ft) above the PWL.

Some of Skinner's peers did not differentiate between lithostratigraphy and biostratigraphy. On the same diagram is a notation (not in Skinner's handwriting, but initialed by CHF, for Charles H. Falkenbach, and dated 9/9/60) that comments, "The Purple White is considered the top of the Chadron." Falkenbach, Schultz, and Stout did not distinguish between Chadron Formation (a lithostratigraphic unit) and Chadronian "age" (a biochronologic unit). This confusion between lithostratigraphy and biochronology is apparent in other paleontologic research of that generation. For example, Schultz and Falkenbach (1968) defined their oreodont taxa partly on their lithostratigraphic occurrence, and Schultz and Stout used oreodonts to define the top of the Chadron Formation.

The Seaman Hills brontothere specimens were not unique. In 1958, Skinner (unpublished section books, vol. 6, p. 40–41) noted another channel sequence in the area southeast of Douglas, Wyoming ~7 m (25 ft) above his "100 correlator white zone" in Douglas, also known as the "5 tuff" of Evanoff et al., 1992), which yielded a partial jaw and large scapula of a brontothere. The large size of the specimens makes it very unlikely that they were reworked. These specimens were recorded from Reno Ranch, South of Tower in the SW sec. 31, T. 32 N., R. 70 W., Douglas 7.5′ Quadrangle, Converse County, Wyoming. Although neither of these occurrences was published by Skinner, they were widely known among White River researchers, and eventually published by Emry et al. (1987, p. 139). Other researchers reported Brule Formation brontothere specimens about the same time. Schultz and Stout (1955, Fig. 10, section 1) show the position of a titanothere metapodial from a channel sandstone incised from ~2 m (7 ft) above the UPW (PWL of Skinner) at Scottsbluff National Monument. These authors (p. 27, footnote 1) comment that "the lowest parts of the Brule (basal Orella or Orella A) should be expected to yield titanothere remains," but they did not elaborate. Terry and LaGarry (this volume) consider these strata to be upper Chadron Formation, not Orella Member of the Brule Formation.

These occurrences of brontotheres above the Chadron Formation show that the original definition of the Chadronian/Orellan boundary has become unworkable. As Emry et al. (1987, p. 139) noted, the Chadronian/Orellan boundary cannot be based on lithostratigraphic criteria, but only on biostratigraphic criteria. They wrote that "it should be placed at the most prominent and most widely recognizable faunal break, which may or may not coincide with lithostratigraphic boundaries. The single criterion of presence or absence of titanotheres is insufficient—the absence of titanotheres at any particular locality does not necessarily mean non-Chadronian."

Similarly, the Wood Committee (Wood et al., 1941) based the Orellan on both the geochron of the Orella Member and on a number of characteristic mammalian taxa. The Whitneyan was based on the geochron of the Whitney Member and on a smaller number of characteristic fossil mammals. As the problem with the Chadronian-Orellan boundary demonstrates, such dual definitions are prone to conflict and ambiguity (see Tedford, 1970; Emry et al., 1987; Woodburne, 1977, 1987; Terry and LaGarry, this volume). In addition, these ages were not true geochronologic ages, that must be based on biostratigraphic units (according to western stratigraphic codes, such as the 1983 North American Stratigraphic Code). Since they were informal stratigraphic ages, the North American land mammal "ages" should properly be put in quotes in publications.

Since the 1950s, some paleontologists (see Savage, 1955, 1962, 1977; Tedford, 1970; Woodburne, 1977, 1987) have attempted to apply more rigorous methods to vertebrate biostratigraphy. Modern vertebrate biostratigraphers accept the possibility that rock units can be time-transgressive over distance, and therefore separate rock units and time units. Also, detailed stratigraphic zonations of mammal fossils provide much higher resolution of time than zonation schemes based on collections in which the only stratigraphic information is the formation they came from. As discussed by Woodburne (1977), detailed zonation could potentially subdivide the Cenozoic Era into time increments of 300,000 yr or less.

However, despite enormous collections, White River biostratigraphy lags far behind zonations proposed for the rest of the Cenozoic. This is primarily due to two factors. The principal impediment is the lack of up-to-date systematics of White River mammals. Since the monographs of Scott et al. (1937–1941), relatively few comprehensive revisions of White River mammals have been published. In addition, some paleontologists oversplit certain groups into many invalid taxa (e.g., Osborn, 1929, on brontotheres; Schultz and Falkenbach, 1956, 1968, on oreodonts). This oversplitting has hindered further investigation, and discouraged many scientists from trying to correct the problem.

A second problem is the biostratigraphic data. Most early collectors of White River fossils had only a vague idea of the stratigraphic or geographic position of each specimen, so their collections cannot be used for high-resolution biostratigraphy. In other cases, some biostratigraphic data was collected (e.g., the Toadstool Park zonation of Schultz and Stout, 1955, for the University of Nebraska State Museum collections, or the zonation of Lillegraven, 1970, for the Slim Buttes collections). The data on these specimens only gives the alphabetically labeled zone (e.g., Orella B in Toadstool Park, or Lillegraven's unit C at Slim Buttes) from which they came, and not the level *within* the zone. Some of these zones are more than 30 m (100 ft) thick, so knowing only the zone does not give very precise information about the exact level of a given fossil in the section. Also, the ranges of taxa within each zone are automatically extended to the lithostratigraphic unit boundaries, whether or not they actually range through the entire unit. This permits only a crude biostratigraphic zonation, when much higher resolution is possible if the collectors make the effort at the time of collection.

Skinner and the field crews of the Frick Laboratory of the American Museum of Natural History in New York were precise in their stratigraphic documentation of fossils. Typically, Skinner would measure a section and establish marker horizons in an area before any extensive collecting was done. Thereafter, all Frick specimens were zoned 2 to 3 m (5–10 ft, some to the nearest foot) from a marker layer (usually a volcanic ash in most collecting areas in the White River Group). These stratigraphic data were usually placed on the specimen itself. Thus, the Frick Collection offers an unparalleled opportunity to construct a range-zone biostratigraphy of the White River Group and replace the land-mammal "ages" with true biostratigraphic stages and ages.

Emry et al. (1987) provided an indication of what such a zonation could be like. They did not propose specific range zones, because too many stratigraphic and systematic problems were unresolved at that time. In 1982, Prothero also made an attempt to zone the late Chadronian through early Whitneyan, but, like Emry et al. (1987), he could not publish a formal biostratigraphy because too many taxonomic problems (especially among key taxa, such as the oreodonts, ischyromyids, and leptomerycids) remained. Since that time, however, a number of systematic studies have been published (various papers in Prothero and Emry, 1996a), so this impediment has been removed in many cases.

Korth (1989) proposed biostratigraphic zones for the Orellan, based on rodent and lagomorph collections in the University of Nebraska State Museum. Unfortunately, these specimens are zoned only to Orella A, Orella B, and so on, with no information as to where they occur within each zone, so the biostratigraphic resolution is no better than the duration of the lithostratigraphic units. In addition, Korth (1989) did not designate type sections for his zones, as required by the 1983 North American Stratigraphic Code. As discussed below, there are additional problems with Korth's zonation. Instead, we have opted to build the biostratigraphy of the Brule Formation from the more finely resolved stratigraphic data of the specimens in the Frick Collection, and we use the Toadstool Park stratigraphy only after high-resolution biostratigraphy has been established in other areas.

The chronostratigraphic calibration of White River biostratigraphic zones has been aided immensely by the addition of magnetostratigraphy and $^{40}Ar/^{39}Ar$ dating (Prothero and Swisher, 1992). Although biostratigraphy is still the fundamental basis for correlations among many White River outcrops, magnetic stratigraphy, and geochronology provide further tests of these correlations. In addition, the magnetic stratigraphy and $^{40}Ar/^{39}Ar$ dates provide time planes for estimation of rates, so that disputes over "how old" or "how long" (e.g., Clark et al., [1967], contended that the Scenic Member spanned only 1,100 to 11,000 yr; Retallack [1983], estimated the duration of Brule deposition in the Pinnacles area of the Badlands) can be resolved. All of the $^{40}Ar/^{39}Ar$ dates used in this study were done by Swisher (Prothero and Swisher, 1992; Swisher and Prothero, 1990). Slightly younger dates for the 5 tuff at Douglas and two of the Flagstaff Rim ashes have been reported by Obradovich et al. (1995). If these dates prove valid, then the chronostratigraphic age assignments in this paper may have to be adjusted slightly back (younger by less than 0.5 m.y.). However, there are no dates by Obradovich on the key Orellan and Whitneyan ashes, so their potential impact on the magnetic time scale is presently unknown. Consequently, we use the time scale of Berggren et al. (1995) and the dates of Prothero and Swisher (1992) for consistency.

Most important, magnetic stratigraphy and $^{40}Ar/^{39}Ar$ dating allow correlation of the North American record to the global time scale, so that the relationship of climatic and faunal events in North America can now be more confidently connected to global climatic and tectonic events. This has resolved many confusing issues in the Eocene and Oligocene (see Prothero, 1994).

The lithostratigraphy and magnetic stratigraphy of many of the collecting areas discussed below are reported elsewhere (Prothero, 1982, 1985a,b, 1996a; Prothero et al., 1982, 1983; Prothero and Swisher, 1992; Evanoff et al., 1992; Terry, this volume; Terry and LaGarry, this volume; LaGarry, this volume). Here we summarize the magnetic stratigraphy of two critical but previously unpublished areas, the Big Badlands of South Dakota and the Seaman Hills of Niobrara County, Wyoming.

MAGNETIC METHODS

Magnetostratigraphic research on the White River Group began with Charles Denham of the Woods Hole Oceanographic Institution in 1976 (Denham, 1984). In 1979, Prothero began his dissertation research on the magnetic stratigraphy of White River sections not studied by Denham (Prothero, 1982). Further sampling was conducted in the summers of 1983, 1986, and 1987. Brief summaries of this research (Prothero et al., 1982, 1983; Prothero, 1985a,b; Swisher and Prothero, 1990; Evanoff et al., 1992; Prothero and Swisher, 1992) have been published, but space limitations prevented a detailed presentation of the polarity stratigraphy of each section. This chapter and previous work (Prothero, 1996a) present the details of 20 years of magnetostratigraphic research in the White River Group. (See Fig. 1 for a map of localities mentioned in the text.)

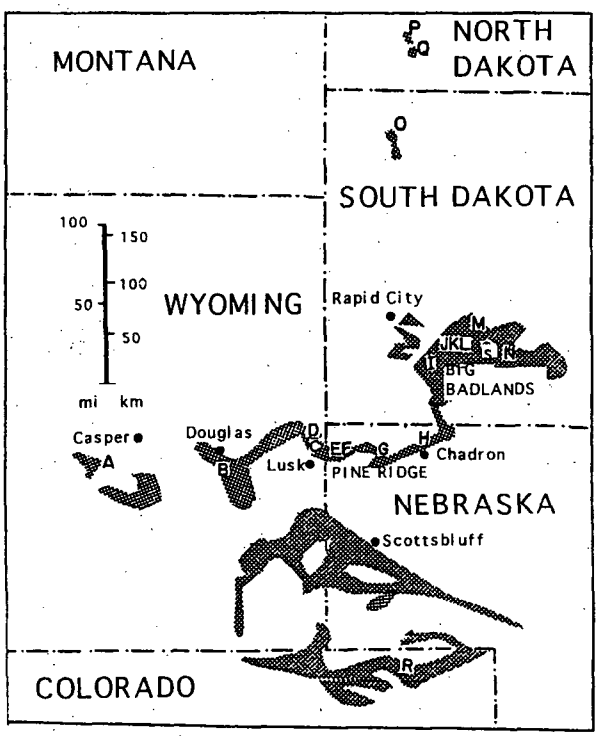


Figure 1. Index map showing localities mentioned in text. Outcrop of White River Group indicated by stipple. A = Flagstaff Rim, Natrona County, Wyoming (Prothero and Swisher, 1992); B = Douglas–Dilts Ranch area, Converse County, Wyoming (see Evanoff et al., 1992); C = Sherrill Hills and Thompson Ranch; D = Boner Ranch, in the Seaman Hills, Niobrara County, Wyoming; E = Geike Ranch; F = Munson Ranch; G = Toadstool Park and Raben Ranch area (E–G in Sioux County, Nebraska); H = Trunk Butte and Morris-Bartlett Ranches, Dawes County, Nebraska; I–N = sections in Big Badlands of South Dakota, including I, Red Shirt Table, J, Cottonwood Pass, K, Sheep Mountain Table, L, type Scenic Member (Chamberlain Pass on some maps), M, Sage Creek and Pinnacles; N, Cedar Pass and Wolf Table–Wanblee; O = Slim Buttes, Harding County, South Dakota (see Prothero, 1985b); P = Little Badlands; Q = Fitterer Ranch, Stark County, North Dakota; R = Flat Top and Chimney Canyons, Logan County, Colorado; S = Pig Dig, Conata Basin, South Dakota.

As previously described in Prothero et al. (1983) and Prothero (1985a), all samples were collected with simple hand tools as horizontally oriented blocks of rock. Three samples were collected per site; most sites were spaced 1.7 m (5.5 ft) stratigraphically. In the laboratory, the samples were trimmed into oriented cubes ~2.5 cm in length.

Different laboratories were used for analysis of the samples as they were collected over the years, and our laboratory procedures have changed. Most samples from sections collected in the field summers of 1979–1980 were run on the ScT cryogenic magnetometer at Woods Hole Oceanographic Institution. These samples were treated primarily with AF (alternating field) demagnetization; thermal demagnetization was undertaken later at Lamont-Doherty Geological Observatory when it became apparent that AF demagnetization could not remove overprints due to iron hydroxides. The 1979–1980 sections include most of the Pine Ridge strata in Sioux County, Nebraska, and in the Lusk and Douglas areas of Wyoming. Prothero and field crew also sampled the Red Shirt Table section in the Big Badlands of South Dakota to complement Denham's sections elsewhere in the Big Badlands (Denham, 1984). A more detailed discussion of the geology and stratigraphy of most of these regions follows; for further details, see Prothero (1982).

In 1983, the original data base was expanded with sampling at Scottsbluff National Monument in the North Platte Valley, Scottsbluff County, Nebraska, and in Cottonwood Pass and the Indian Creek drainage west of Sheep Mountain Table in the Big Badlands, near the type sections of Clark's (1937) Ahearn, Crazy Johnson, and Peanut Peak Members of the Chadron Formation. The Scottsbluff section appeared in Prothero and Swisher (1992, Fig. 2.6), but the rest of the 1983 sections have not been published. These samples were run on the ScT cryogenic magnetometer at the laboratory of the South Dakota School of Mines in Rapid City, using mostly thermal demagnetization, with AF demagnetization of pilot samples undertaken to determine coercivity behavior.

In the early 1980s, it became apparent that the original magnetic analysis by Denham and Farmer was inadequate because they used only one sample per site (preventing any calculation of site statistics) and used almost no thermal demagnetization to remove overprinting that might be due to high-coercivity iron hydroxides. The resampling and thermal demagnetization analy-

sis of samples from the Flagstaff Rim section (Prothero, 1985a; Prothero and Swisher, 1992, Fig. 2.4) radically changed the pattern originally obtained by Denham (*in* Prothero et al., 1982, Fig. 1; Prothero et al., 1983, Fig. 6). In addition, the Big Badlands sections needed to be resampled, since none of the results reported by Denham (1984) were based on thermal demagnetization. In 1986, Prothero and field crew conducted a dense resampling program in the Big Badlands, following the measured sections and field notes of the Frick Laboratory (particularly those of Morris Skinner). The Cedar Pass, Pinnacles, and Wolf Table–Wanblee sections were sampled that summer. In 1987, Prothero and crew continued the Big Badlands sampling with a section at Sheep Mountain Table. These Badlands magnetic sections were summarized in Prothero and Swisher (1992, Fig. 2.7). In 1994, Whittlesey added to the Badlands coverage with sections at Sage Creek and in the type Scenic area (labeled Chamberlain Pass on some maps). In 1995, she also sampled a section in the Conata Basin near the locality called Pig Dig.

The samples that were taken in 1979–1980 and 1983 were treated with thermal demagnetization at 300°-500°C, based on detailed stepwise thermal and AF demagnetization of a pilot suite (e.g., Prothero et al., 1983, Fig. 4). All of the samples run since 1986 have been analyzed on a 2G cryogenic magnetometer using extensive thermal and AF demagnetization at the paleomagnetics laboratory of the California Institute of Technology. AF demagnetization (Fig. 2) showed that most samples declined in intensity rapidly, so that the primary carrier of remanence was a low-coercivity mineral such as magnetite. However, in a few samples there is little response to AF treatment, suggesting that hematite or goethite was more important than magnetite. Thermal demagnetization to 600°C removed nearly all the magnetization in some specimens, suggesting that magnetite was the important magnetic mineral in these specimens. Others, however, still had significant remanence above 600°C, probably due to hematite. These analyses showed that overprints were removed between 200° and 300°C, with the best results obtained between 300° and 400°C.

Isothermal remanent magnetization acquisition studies (IRM) (Fig. 3) clearly showed saturation in many samples between 300 and 1,000 mT (millitesla), indicating that the carrier of remanence was mostly magnetite. A few samples, however, did not saturate at fields higher than 1300 mT, indicating the presence of hematite. A modified Lowrie-Fuller anhysteretic remanent magnetization (ARM) test (e.g., Johnson et al., 1975) was also conducted during the IRM analysis (see Pluhar et al., 1991, for details). This test compares the resistance of AF demagnetization of both an IRM acquired in a 100-mT peak field, and an ARM gained in a 100-mT oscillating field. In most samples, the ARM (black squares) demagnetized at higher peak fields than did the IRM (open squares), indicating that the remanence was carried by single-domain or pseudo-single-domain grains.

Based on these results, the vectors obtained between 300° and 500°C were averaged using the methods of Fisher (1953; see Butler, 1992). Class I sites of Opdyke et al. (1977) showed a clustering that differed significantly from random at the 95% confidence level. In Class II sites, one sample was lost or crumbled (so no site statistics could be calculated), but the remaining samples gave a clear polarity indication. In Class III sites of

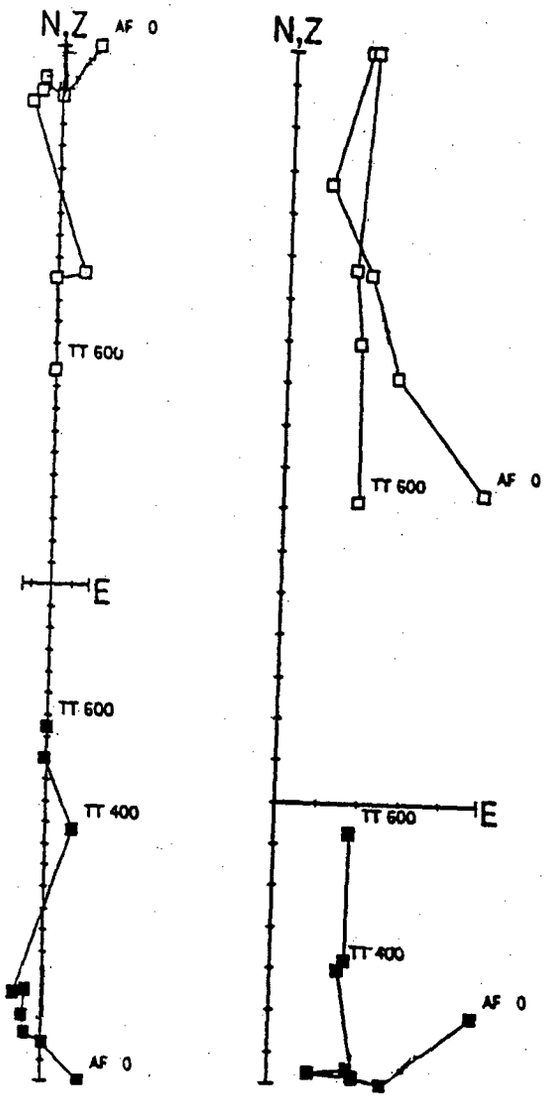


Figure 2. Typical orthogonal demagnetization plots ("Zijderveld" plots) of alternating field (AF) and thermal demagnetization of representative samples. Declination is shown by solid boxes (with the AF field in Gauss or temperature (TT) in degrees Centigrade indicated); inclination by open boxes. Each division = 10^{-6} emu. In most samples, the intensity drops rapidly through AF demagnetization, suggesting that magnetite is the primary carrier of the remanence. In each case, a normal overprint is removed by 200°C, and a stable reversed component is apparent between 300° and 500°C. This component was used in further analysis.

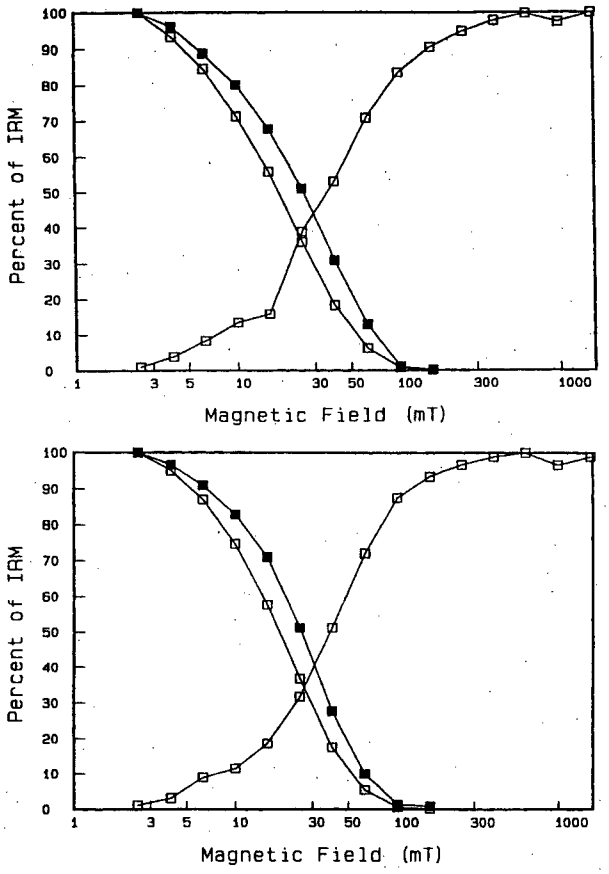


Figure 3. IRM (isothermal remanent magnetization) acquisition (ascending curve on the right) and Lowrie-Fuller test (two descending curves on left) of typical samples. Open boxes indicate IRM; solid boxes indicate ARM (anhysteretic remanent magnetization). In most samples, the IRM saturates by 300 mT (millitesla), showing that magnetite is the dominant magnetic mineral phase. In the Lowrie-Fuller test, the ARM (black squares) is more resistant to AF demagnetization than the IRM (open squares) suggesting that single-domain or pseudo-single-domain grains predominate (see Pluhar et al., 1991, for details of the methods).

Opdyke et al. (1977), two samples showed a clear polarity preference, but the third sample was divergent because of insufficient removal of overprinting. A few samples were considered indeterminate if their magnetic signature was unstable, or their direction uninterpretable.

The means for the normal and reversed sites at each locality were also averaged using the methods of Fisher (1953). The mean of reversed sites (D = 164.5, I = -51.9, k = 26.1, α_{95} = 5.9) is antipodal to the mean of normal sites (D = 356.0, I = 57.3, k = 17.7, α_{95} = 10.6) within the α_{95} error estimate. This positive reversal test suggests that the magnetization is primary and not due to secondary overprinting. Most of the strata are horizontal, so no fold test could be conducted.

In some sections, it was not possible to completely remove the overprinting on every site, so that there are isolated single-site "polarity events." In most cases (such as single normal sites within a long reversed interval), these are most likely due to unremoved normal overprinting. Consequently, the correlations discussed below are based only on magnetozones that are at least two or more sites thick; single-site polarity events are not correlated between regions, since they are not consistently found in every section. However, the possibility that these short polarity events are real cannot be ruled out, since detailed analysis of Oligocene deep-sea cores (Hartl et al., 1993) demonstrates that there were a number of brief polarity events during the Eocene and Oligocene.

MAGNETIC RESULTS

As mentioned above, three areas with a good record of the Chadronian, Orellan, and/or Whitneyan also yield brontothere bones above the Chadron Formation. The Douglas sections have already been described by Evanoff et al. (1992). The Scottsbluff section was published in Prothero and Swisher (1992, Fig. 2.6). The Seaman Hills sections and the Big Badlands sections have not yet been fully published and are described below.

Badlands National Park, South Dakota

The Big Badlands of South Dakota have always been among the most important collecting areas for White River mammals, and their stratigraphy has been studied by a number of people over the years (e.g., Sinclair, 1921; Wanless, 1923; Clark, 1937; Clark et al., 1967; Harksen and Macdonald, 1969a,b; Retallack, 1983, among others; reviewed by Emry et al., 1987). In addition to these published works, Morris Skinner and parties of the Frick Laboratory have made extensive and well-documented collections in the Big Badlands since the mid-1940s.

Most magnetic sections were measured following Skinner's unpublished field notes, so that Frick Collection fossils could be tied as closely as possible to the magnetic stratigraphy. Sections were chosen to be spaced out throughout the arc of the Badlands escarpment, from Wolf Table on the east to Red Shirt Table on the west. Locations of each section are given in Figures 1 and 4.

As discussed by Prothero and Swisher (1992), there is a clear polarity pattern throughout the White River Group. An early Orellan normal zone is found in nearly every section in other areas, and in most Big Badlands sections (Fig. 4). Based on $^{49}Ar/^{39}Ar$ dates from below and above this zone in Douglas, Wyoming, and Scottsbluff, Nebraska (Swisher and Prothero, 1990; Prothero and Swisher, 1992), this early Orellan normal magnetozone correlates with Chron C13n (33.0–33.4 Ma in the time scale of Berggren et al., 1995).

In the Sage Creek, Pinnacles, and Cedar Pass sections, there is no zone of normal polarity associated with early Orellan mammals. However, the scientists of the Frick Laboratory have long known that the earliest Orellan index taxa (discussed in the following section) found at Lusk and Douglas, Wyoming, are absent

from the Big Badlands. This suggests that there is a subtle earliest Orellan unconformity below the late early Orellan "lower nodular zone" fossils in the Badlands. M. F. Skinner (unpublished field notebooks) had known this since the 1950s, although his ideas reached print only as a figure in Mellett (1977, Fig. 71, p. 129). Skinner's discovery was later published by Prothero (1985b) and Prothero and Swisher (1992). We agree with Skinner that there is an unconformity in this position, and we believe that it removes Chron C13n (and therefore at least 400,000 yr of strata) directly beneath the lower nodules at Sage Creek, Pinnacles, and Cedar Pass. The truncation of the biostratigraphic ranges at this level support this interpretation.

The rest of the Scenic Member is of reversed polarity, as is the lower Poleslide Member. In the middle Poleslide is another normal magnetozone that correlates with the mid-Whitneyan normal magnetozone in Nebraska. This zone includes the Upper Whitney Ash at Scottsbluff, which is $^{40}Ar/^{39}Ar$ dated at 30.58 ± 0.18 Ma (Swisher and Prothero, 1990; Prothero and Swisher, 1992). Based on this date, the mid-Whitneyan normal zone must be Chron C12n, estimated at 30.4 to 30.9 Ma (Berggren et al., 1995). The rest of the Poleslide Member correlates with Chron C11r. Near the base of the Rockyford Ash is the beginning of Chron C11n (30.0 Ma). The correlation of the Sharps Formation with units in Nebraska is discussed further by Tedford et al. (1996).

As is apparent from Figure 4, some of the classic lithostratigraphic units of the Badlands are not good time markers. In particular, the fossiliferous "lower nodular zone" used by nearly all collectors to zone their specimens appears to be noticeably diachronous. At Red Shirt Table, the type Scenic area (known on some maps as Chamberlain Pass), Cottonwood Pass, and Wolf Table, it occurs within Chron C13n, but at Sheep Mountain Table, the top of the lower nodules continues into Chron C12r. At Sage Creek, Pinnacles, and Cedar Pass, the lower nodules appear to lie entirely within early Chron C12r. This is almost a million years of age discrepancy between the base of the lower nodules at the type Scenic and at Sage Creek, yet these sections are only a few miles apart. In this chapter, when we report biostratigraphic data of a Big Badlands fossil with reference to the lower nodular zone, we are referring to collections made in the Sheep Mountain Table–Cottonwood Pass–Big Corral Draw area, where most of the Frick collections come from, and where the lower nodular zone does not transgress time.

The variability of the lower nodules was apparent to M. F. Skinner (unpublished field notes) and to many other Badlands researchers. The top and bottom of the nodules are often hard to define, and in some places this unit is not distinct at all. Thus, it should not be surprising that this marker unit is not the same age in every section. The physical causes of the nodular zones in White River Group are still controversial. Retallack (1983) attributed the nodules in the Pinnacles' section to pedogenic processes and implied that there was a synchronous soil-forming event in the early Orellan responsible for the lower nodules. Others have argued that they are due to groundwater-related carbonate concentrations, so they need not be synchronous. Wells et al. (1995; this volume) attribute the nodular zones in northwestern Nebraska to the jointing and erosion of sheet sandstones, and deny that they are true nodular zones like those of the Big Badlands. Whatever their cause, it is clear that the lower nodules cannot be used as a time marker, but only as a local lithostratigraphic unit for zoning fossil collections. Consequently, when a Badlands fossil is collected from the lower nodules, one needs to know exactly *where* it was collected in order to use it for biostratigraphy.

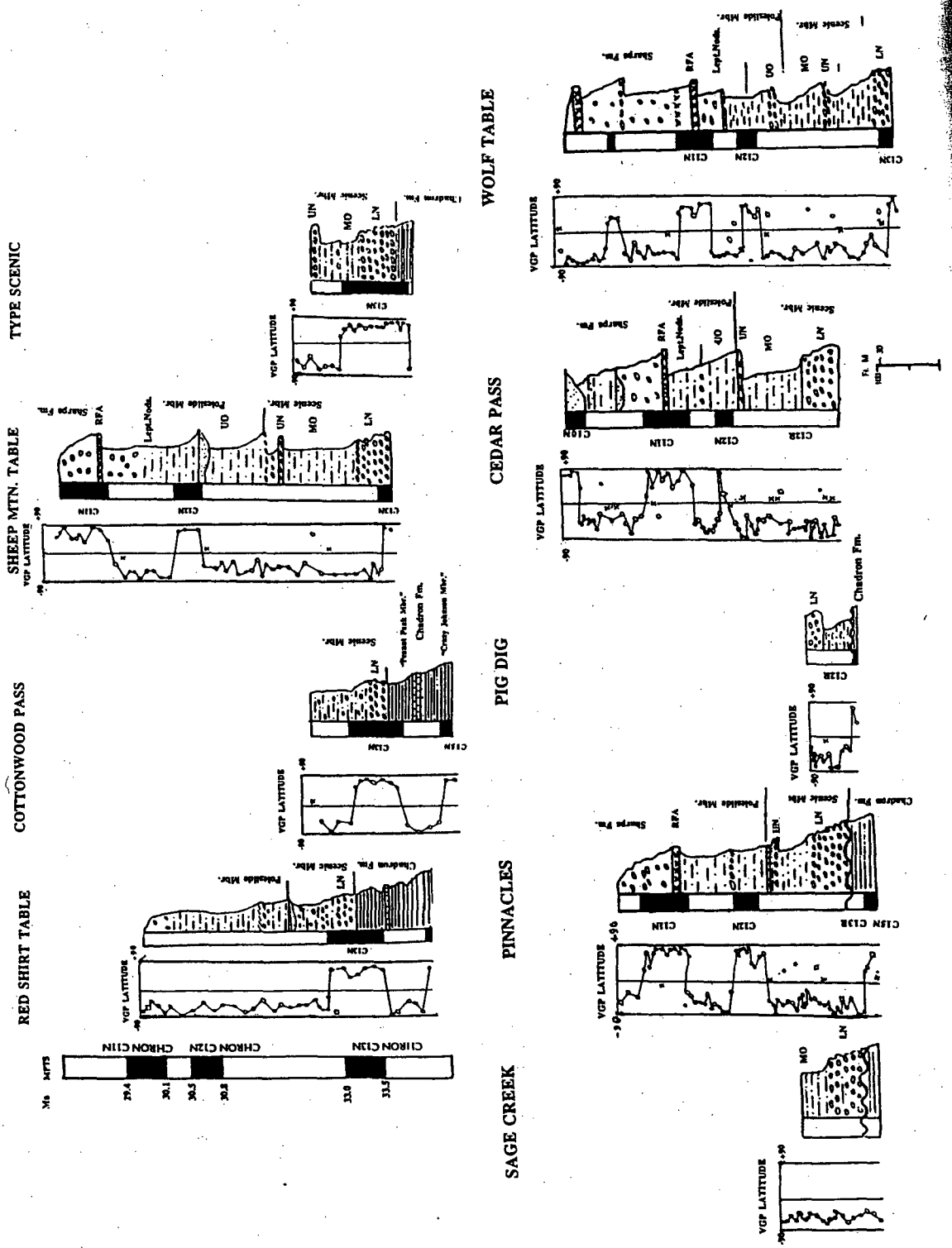
Seaman Hills, north of Lusk, Niobrara County, Wyoming

Perhaps the most important area for studying the Chadronian-Orellan transition is the Seaman Hills, southeastern Wyoming (Figs. 1, 5). Here, the Chadronian-Orellan transition is preserved in much greater detail than in the Big Badlands or Toadstool Park, with no significant unconformities (based on the continuity of the biostratigraphic record; see below). The lithology is uniformly fine volcaniclastic siltstones, with few sandstones or nodular zones, so there is little facies variation to affect fossil occurrences. The PWL is prominent throughout the area, so most of the fossils collected by Morris Skinner et al. in the Frick Laboratory are zoned to this marker horizon. In addition, fossils occur almost uniformly throughout the section, unlike the patchy distribution in the Douglas section (Evanoff et al., 1992).

Despite its great significance, however, the geology and stratigraphy of the Seaman Hills have never been adequately documented. Many important Frick specimens shipped to the American Museum ("Lusk-O" box numbers) have been described (e.g., Schultz and Falkenbach, 1968; Mellett, 1977; Prothero and Shubin, 1989), but the stratigraphic level of these specimens is known almost exclusively from Morris Skinner's unpublished field notes. Singler and Picard (1980, Fig. 3, section 6) gave a brief summary of the local stratigraphy in one section, but the area is much more complex, as Skinner realized. We show three magnetostratigraphic sections in Figure 5 that incorporate Skinner's understanding of the stratigraphy. As in other White River sections, there is a late early Orellan normal magnetozone (from about 8–30 m [25-90 ft] above the PWL on all three sections), which we suggest correlates with Chron C13n. Thus, there is an 8-m (25-ft) interval representing late Chron C13r (between the PWL and the base of Chron C13n), which yields the last brontothere and also preserves the Chadronian-Orellan transition in more detail than any other place. In the Boner Ranch section (Fig. 5), there is also a late Chadronian normal magnetozone, which apparently correlates with Chron C15n (based on comparison with Douglas, Wyoming—see Evanoff et al., [1992], and Prothero and Swisher [1992]).

TAXONOMIC OVERVIEW

Before discussing the biostratigraphic data in each collecting area, it is necessary to summarize the current systematics of the key mammalian groups.


TYPE SCENIC
SHEEP MTN. TABLE
COTTONWOOD PASS
RED SHIRT TABLE
WOLF TABLE
CEDAR PASS
PIG DIG
PINNACLES
SAGE CREEK
VGP LATITUDE
Sharpe Fm.
Scenic Mbr.
Potsville Mbr.
Chadron Fm.
UN
MO
LN
UO
RFA
CHRON C13N

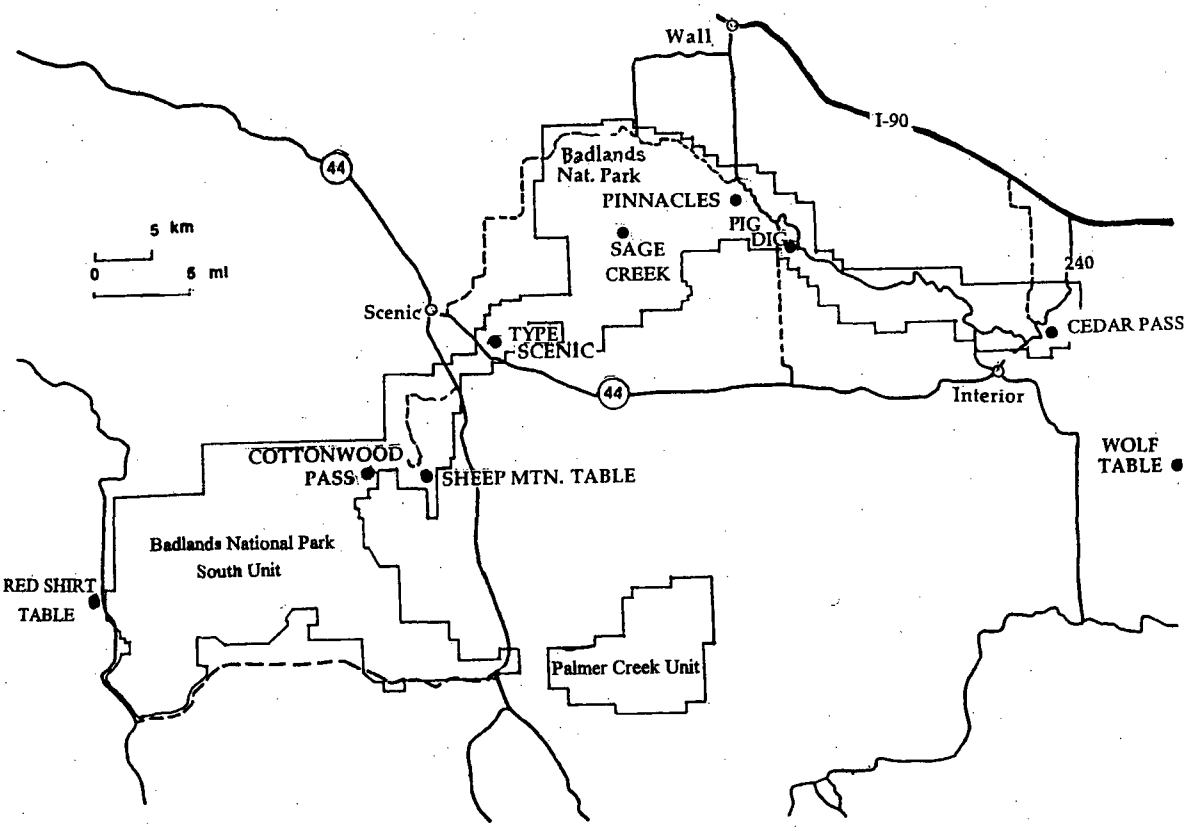


Figure 4 (this and facing page). Magnetic stratigraphy of sections in the Big Badlands of South Dakota (modified from Prothero and Swisher, 1992, Fig. 2.7). Stratigraphic terminology follows the unpublished field notes of Morris Skinner, except for Pinnacles, which follows Retallack (1983). Abbreviations: Lept. Nods. = *Leptauchenia* nodules; LN = lower nodular zone; MO = middle Oredon beds; RFA = Rockyford Ash; UN = upper nodular zone; UO = upper Oreodon beds. Solid circles = Class I sites of Opdyke et al. (1977), which are statistically separated from a random distribution at the 95% confidence level. Triangles = Class II sites, which could not be statistically analyzed because only two samples remained; the third was lost or crumbled. Open circles = Class III sites, in which two samples showed a clear polarity preference, but the third sample was divergent. "x" = site of indeterminate polarity. Locations of sections are as follows: Red Shirt Table: S1/2 sec. 34, T. 42 N., R. 47 W., Red Shirt SW 7.5′ Quadrangle, Shannon County, South Dakota; Cottonwood Pass: NE SE sec. 10, T. 4 S., R. 12 E., Heutmacher Table 7.5′ Quadrangle; Type Scenic area section (also called "Chamberlain Pass")—on the corner of four quadrangles: SE sec. 25, T. 3 S., R. 13 E., Scenic 7.5′ Quadrangle, and SW sec. 30 T. 3 S., R. 4 E., Quinn Table SW 7.5′ Quadrangle, and NE sec. 36, T. 4 S., R. 13 E., Sheep Mountain Table 7.5′ Quadrangle, and NW sec. 31, T. 4 S., R. 14 E., Imlay 7.5′ Quadrangle; Sage Creek: SE NW SW sec. 15 and NE NE NE sec. 21, T. 25 N., R. 15 E., Quinn Table NE 7.5′ Quadrangle; Sheep Mountain Table: SW NW NW sec. 10, to SE SE NW sec. 9, T. 43 N., R. 44 W., to SE NE SE sec. 4, T. 42 N., R. 44 W., to SE NE SW sec. 28, to SE NE NE sec. 28, T. 43 N., R. 44 W., Sheep Mountain Table 7.5′ Quadrangle; Pinnacles: Route indicated by Retallack (1983), from SE SE NE sec. 20 to SW NE NE sec. 20, to NE NW NE sec. 20, to SE SW sec. 17, T. 2 S., R. 16 E., Wall SW 7.5′ Quadrangle; "Pig Dig": NW NW sec. 34, T. 2 S., R. 16 E., Wall SW 7.5′ Quadrangle; Cedar Pass: Lower Scenic, NE NE SW sec. 27, Cottonwood SW 7.5′ Quadrangle; upper Scenic, SE NW NE to NW SE NE sec. 27, T. 3 S., R. 18 E., Cottonwood SW 7.5′ Quadrangle; Poleslide, NE NE NW sec. 35 to SE SE SW sec. 26, Interior 7.5′ Quadrangle; Wolf Table–Wanblee: SW NE NE sec. 31, T. 43 N., R. 87 W., School Section Butte 7.5′ Quadrangle, to NE NW SE sec. 32, to NE SW NW sec. 4, to SW SE SE sec. 9, to SW NE NE sec. 16, T. 42 N., R. 87 W., Wanblee NW 7.5′ Quadrangle.

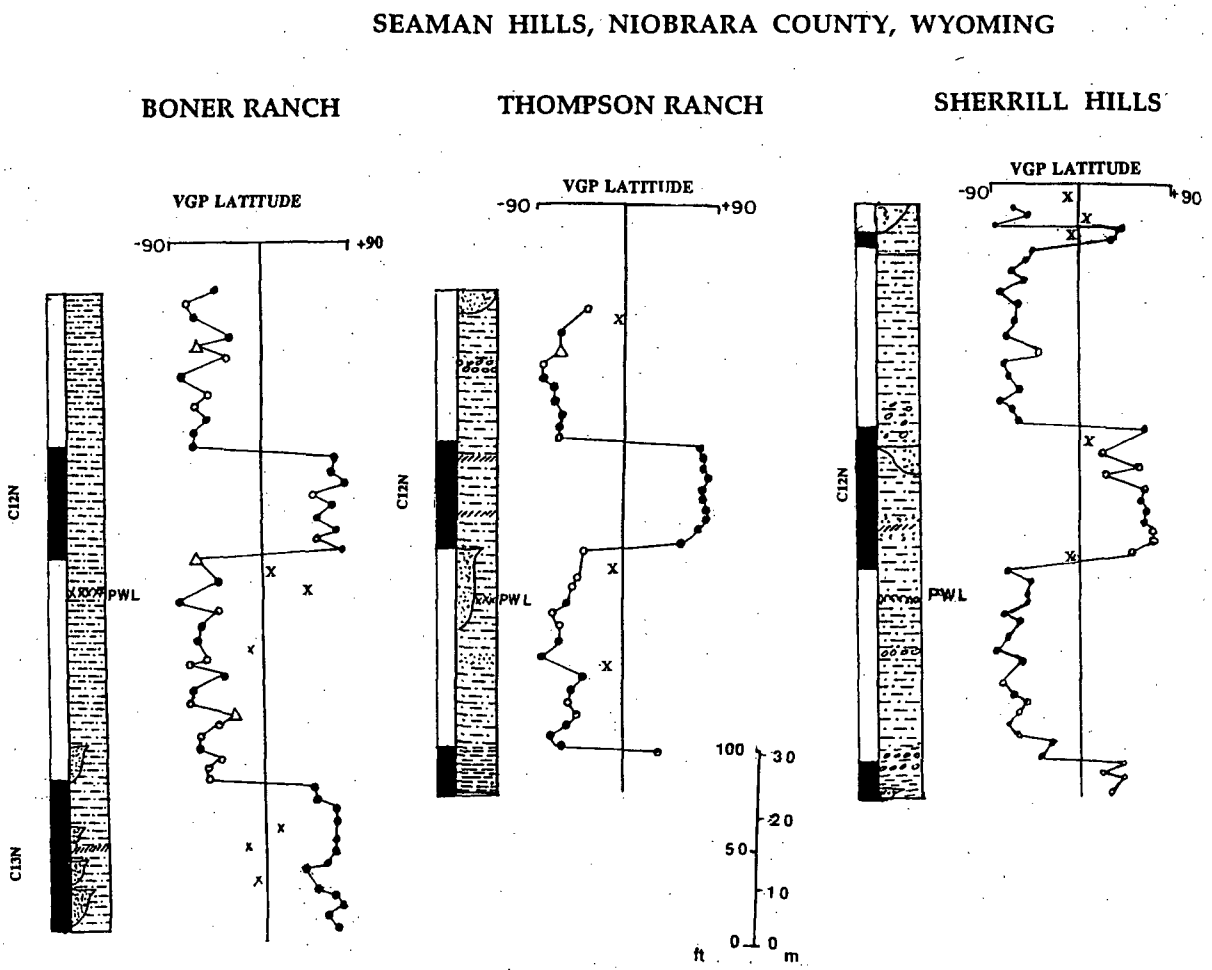


Figure 5. Magnetic stratigraphy of sections in the undifferentiated White River Formation in the Seaman Hills, Niobrara County, Wyoming. Stratigraphy based on the unpublished field notes of Morris Skinner. PWL = Persistent White Layer of Skinner; all other symbols as in Figure 4. Locations of sections are as follows: Thompson Ranch: NE sec. 28, T. 35 N., R. 61 W., Shepherds Point 7.5′ Quadrangle; Boner Ranch: SW sec. 9, T. 35 N., R. 61 W., South Oat Creek 7.5′ Quadrangle; Sherrill Hills: SW NE SW to SE SE SE sec 17, T. 34 N., R. 60 W., Sherrill Hills Quadrangle.

Artiodactyls

Oreodonts are by far the most common larger mammal fossils in the Brule Formation. Scott et al. (1941) recognized only three genera and less than a dozen species from the White River Group, and Thorpe (1937) was similarly conservative. Unfortunately, the Schultz and Falkenbach (1956, 1968) monographs obscured the taxonomy of oreodonts. These authors oversplit the family into dozens of invalid species, genera, and even subfamilies based on features due to post-mortem deformation (for example, *Platyochoerus*, or "flat pig," was a dorsoventrally crushed *Miniochoerus*, and *Stenopsochoerus*, or "narrow pig," was a laterally crushed *Miniochoerus*). Their taxonomy is also plagued by inadequate statistical comparisons, taxonomic definitions that were based on stratigraphic boundaries with no morphologic change, and so on. The problems with their taxonomy have been widely recognized (e.g., Harksen and Macdonald, 1969a, p. 13; Lander, 1977; Savage and Russell, 1983, p. 195, footnote 2; Gustafson, 1986, p. 16; Emry et al., 1987, p. 140). Stevens and Stevens (1996) and CoBabe (1996) have recently revised the group, although a full revision of post-Whitneyan oreodonts is not yet in print.

Stevens and Stevens (1996) have reduced the multiple subfamilial lineages of Schultz and Falkenbach (1968) to two lineages in the early Oligocene. The main lineage is the most common of all Badlands fossils, the larger oreodont *Merycoidodon* (incor-

rectly referred to "*Prodesmatochoerus*" by Lander, 1977). Although this genus was grossly oversplit by Schultz and Falkenbach (mostly because they are so common, and intraspecific variability is high), Stevens and Stevens (1996) have recognized only *Merycoidodon culbertsoni* in the late Chadronian and early Orellan. In the late early Orellan (early Chron C12r, upper nodules or "middle Oreodon beds" in the Badlands, and somewhere within Orella C at Toadstool Park), most *Merycoidodon* specimens acquire inflated auditory bullae. This change is the criterion for the recognition of *M. bullatus* of Stevens and Stevens (1996), known as *Eporeodon bullatus* in older literature (e.g., Thorpe, 1937; Scott, 1941), and equivalent to numerous Schultz and Falkenbach (1968) taxa.

Large oreodonts referable to *M. major* appear in the late Whitneyan (late Chron C12r, or "*Leptauchenia* nodules" in the Big Badlands, and Whitney B at Toadstool Park). They are the most abundant and reliable indicator of the late Whitneyan. Stevens and Stevens (1996) reported *Mesoreodon minor* as a zonal indicator of the latest Whitneyan–earliest Arikareean ("brown siltstone" and lower Sharps Formation—see Tedford et al., 1996).

In addition to the *Merycoidodon* lineage of larger oreodonts, another lineage was present in the Orellan. Long known as *Merycoidodon gracilis* in older literature (e.g., Thorpe, 1937; Scott, 1941), Stevens and Stevens (1996) referred all these taxa to *Miniochoerus*. This same lineage was split into multiple genera and dozens of species by Schultz and Falkenbach (1956, 1968) and incorrectly referred to *Oreonetes* by Lander (1977). According to Stevens and Stevens (1996), late Chadronian *Miniochoerus chadronensis* is significantly smaller than contemporary *Merycoidodon culbertsoni*, but otherwise it is difficult to distinguish. In the early Orellan, *Miniochoerus* undergoes a dwarfing trend (30% size reduction in about a million years), so that the size of specimens along this chronocline can be used to zone the early Orellan (Prothero and Heaton, 1996, Fig. 2). Stevens and Stevens (1996) referred specimens from just below the PWL to ~17 m (50 ft) above the PWL at Lusk to *M. chadronensis* (mean M^{1-3} length, 47 mm; observed range, 44–50 mm). Stevens (1977) originally referred specimens that were intermediate in size (occurring between 0 and 17 m (0 and 50 ft) above the PWL at Lusk) not to *M. chadronensis*, but to *M.* "*douglasensis*" (mean M^{1-3} length, 42 mm; observed range, 39–45 mm). The smaller miniochoeres (from the late early Orellan) were referred to *M. gracilis* (mean M^{1-3}length, 34 mm; observed range, 31–37 mm) and *M. affinis* (mean M^{1-3} length, 38 mm; observed range, 36–41 mm). *M. affinis* first appears about 17 m (50 ft) above the PWL at Lusk, Wyoming (mid-Chron C13n), and *M. gracilis* first appears about 27 m (80 ft) above the PWL at Lusk (early Chron C12r). Both species range through the early Orellan without further change. In the late Orellan (mid Chron C12r, upper nodular zone in the Big Badlands, Orella D at Toadstool Park [=lower Whitney of LaGarry, this volume], upper Fitterer Ranch in North Dakota), a slightly larger but more advanced species, *M. starkensis*, appears (Stevens and Stevens, 1996). This is the last species of *Miniochoerus*, persisting until the mid-Whitneyan (late Chron C12r, lower "*Leptauchenia* nodules" in the Badlands, and Whitney A in Nebraska).

In addition to merycoidodontines and miniochoerines, a third subfamily of oreodonts is critical to Brule biostratigraphy. The small hypsodont oreodonts known as leptauchenines are so abundant in the Whitneyan that they were the original indicator of that age. CoBabe (1996) reduced 7 genera and 31 species recognized by Schultz and Falkenbach (1968) to only 2 genera (*Leptauchenia, Sespia*), with three species between them. Most specimens are referable to *L. decora*, whose first abundant occurrence marks the beginning of the Whitneyan, although a few rare specimens ("*Pseudocyclopidius*" and "*Hadroleptauchenia*" of Schultz and Falkenbach, 1968) are known from the Orellan. The beginning of the Whitneyan also yields the larger species *Leptauchenia major* (formerly *Cyclopidius* of Thorpe, 1937, and Scott, 1941, and oversplit by Schultz and Falkenbach, 1968). *Sespia* first appears in the early Arikareean (Tedford et al., 1996).

Although oreodonts are the most abundant artiodactyl fossils, others are also important. Heaton and Emry (1996) have updated the taxonomy of the common ruminant *Leptomeryx*. Most of its species occur in the Chadronian, but the first appearance of *L. evansi* is an indicator of the earliest Orellan. Heaton and Emry (1996) recognize no other valid Orellan species, despite the number of names that have been applied to this abundant group.

The small ruminant *Hypertragulus calcaratus* first appeared at the beginning of the Orellan. Tedford et al. (1996) mark the beginning of the Arikareean by the small ruminant *Nanotragulus loomisi*. The tiny, hypsodont ruminant *Hypisodus* was reviewed by Haake and Galbreath (1979), but so far has not proven useful for biostratigraphy.

Prothero (1996b) has reviewed the camels of the White River Group. The primitive Chadronian species *Poebrotherium eximium* (distinguished by its short diastema between P_1 and P_2) last occurs about 13 m (40 ft) above the PWL at Lusk, and about 27 m (80 ft) above the 5 tuff at Douglas (i.e., within Chron C13n in both places), so its LAD is a marker of the late early Orellan. *Paratylopus labiatus* first occurs just before Chron C13n at both Lusk and Douglas.

The horned tylopod *Protoceras* was the original basis for the upper Poleslide "*Protoceras* channels" in the Big Badlands. *Protoceras* is unknown outside the Badlands, so its value as an index fossil is limited.

There is currently no up-to-date published taxonomy of the entelodonts, anthracotheres, tayassuids, or the leptochoerines. These artiodactyls, however, are too rare in most White River localities to be very useful in biostratigraphy, although their ranges are mentioned below when they are known.

Lagomorphs

Korth and Hageman (1988) recently reviewed the lagomorphs from the Toadstool Park area, Nebraska. Korth (1989) renamed Orella A (= upper Chadron Formation of Terry and LaGarry, this volume) as the *Palaeolagus hemirhizis* zone. This

seems to be an improvement over the lithostratigraphically based zonations of Schultz and Falkenbach (1968), but there are problems with this taxon and with the zone itself.

As indicated above, none of the University of Nebraska State Museum collections are zoned *within* Orella A, so it is impossible to tell exactly where the FAD of *Palaeolagus hemirhizis* is located (or any other taxa that might define the earliest Orellan at Toadstool Park). Such imprecision is no longer adequate when higher resolution is possible in the Lusk and Douglas areas. In addition, the upper 13 feet of Orella A are correlative with Chron C13n, so they are not earliest Orellan, but late early Orellan. This makes it questionable whether the Toadstool Park section can be used for high-resolution biostratigraphy.

Besides the lithostratigraphic problems, there are also taxonomic problems with *P. hemirhizis*. Korth and Hageman (1988) erected the taxon for Orella A specimens that are intermediate between Chadronian *P. temnodon* and Orella B *P. haydeni*. *Palaeolagus hemirhizis* was defined on the basis of its intermediate size (but the size distributions between the three taxa overlap greatly) and on the presence of P^3–M^2 roots on ~50% of the specimens. However, Prothero has closely examined and measured the precisely zoned lagomorph specimens in the Frick Collection, and found that these criteria are very difficult to use. For example, earliest Orellan lagomorphs in the Munson Ranch in Sioux County, Nebraska (Prothero, 1996a), and in the Lusk and Douglas areas show the same range of size variation reported by Korth and Hageman (1988), but at each level, they can be assigned to *P. haydeni* or to *P. temnodon* based on the presence or absence of roots on their upper cheek teeth. We suspect that *P. hemirhizis* is an artificial construct of two different species of rabbits, which were lumped together because of the poor resolution of the University of Nebraska State Museum collections.

Even if the taxonomy of *P. hemirhizis* is accepted, there are other problems with its first occurrence marking the beginning of the Orellan. Storer (1994) reported specimens of *Palaeolagus* cf. *P. hemirhizis* from the Chadronian Kealey Springs West local fauna of the Cypress Hills Formation in southwestern Saskatchewan. In fact, Korth and Hageman (1988, p. 147) conceded that *P. hemirhizis* may occur in the Chadronian of Nebraska and Colorado as well. Thus, we hesitate to name the earliest Orellan biostratigraphic zone after a taxon of doubtful validity that probably occurs in the Chadronian.

Palaeolagus intermedius is much rarer, but in most localities (e.g., Lusk and Douglas, Wyoming; Munson Ranch, Nebraska; Cedar Creek, Colorado) its FAD is also an indicator of the earliest Orellan, since it first appears 3 m (10 ft) above the PWL at Lusk and 3 m (10 ft) above the 5 tuff at Douglas. Korth and Hageman (1988) have only recorded it from Orella D (= lower Whitney of Terry and LaGarry, this volume) at Toadstool Park, although Prothero's examination of the University of Nebraska State Museum collections turned up specimens assigned to this taxon from Orella A–C.

The distinctively small species *Palaeolagus burkei* first occurs in mid-Chron C12r (at the base of the upper nodules in the Badlands and in Orella C at Toadstool Park). It is a good indicator for the early late Orellan.

Korth and Hageman (1988) reported *Megalagus turgidus* from the entire Orella Member, and point out that it increases in size through the Orellan. In the Frick Collections, *M. turgidus* first appears 10 m (30 ft) below the PWL at Lusk and 10 m (30 ft) below the 5 tuff at Douglas, so its FAD could be used as an indicator of the latest Chadronian. However, it is difficult to distinguish from the typical Chadronian species *M. brachyodon*.

Litolagus molidens is an extremely rare rabbit. Four specimens are known between 13 and 30 m (40–90 ft) above the PWL at Lusk (late C13r-early C13n) and one specimen from 10 m (30 ft) below the 5 tuff at Douglas (late C13r). Its rarity makes this range discrepancy less surprising, but its presence does seem to indicate latest Chadronian–earliest Orellan.

Rodents

One of the most common Orellan rodents is the ischyromyids. Their taxonomy has recently been revised by Heaton (1993, 1996), who found that both the larger *Ischyromys typus* and the smaller *I. parvidens* are present in the earliest Orellan, although the latter species is far more abundant. The supposed anagenetic size increase reported by Howe (1966) is actually the replacement of *I. parvidens* (which last appears in latest Chron C13n, or in Orella B at Toadstool Park, and just below the upper nodules in the Big Badlands) by the larger species *I. typus*. Thus, the last appearance of *I. parvidens* (leaving only *I. typus*) is a good indicator of the late early Orellan. *I. typus* continues to be abundant throughout the late Orellan, so its last appearance is also a good marker of the Whitneyan.

Another common rodent group is the eumyine cricetids, which have been reviewed by Wood (1980), Martin (1980), and Korth (1981). The most common species is *Eumys elegans*, which first occurs in mid-Chron C13n (27 m or 80 ft above the PWL at Lusk, Wyoming, Orella B at Toadstool Park). Korth (1989) used it as the nominal taxon of his *Eumys elegans* zone (Orella B-C). However, a number of specimens supposedly referable to *E. elegans* have been reported from the Chadronian. Some of these may be "float" from the Orellan (according to Ostrander, 1985) or may actually be referable to *Scottimus viduus* (Martin, 1980; Korth, 1981). Storer (1994) reported *E. elegans* from the late Chadronian of Saskatchewan, but it is possible that this locality is actually Orellan with some late Chadronian taxa that persist unusually late in Saskatchewan. *Eumys* also occurs in the early Orellan (Chron C13r) of Montana (Tabrum et al., 1996).

A number of other species of eumyines are recognized by Martin (1980) and/or Korth (1981). *Eumys obliquidens* is restricted to the late early Orellan (mid-Chron C13n). *E. parvidens* is restricted to the early late Orellan (early Chron C12r, Orella C at Toadstool Park, upper nodules in the Big Badlands). *Wilsoneumys planidens* is also restricted to the late Orellan (mid-Chron C12r, Orella D at Toadstool Park). *Scottimus lophatus* is restricted to the later Whitneyan (latest Chron C12r, upper Whit-

ney B and Whitney C at Toadstool Park, late Whitneyan in Slim Buttes, South Dakota, according to Lillegraven, 1970).

Diplolophus insolens is a rare (only nine specimens are known) enigmatic rodent whose occurrence was the basis for the *D. insolens* bench marker bed that separates Orella C and D at Toadstool Park (Barbour and Stout, 1937; Schultz and Stout, 1955; Wood, 1980; see LaGarry, this volume). Korth (1989) renamed Orella D as the *D. insolens* zone. It also occurs early in Chron C12r in Colorado (Galbreath, 1953), and at an unknown horizon in the Big Badlands.

Korth (1989) summarized a number of other rodent occurrences that could be used to characterize zones of the Orellan. Unfortunately, their exact occurrence in the Orellan is difficult to reconstruct, since they are zoned only as Orella A, Orella B, and so on. In addition, these rodents have not yet been studied in the Frick Collection, so their stratigraphic ranges in the important Lusk and Douglas areas are presently undocumented. Once these rodents have been adequately studied outside the Toadstool Park area, they may prove useful to Orellan-Whitneyan biostratigraphy.

Perissodactyls

The species-level taxonomy of brontotheres is still problematic since the oversplitting of Osborn (1929). Mader (1989) has revised the group at the generic level, but the number of Chadronian brontothere species and criteria for their recognition are still unknown.

Prothero and Shubin (1989) recently revised the horses from the Orellan and Whitneyan. *Mesohippus bairdi* and *M. westoni* range up from the middle Chadronian, and last occur in the late Orellan (mid-Chron C12r, upper nodules in the Big Badlands, Orella C–D at Toadstool Park). *M. exoletus* first occurs 2 m (5 ft) below the top of the Chadron Formation in the Big Badlands and Lusk, and last occurs in the late Orellan (mid-Chron C12r, upper nodules in the Big Badlands, Orella C–D at Toadstool Park). *M. barbouri* ranges from the late early Orellan to the late Orellan (early Chron C12r, lower nodules to upper nodules in the Big Badlands, and also in the late Orellan Harvard Fossil Reserve in Torrington, Wyoming: see Schlaikjer, 1935). *Miohippus obliquidens* is an extremely long-ranging taxon, occurring from the late Chadronian through the late Whitneyan.

Prothero and Shubin (1989) referred certain late Chadronian specimens to *Miohippus assiniboiensis*, but Storer and Bryant (1993) have shown that they are correctly referred to *M. grandis*, which ranges up from the early Chadronian (Calf Creek Local Fauna in Saskatchewan) and last occurs in the upper Chadron Formation in the Big Badlands (Clark et al., 1967) and 10 m (30 ft) above the PWL at Lusk and 10 m (30 ft) below the 5 tuff in Douglas. Thus, its LAD could be used to recognize the Chadronian-Orellan boundary. *Miohippus gidleyi* and *M. intermedius* are rarer, but both are restricted to the late Whitneyan (Chron C12n, "*Leptauchenia* beds" in the Big Badlands).

Prothero (1996c) revised the rhinocerotid *Hyracodon*. In the late Chadronian, *H. priscidens* is replaced by the type species, *H. nebraskensis*, which ranges to the end of the Whitneyan with no observable change. However, in the early Whitneyan ("upper Oreodon beds" in the Big Badlands, late Chron C12r), a larger, more advanced species referable to *H. leidyanus* appears; it also persists to the end of the Whitneyan. The extinction of both of these *Hyracodon* species is one good indicator of the Whitneyan-Arikareean boundary (Tedford et al., 1996).

The common Badlands rhinocerotid, *Subhyracodon occidentalis*, also first occurs in the late Chadronian and shows no noticeable change in the Orellan (Prothero et al., 1989). In the early Whitneyan, it is replaced by *Diceratherium tridactylum*, but all of these rhinos are too rare to be useful to biostratigraphy. Some rare rhinocerotids have unusual distributions. For example, both *Penetrigonias dakotensis* and *Amphicaenopus platycephalus* occur in the Chadronian and Whitneyan, but not in the intervening Orellan. However, both are very rare, so this distributional anomaly is probably due to sampling error.

The amynodontid rhinocerotoid *Metamynodon planifrons* was the basis for the name of the *Metamynodon* channels in the lower Scenic Member in the Big Badlands. Like *Protoceras*, however, its utility for correlation is limited by the fact that it is restricted to South Dakota.

White River tapirs have not been revised recently, but their extremely rarity throughout the Oligocene limits their biostratigraphic utility.

Creodonts and carnivorans

Mellett (1977) revised the common White River creodont *Hyaenodon*, using the large collections compiled by the Frick Laboratory. Both *H. horridus* and *H. crucians* first occur in the middle Chadronian. The common small species *H. crucians* last occurs near the base of the upper nodules (mid Chron C12r), so its LAD is an indicator of the latest Orellan. The largest species, *H. horridus*, last appears in the mid-Whitneyan (late Chron C12r, upper *Oreodon* beds in the Big Badlands, mid-Whitney B in Toadstool Park), so its LAD is a good indicator of the late Whitneyan. *H. brevirostrus* is restricted to the early Arikareean.

Wang (1994) and Wang and Tedford (1996) have recently revised the early North American canids. They have shown the group to be much more diverse than previously expected, with a number of species occurring in the Orellan and Whitneyan. The commonest canid is *Hesperocyon gregarius*, which ranges from the Duchesnean through Whitneyan. "*Mesocyon*" *temnodon* first occurs in the Orellan in South Dakota, and in the Whitneyan in Nebraska and Colorado. *Cynodesmus thooides* first occurs in the Whitneyan of Nebraska, South Dakota, and Montana. *Osbornodon renjiei* occurs in the Orellan of North Dakota, and in the Whitneyan of South Dakota and Nebraska. *O. sesnoni*, *Parenhydrocyon josephi*, and *Ectopocynus antiquus* are known from the Whitneyan of South Dakota. The small, primitive borophagine *Cormocyon pavidus* occurs in the Orellan of Colorado and in the Whitneyan of South Dakota and Nebraska. *Oxetocyon cuspidatus* is restricted to the Whitneyan of South Dakota and Nebraska.

The amphicyonids, or "beardogs," have been updated by Hunt (1996). The last appearance of *Brachyrhynchocyon* may occur about 3 m (10 ft) above the PWL at Arner Ranch (UNSM locality Sx-29) in Sioux County, Nebraska. *Daphoenictis* last appears at the end of the Chadronian, and *Paradaphoenus* first appears in the earliest Orellan. *Daphoenus vetus* and *D. hartshornianus* range from the late Chadronian through early Orellan.

Considerably more common in the White River Group are the cat-like nimravids, which have been recently reviewed by Bryant (1996). *Dinictis felina* ranges from the mid-Chadronian to the late Whitneyan. *Pogonodon platycopis* occurs in the Orellan of South Dakota and in the Whitneyan of South Dakota and Nebraska. *Hoplophoneus primaevus* occurs throughout the Orellan and Whitneyan (and may range into the Chadronian). *H. occidentalis* occurs in the late early Orellan of South Dakota (lower nodules), North Dakota, Nebraska, and Wyoming, and in the Whitneyan of South Dakota. *H. sicarius* is known from three specimens, one from the lower nodules in the Badlands, and two from unknown levels in Wyoming. *H. dakotensis* first occurs in the late Whitneyan "*Protoceras* beds" of South Dakota and ranges into the early Arikareean. *Eusmilus cerebralis* first occurs in the late Whitneyan of South Dakota also. *Nimravus brachyops* first occurs in the late Whitneyan of Nebraska, South Dakota, and possibly Saskatchewan.

In addition to amphicyonids, canids, and nimravids, there are a number of smaller White River carnivorans that were recently reviewed by Baskin and Tedford (1996). Most are known from only a few specimens, so their biostratigraphic utility is limited. For example, *Drassonax harpagops* is known from two specimens, one from the Scenic Member in the Big Badlands and one from the Orella Member in Nebraska; both are from unknown levels. *Palaeogale sectoria* occurs in the Orellan of Colorado, Montana, and South Dakota. The majority of the taxa discussed by Baskin and Tedford (1996), however, are from the Chadronian.

Insectivorans

Numerous large collections of insectivorans from the Brule Formation exist, but their systematics have not been adequately published to be useful for biostratigraphy. For example, M. J. Novacek (personal communication) has revised the leptictids, and recognizes *Leptictis dakotensis* (late Chadronian to late Whitneyan), as well as some Chadronian taxa. The apternodontids are also abundant, but lack an updated analysis. *Centetodon* was recently revised by Lillegraven et al. (1981), but it is also very rare and sporadically distributed. The insectivorans might prove biostratigraphically useful once their systematics is updated, and when enough screen-washing is done throughout the section to get a more uniform sampling of micromammals (Emry et al., 1987).

BIOSTRATIGRAPHY

Chadronian-Orellan boundary and earliest Orellan

The Chadronian-Orellan transition is best preserved in detail in the Lusk (Fig. 6) and Douglas (Evanoff et al., 1992) sections in Wyoming. The truncation of the base of the biostratigraphic ranges of most early Orellan taxa in the Big Badlands sections (Fig. 7) is further evidence of an unconformity beneath the lower nodules. Thus, we will examine key biostratigraphic datum levels between the PWL (the traditional top of the Chadron Formation in Wyoming) and the base of Chron C13n (which is late early Orellan), mostly in the Lusk and Douglas sections.

This earliest Orellan time interval is also preserved in "Orella A" (= upper Chadron of Terry and LaGarry, this volume) in the Toadstool Park area, Sioux County, Nebraska (Schultz and Stout, 1955, 1961). Schultz and Falkenbach (1968) based their "Oreodont Faunal Zone Orella A" on this unit, and Korth (1989) also based his biostratigraphy on the Toadstool Park sequence and the University of Nebraska State Museum collections. Indeed, the lower 7 m (20 ft) of "Orella A" at Toadstool Park (Prothero and Swisher, 1992, Fig. 2.5) does correlate with Chron C13r above the PWL (UPW of Schultz and Stout, 1955), so lower Orella A fossils are equivalent to those found in the lower 8 m (25 ft) at Lusk, and the lower 13 m (40 ft) at Douglas. Unfortunately, the existing University of Nebraska State Museum collections do not indicate at what level specimens were found *within* "Orella A." So specimens with the "Orella A" field data might come not only from the earliest Orellan (late Chron C13r) part of the unit (the lower 7 m, or 20 ft) but also from the late early Orellan (early Chron C13n) part of Orella A (the upper 4 m, or 13 ft). This unfortunate lack of stratigraphic resolution and mixing of collections from different levels makes the current Toadstool Park collections less than ideal for examination of the detailed changes in the fauna.

As discussed above, the LAD of brontotheres was part of the Wood Committee's (1941) original definition of the Chadronian. However, these last few brontothere specimens occur very sporadically. Even though they have long been part of the classic concept of the Chadronian, brontotheres are so scarce that they are a very impractical marker of the end of that land-mammal "age." In addition, their last appearance is highly variable within late Chron C13r. At Scottsbluff (Prothero and Swisher, 1992, Fig. 2.6) they occur just above the PWL (or UPW). At Lusk, they occur in a channel cut down from 8 m (25 ft) above the PWL, and thus at the C13r–C13n boundary (~33.5 Ma). But in Douglas, they last occur 3 m (15 ft) below the C13r–C13n boundary. This lack of consistency between last occurrences in various areas makes the brontothere LAD less useful than other datum levels.

According to Stevens and Stevens (1996), there are two valid generic lineages of oreodonts in the Orellan. The larger is the well-known and abundant *Merycoidodon culbertsoni*. Although specimens are extremely abundant and highly variable, there is no consistent trend that allows any other species in this lineage to be distinguished in the late Chadronian or early Orellan. Thus, the merycoidodont "faunal zones" of Schultz and Falkenbach (1968) are invalid. However, the other, smaller lineage of Orellan oreodonts, the miniochoeres, do show changes that are biostratigraphically useful, as discussed above (Fig. 6).

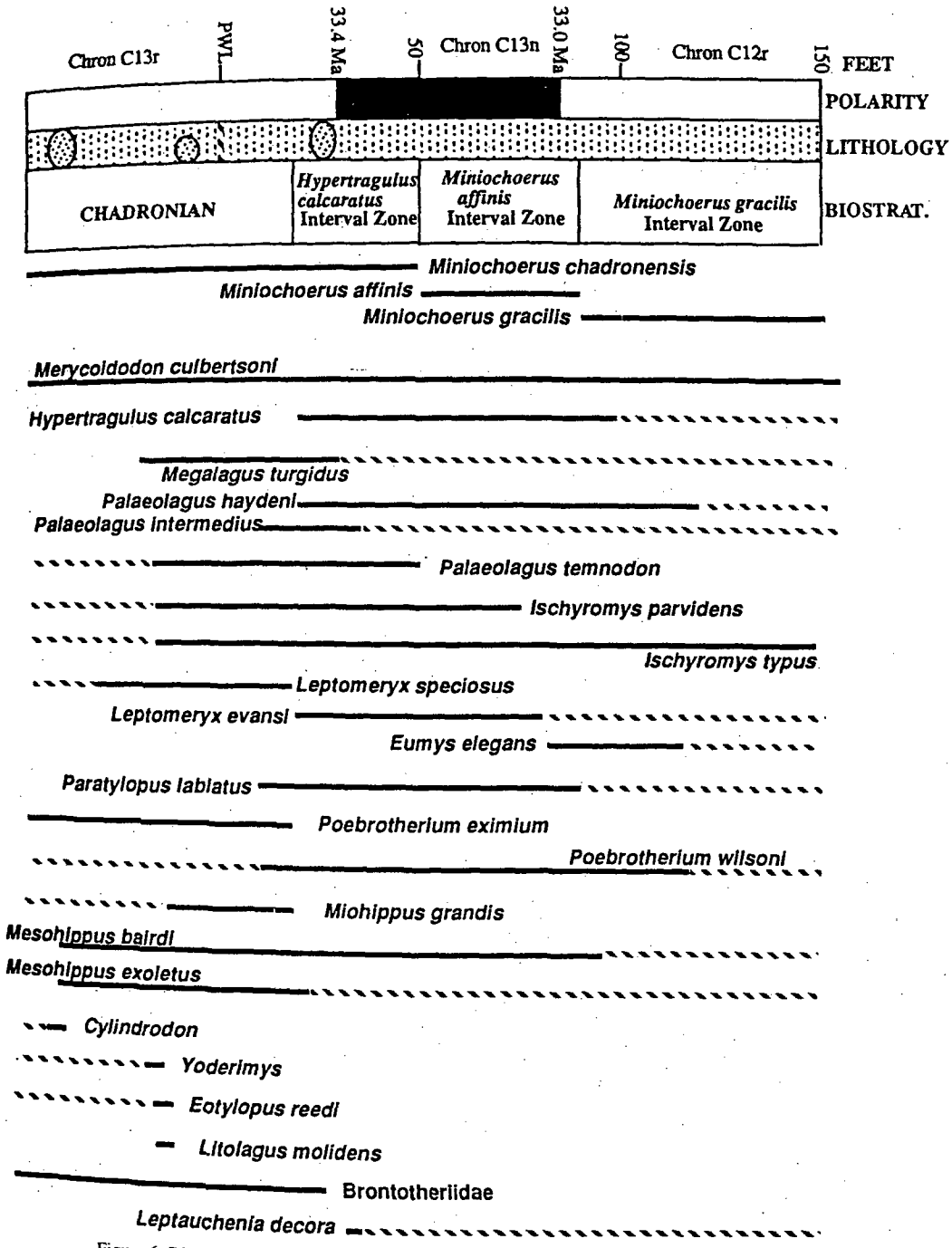


Figure 6. Biostratigraphy of key taxa in the undifferentiated White River Formation, Seaman Hills area, Niobrara County, Wyoming. Magnetic stratigraphy and lithostratigraphy as in Figure 5. Solid lines show documented ranges of taxa; dashed lines are inferred range extensions based on occurrences elsewhere in the White River Group (modified from Prothero, 1982).

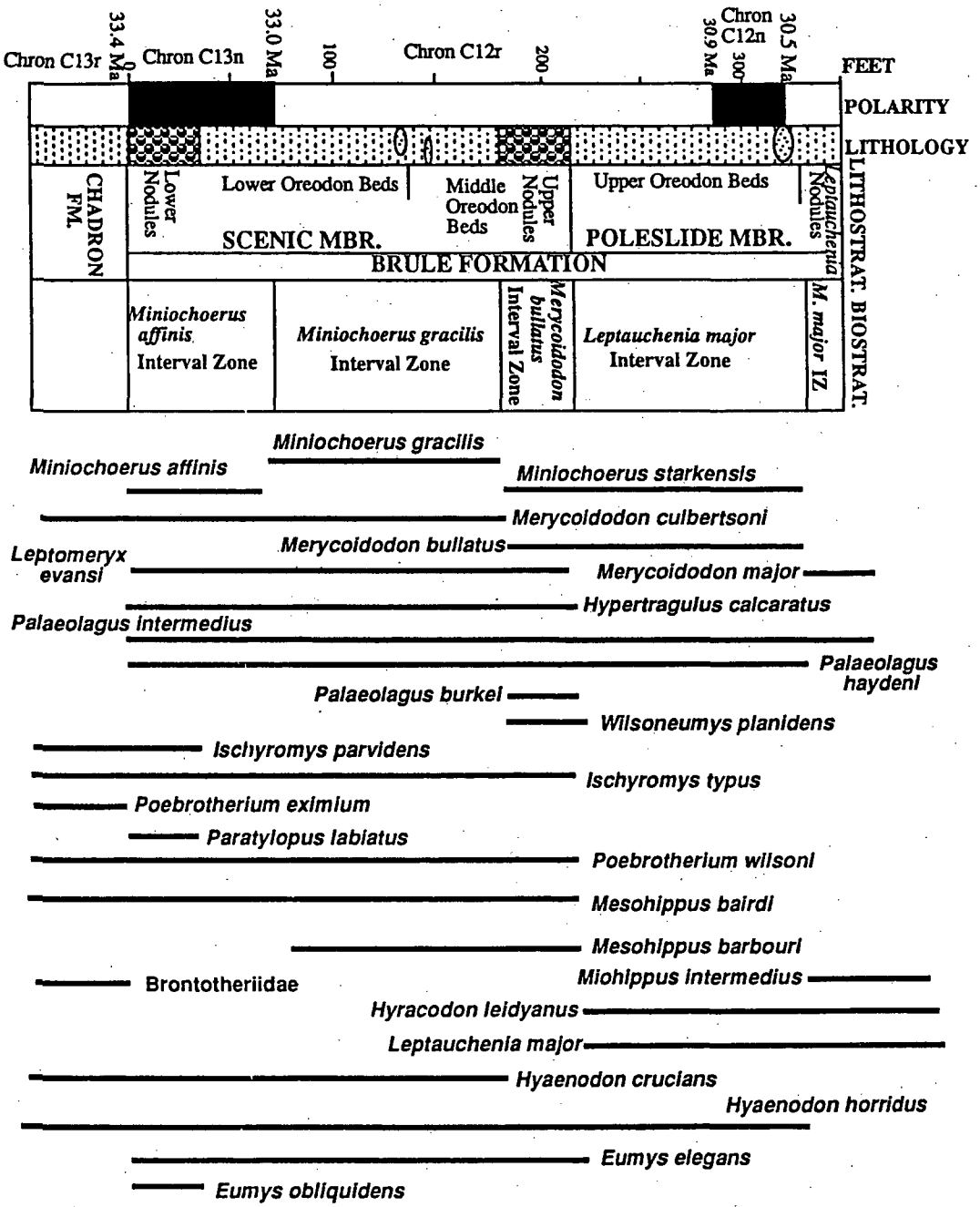


Figure 7. Biostratigraphy of key taxa in the Sheep Mountain Table–Cottonwood Pass–Big Corral Draw areas of the Big Badlands, South Dakota (based on the Frick Collection). Magnetic stratigraphy and lithostratigraphy as in Figure 4. Note the truncation of ranges at the base of the lower nodules and the absence of early Orellan taxa (*Hypertragulus calcaratus* Interval Zone), indicating an unconformity at this level. Based on data in Prothero (1982).

Next to oreodonts, the most common mammals in the Orellan are the small deer-like artiodactyls *Leptomeryx* and *Hypertragulus*. Heaton and Emry (1996) recognized a gradual transition from *L. speciosus* of the late Chadronian to *L. evansi* of the early Orellan. Although several highly variable characters are used to distinguish the two species, one of the easiest to recognize is the lingual to labial shift of the ridge behind the P_3 protoconid. Prothero (1982) found that this shift first occurs in specimens about 7 m (20 ft) above the PWL at Lusk and 13 m (40 ft) above the 5 tuff at Douglas. Given the great number of *Leptomeryx* lower jaws in all early Orellan localities, this biostratigraphic datum is potentially very useful for recognizing the beginning of the Orellan.

Even more useful is the FAD of *Hypertragulus calcaratus*. Unlike *Leptomeryx evansi*, which is only subtly distinguished from the similar *L. speciosus* of the Chadronian, *H. calcaratus* appears in the White River group with no late Chadronian predecessor. *Hypertragulus heikeni* (Ferrusquia-Villafranca, 1969) was reported from the early Chadronian Rancho Gaitan local fauna in northeastern Chihuahua, Mexico, but no *H. calcaratus* have been reliably reported from *in situ* Chadronian deposits in the White River Group. (The reports by Ostrander [1985] and Gustafson [1986] are apparently loose Orellan specimens reworked out of Pleistocene terraces in the Chadron area). *H. calcaratus* appears suddenly about 7 m (20 ft) above the PWL at Lusk and 7 m (20 ft) above the 5 tuff at Douglas, so it is very easy to recognize and use as the indicator of the beginning of the Orellan. It also first appears at the beginning of the Orellan in northeastern Colorado (Galbreath, 1953), Slim Buttes in northwestern South Dakota (Lillegraven, 1970), and the Big Badlands and Toadstool Park (Prothero, 1982).

As Emry et al. (1987, p. 138) pointed out, the Chadronian-Orellan transition is a subtle one. "At most localities there is no significant lithologic change across the boundary, no evidence of a significant break in the stratigraphic record, and very little change in the fauna, except that in the Brule fossil mammal remains are much more abundant and the titanotheres that were a prominent part of the Chadronian fauna are missing." Emry et al. (1987, p. 139) recommended that the boundary "be based on biostratigraphic criteria and that it should be placed at the most prominent and most widely recognizable faunal break, which may or may not coincide with lithostratigraphic boundaries." Clearly, the Chadron-Brule contact cannot be used as the boundary, since it is time-transgressive in some places, such as the Big Badlands. Where it is marked by a synchronous volcanic ash (such as the PWL), there are no biostratigraphic datums that coincide with it precisely.

Although the boundary is subtle with no major "faunal break," a number of taxa can be used to characterize the Chadronian-Orellan transition and the earliest Orellan. The most useful of these are the FADs of *Hypertragulus calcaratus, Leptomeryx evansi, Palaeolagus intermedius*, and small *Miniochoerus chadronensis* (=*M.* "*douglasensis*" of Stevens, 1977), and the LAD of *Poebrotherium eximium* and *Miohippus grandis*. Murphy (1977) and Woodburne (1977, 1987) recommended that biostratigraphic zones be *characterized* by a number of distinctive taxa, but *defined* on the basis of a single taxon to prevent confusion and ambiguity. We recommend that the beginning of the Orellan be defined by the FAD of *Hypertragulus calcaratus*, and characterized by the rest of the taxa listed above. Based on this criterion, the Chadronian-Orellan boundary occurs about 7 m (20 ft) above the 5 tuff at Douglas, Wyoming, 7 m (20 ft) above the PWL in the Seaman Hills, Wyoming, and somewhere within the lower 7 m (20 ft) of Orella A in Toadstool Park, Nebraska. In compliance with the 1983 North American Code of Stratigraphic Nomenclature, we formally designate the earliest Orellan as the *Hypertragulus calcaratus* Interval Zone, using the concept of interval zones that are based on the successive first occurrences of index taxa as described in the Code (NACSN, 1983, Fig. 4C1). This type of zone was also used by Archibald et al. (1987) to zone the Paleocene. We designate the type section for the *Hypertragulus calcaratus* Interval Zone as the strata 7 to 17 m (20–50 ft) above the 5 tuff in the Reno Ranch East section (Evanoff et al., 1992, Fig. 6.6), located in NW NW sec. 32, and W1/2 sec. 29, T. 32 N., R. 70 E., Irvine 7.5′ Quadrangle, Converse County, Wyoming. The top of this zone is marked by the FAD of the name-bearer of the overlying zone, *Miniochoerus affinis*.

Chronostratigraphically, the Chadronian-Orellan boundary falls midway between the 5 tuff (33.9 Ma, according to Swisher and Prothero, 1990, and Prothero and Swisher, 1992) and the Chron C13r–C13n boundary (33.5 Ma, according to Berggren et al., 1995). Thus, its age would be ~33.7 Ma if sedimentation rates are assumed to be constant. However, the new $^{40}Ar/^{39}Ar$ dates reported by Obradovich et al. (1995) may recalibrate this geochronologic estimate to a slightly younger age.

Late early Orellan

A number of faunal events can be used to characterize the late early Orellan. The FAD of *Miniochoerus affinis, Eumys elegans* (in the White River Group, but not in Montana or Saskatchewan), *Pelycomys brulanus, Adjidaumo minutus, Cedromus wardi*, and *Hoplophoneus occidentalis* all occur in strata that correlate with early Chron C13n; *Ischyromys parvidens* last occurs within this interval. Korth (1989) called this the *Eumys elegans* zone, but as noted above, there are problems with the FAD of *Eumys*. Instead, we call the late early Orellan the *Miniochoerus affinis* Interval Zone, and designate its type section as the strata between 17 and 27 m (50–80 ft) above the PWL in the Boner Ranch section (SWsec.9,T.35N., R.61W., South Oat Creek 7.5′ Quadrangle) in the Seaman Hills, Niobrara County, Wyoming. The top of this zone is marked by the FAD of the zonal indicator of the overlying zone, *Miniochoerus gracilis*. This zone can also be recognized in the strata 17 to 25 m (50–75 ft) above the 5 tuff in Douglas, Wyoming, and correlates with some portion of Orella B in Toadstool Park. Most of the "early" Orellan fossils from the lower nodular zone of the Big Badlands are correlative with this zone. It is also correlative with most of Chron C13n (33.0–33.5 Ma).

Early late Orellan

The FAD of the dwarfed oreodont *Miniochoerus gracilis* and the horse *Mesohippus barbouri*, and the last appearance of *Ischyromys parvidens* occur 27 m (80 ft) above the PWL in the Lusk sections, ~57 m (170 ft) above the 5 tuff in the Douglas sections, at the base of Orella C in Toadstool Park, Nebraska, and at the top of the lower nodular zone in the Big Badlands. These taxa can be used to recognize an interval that might be called early late Orellan, or what we designate as the *Miniochoerus gracilis* Interval Zone. According to Korth (1989), there are several rodent taxa (*Agnotocastor readingi, Paradjidaumo validus, Eutypomys thomsoni,* and *Eumys parvidens*) restricted to Orella C in the Toadstool Park area; if their ranges are found to be consistent in the rest of the White River Group, they could also serve as indicators of this zone. The top of this zone is marked by the FAD of the zonal indicator of the overlying zone, *Merycoidodon bullatus*. We designate its type section as the strata 27 to 50 m (80–150 ft) above the PWL in the Boner Ranch section (SW sec. 9, T. 35 N., R. 61 W., South Oat Creek 7.5 quadrangle) in the Seaman Hills, Niobrara County, Wyoming. This zone begins in latest Chron C13n (33.1 Ma) and continues through the early part of Chron C12r (to 32.5 Ma).

Latest Orellan

Several distinctive taxa mark the latest Orellan, including the first appearances of the oreodont with large bullae, *Merycoidodon bullatus*, the last of the miniochoeres, *Miniochoerus starkensis*, the rabbit *Palaeolagus burkei*, the rodents (*fide* Korth, 1989) *Prosciurus magnus, Ecclesimus tenuiceps, Tenudomys basilaris, Pelycomys placidus, Heliscomys vetus, Heliscomys mcgrewi, Wilsoneumys planidens, Campestrallomys annectens*, and the last appearance of a great number of typically Orellan taxa, including *Hyaenodon crucians, Ischyromys typus, Paratylopus labiatus, Archaeotherium mortoni, Thinohyus lentus, Stibarus quadricuspis, Leptochoerus emilyae, Subhyracodon occidentalis, "Hesperocyon" coloradensis, Prosciurus, Pelycomys, Protosciurus, Oligospermophilus, Eutypomys, Adjidaumo, Heliscomys, Wilsoneumys, Eoeumys, Tenudomys, Pipestoneomys, Megalagus, Palaeolagus intermedius, Centetodon marginalis, Leptictis haydeni, Herpetotherium fugax, Copedelphys stevensoni, Nanodelphys hunti,* and all surviving species of *Mesohippus* (*M. bairdi, M. exoletus, M. westoni*, and *M. barbouri*). However, the ranges of many of the micromammals in this list may eventually be extended upward once adequate screenwashing is done in the Whitneyan.

This interval was called the "*Diplolophus insolens* zone" by Korth (1989), but as discussed above, this rodent is too rare to serve as an index fossil. We redesignate it as the *Merycoidodon bullatus* Interval Zone, with a type section in the "upper nodular zone" strata (termed "middle *Oreodon* beds" by some workers) on the east side of Sheep Mountain Table, just east of where the road climbs to the top (SW,NE,NW sec. 9, T. 43 N., R. 44 W., Sheep Mountain Table 7.5′ Quadrangle, Shannon County, South Dakota). We choose the Badlands sections over those in Nebraska or Colorado to typify the late Orellan and Whitneyan because the Frick collections from the Sheep Mountain Table–Cottonwood Pass–Corral Draw area are much larger and better documented than any other for this stratigraphic interval. Correlative strata include the upper part of Orella C and Orella D in the Toadstool Park area, and the upper part of the Cedar Creek Member in Colorado (Galbreath, 1953). These strata correlate with early Chron C12r (32.0–32.5 Ma). See Figure 8 for biogstratigraphic data within the Orellan and Whitneyan.

Early Whitneyan

The resolution of the biostratigraphic data for the Whitneyan collections is not as good as that for the Orellan. Most of the fossils come from the Big Badlands of South Dakota, where they are typically recorded as coming from Upper Oreodon, *Protoceras* beds, and *Leptauchenia* beds of Osborn (1907), Osborn and Matthew (1909), Wanless (1923), Skinner (unpublished field notes), and other workers; seldom is the exact stratigraphic level within these three units (each ~30m, or 100 ft thick) recorded, even by the Frick Laboratory. The other large Whitneyan collections come from the Toadstool Park–Roundtop area (where they are zoned as Whitney A, B, and C of Schultz and Stout, 1955). Hence the Whitneyan cannot be as finely subdivided as the Orellan with the present database.

Nevertheless, the early Whitneyan can be recognized by a number of species that first occur in Whitney A or in the "upper Oreodon beds." Traditionally, the Whitneyan was indicated by the first abundant occurrence of *Leptauchenia decora*; this is by far the most common fossil in these beds. In addition, the early Whitneyan is marked by the first occurrences of *Leptauchenia* (formerly "*Cyclopidius*") *major, Hyracodon leidyanus, Paratylopus primaevus, Paralabis cedrensis, Diceratherium tridactylum, Protapirus obliquidens, Ectopocynus antiquus, Oxetocyon cuspidatus, Cynodesmus thooides, Agnotocastor praetereadens,* and *Oropyctis pediasius*. A number of taxa, including *Miniochoerus starkensis, Stibarus obtusilobus, Hyaenodon horridus, Cedromus wilsoni, Metadjidaumo hendryi, Agnotocastor praetereadens,* and *Oropyctis pediasius* last occur in the early Whitneyan.

Of this list, we feel that the best index fossil is *Leptauchenia* (formerly "*Cyclopidius*") *major*, so we designate this interval as the *Leptauchenia major* Interval Zone. Its type section is the upper Oreodon beds (0–30m, or 0-90 ft above the upper nodular layer, or the Scenic-Poleslide contact) on the south side of Sheep Mountain Table, SE,NE,SE sec. 4, T. 42 N., R. 44 W., Sheep Mountain Table 7.5′ Quadrangle, Shannon County, South Dakota. It correlates with mid-Chron C12r (32.0–31.4 Ma).

Late Whitneyan

A number of taxa first occur in the later part of the Whitneyan ("*Protoceras* beds" and "*Leptauchenia* beds" in the Big Badlands,

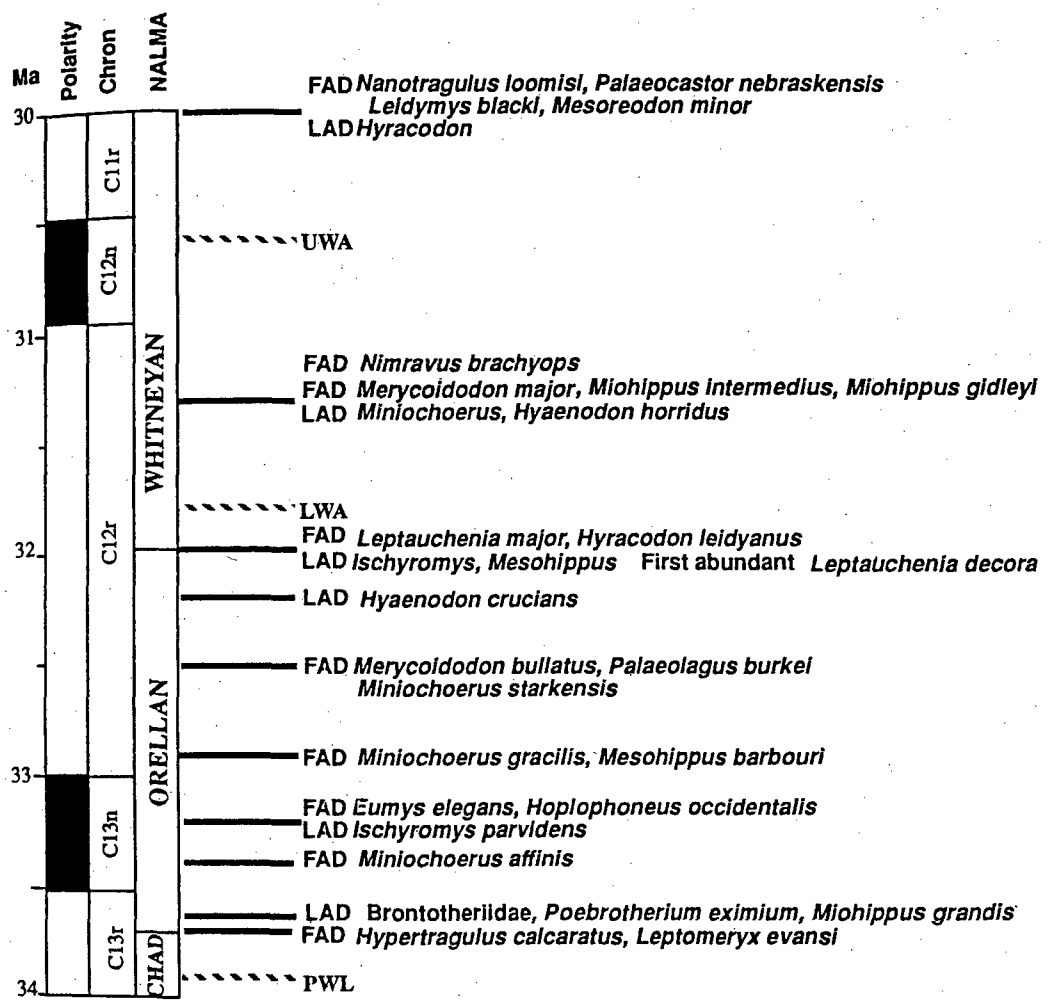


Figure 8. Important biostratigraphic datum levels within the Orellan and Whitneyan. Abbreviations: FAD = first appearance datum; LAD = last appearance datum; PWL = Persistent White Layer; LWA = Lower Whitney Ash; UWA = Upper Whitney Ash; NALMA = North American land-mammal "age." Time scale after Berggren et al. (1995) and Prothero and Swisher (1992).

and Whitney B–C of Toadstool Park). They include the large oreodont *Merycoidodon major*, the protoceratid *Protoceras celer*, the camel *Pseudolabis dakotensis*, the horses *Miohippus intermedius, M. annectens, M. equinanus*, and *M. gidleyi*, the nimravids *Hoplophoneus dakotensis, Eusmilus cerebralis*, and *Nimravus brachyops*, the creodont *Hyaenodon brevirostrus*, and the rodents *Eumys brachyodus* and *Scottimus lophatus*. The rabbit *Palaeolagus burkei* and the oreodont *Miniochoerus starkensis* last occur at this level. Of these taxa, *Merycoidodon major* is the most abundant and easy to recognize, so we designate the late Whitneyan as the *M. major* Interval Zone. Its type section is the "*Protoceras*" and "*Leptauchenia* beds" (30 to 103 m, or 90–310 ft above the upper nodular layer, or from 30 m (90 ft) above the Scenic-Poleslide contact to the Rockyford Ash) on the south side of Sheep Mountain Table (the type section of the Poleslide Member illustrated by Harksen and Macdonald, 1969a), NE,SE,SW sec. 28, T. 43 N., R. 44 W., Sheep Mountain Table 7.5′ Quadrangle, Shannon County, South Dakota. This interval correlates with late Chron C12r to early C11r (31.4–30.0 Ma).

Whitneyan-Arikareean boundary

Tedford et al. (1996) have discussed the problem of the recognition of the Whitneyan-Arikareean boundary; they recom-

mend basing it on the FAD of *Nanotragulus loomisi, Palaeolagus hypsodus, Palaeocastor nebrascensis, Leidymys blacki,* and *Mesoreodon minor*. In addition to these, a number of taxa first occur in the earliest Arikareean, including *Palaeolagus philoi, Shunkehetanka geringensis, Sespia nitida, Diceratherium armatum, Diceratherium annectens, Sanctimus stuartae, Geringia mcgregoryi*, and *Plesiosminthus*. These taxa first occur low in the Sharps Formation in South Dakota (near the Rockyford Ash), and near the second Nonpareil Ash Zone (NPAZ) of the "brown siltstone" member of the Brule Formation in Nebraska. These strata correlate with the base of Chron C11n (30.0 Ma). In addition, *Leptomeryx, Merycoidodon, Paratylopus, Paralabis, Perchoerus, Heptacodon, Leptochoerus, Colodon, Protapirus, Hesperocyon, Osbornodon, Dinictis, Paradjidaumo, Eumys*, and *Scottimus* last occur in the late Whitneyan; none of these genera have yet been reported from the Arikareean as currently defined.

CONCLUSIONS

In the 150 yr since White River fossils were first discovered and described, an enormous volume of fossils and data has accumulated. Early in this century, Osborn and Matthew made the first attempts to erect a biostratigraphic zonation of the White River Group, but their lead was not followed by the Wood Committee. In the 90 yr since the work of Osborn and Matthew, White River biostratigraphy has not kept pace with the biostratigraphic studies of the Paleocene (Archibald et al., 1987) or Eocene (Gingerich, 1983; Krishtalka et al., 1987), although many of the biostratigraphic zones proposed in these chapters lack type sections and therefore do not conform to the 1983 North American Stratigraphic Code (see Prothero, 1995). Now the systematic data base has been updated and detailed biostratigraphic, magnetostratigraphic, and radioisotopic data for most of the key sections are available. White River biostratigraphy can now conform to the methods and principles used by invertebrate biostratigraphers since the time of Oppel more than a century ago.

A biostratigraphic zonation for the Chadronian is still in preparation by R. Emry (personal communication), although its outlines were sketched by Prothero and Emry (1996b). In this chapter, we propose the following biostratigraphic zones for the Orellan and Whitneyan:

The Chadronian-Orellan boundary is defined on the first appearance of *Hypertragulus calcaratus*, and is characterized by the first occurrences of *Leptomeryx evansi, Palaeolagus intermedius*, and small *Miniochoerus chadronensis*. Brontotheres are no longer part of the boundary definition. Based on $^{40}Ar/^{39}Ar$ dates and magnetostratigraphy, this boundary is dated at 33.7 Ma (late Chron C13r) following the time scale of Berggren et al. (1995), although it may be slightly younger if the dates of Obradovich et al. (1995) are correct.

The earliest Orellan *Hypertragulus calcaratus* Interval Zone occurs in late Chron C13r (33.7–33.5 Ma).

The late early Orellan *Miniochoerus affinis* Interval Zone spans most of Chron C13n (33.5–33.1 Ma).

The early late Orellan *Miniochoerus gracilis* Interval Zone occurs in latest Chron C13n and early Chron C12r (33.1–32.5 Ma).

The latest Orellan *Merycoidodon bullatus* Interval Zone occurs in mid–Chron C12r (32.5–32.0 Ma).

The early Whitneyan *Leptauchenia major* Interval Zone spans late Chron C12r (32.0–31.4 Ma).

The late Whitneyan *Merycoidodon major* Interval Zone spans latest Chron C12r to late Chron C11n (31.4–30.0 Ma).

The Whitneyan-Arikareean boundary falls early in Chron C11n (30.0 Ma).

ACKNOWLEDGMENTS

We thank Malcolm McKenna and Richard Tedford for access to the Frick Collection of the American Museum of Natural History, Mike Voorhies, and Bob Hunt for access to the University of Nebraska State Museum collections, and Larry Martin for access to the University of Kansas Museum of Natural History collections. Much of the information in this chapter was part of D. R. P.'s doctoral dissertation, supervised by Malcolm McKenna, Dennis Kent, and Jim Hays.

We also thank Charles Denham, Dennis Kent, Neil Opdyke, William Roggenthen, and Joseph Kirschvink for access to their paleomagnetics laboratories. D. R. P. thanks Bob Emry, Emmett Evanoff, Phil Bjork, Margaret Stevens, Connie Gawne, Tim Heaton, Mike Voorhies, Malcolm McKenna, and the late Edwin Galbreath for their help with the stratigraphy and paleontology of the White River Group. K. E. W. thanks Rachel Benton, Scott Foss, and Valerie Naylor for their help with research in the Badlands. The field sampling would never have been possible without several field crews: in 1979, Karen Gonzalez and Heidi Shlosar; in 1980, Priscilla Duskin, Jon Frenzel, and Heidi Shlosar; in 1983, Allison Kozak, Rob Lander, and Annie Walton; in 1986, Allison Kozak, Dana Gilchrist, and Kecia Harris; in 1987, John Foster, Steve King, and Susan Briggs. Albert Whittlesey helped his daughter with the 1994 field sampling. Karen Gonzalez (1979), Priscilla Duskin (1980), Rob Lander (1983), Steve King (1987), and Ed Vance (1994) also helped with the processing of samples and laboratory analysis. We thank Bob Emry, Bob Hunt, Bill Korth, Hannan LaGarry, Bruce MacFadden, Margaret Stevens, John Storer, Dick Tedford, and Dennis Terry for reviews of the manuscript.

Prothero's research was supported by National Science Foundation grants EAR87-08221 and EAR91-17819, a grant from the Donors of the Petroleum Research Fund of the American Chemical Society, a grant in aid of research from Sigma Xi, and by field funds from the Department of Geological Sciences of Columbia University. Whittlesey was supported by the F. V. Hayden Paleontological Internship in Badlands National Park during this research.

This chapter is dedicated to the memory of Morris Skinner,

on whose work it was built. Morris deciphered most of the subtleties of the White River Group in the 1950s, but never published much of his immense store of knowledge. We who follow in his footsteps are just now putting into print the basic framework that Morris Skinner figured out 40 years ago.

REFERENCES CITED

Archibald, J. D., Clemens, W. A., Gingerich, P. D., Krause, D. W., Lindsay, E. H., and Rose, K. D., 1987, First North American land mammal ages of the Cenozoic Era, *in* Woodburne, M. O., ed., Cenozoic mammals of North America, geochronology and biostratigraphy: Berkeley, University of California Press, p. 24–76.

Barbour, E. H., and Stout, T. M., 1937, The White River Oligocene rodent *Diplolophus*: Bulletin of the University of Nebraska State Museum, v. 2, no. 3, p. 29–36.

Baskin, J. A., and Tedford, R. H., 1996, Small arctoid and feliform carnivorans, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingom, Cambridge University Press, p. 486–497.

Berggren, W. A., Kent, D. V., Aubry, M.-P., Swisher, C. C., III, and Miller, K. G., 1995, A revised Paleogene geochronology and chronostratigraphy: SEPM Special Publication 54, p. 129–212.

Bryant, H. R., 1996, Nimravidae, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 453–475.

Butler, R. F., 1992, Paleomagnetism: Boston, Blackwell, 319 p.

Clark, J., 1937, The stratigraphy and paleontology of the Chadron Formation in the Big Badlands of South Dakota: Carnegie Museum Annals, v. 25, p. 261–350.

Clark, J., Beerbower, J. R., and Kietzke, K. K., 1967, Oligocene sedimentation, stratigraphy, paleoecology and paleoclimatology of the Big Badlands of South Dakota: Fieldiana, Geology Memoir 5, p. 1–158.

CoBabe, E., 1996, Leptaucheniinae, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 574–580.

Darton, N. H., 1899, Preliminary report on the geology and water resources of Nebraska west of the 103rd meridian: U.S. Geological Survey Annual Report 19, part 4, p. 719–785.

Denham, C. R., 1984, Statistical sedimentation and magnetic polarity stratigraphy, *in* Berggren, W. A., and Van Couvering, J. A., eds., Catastrophes and earth history, the new uniformitarianism: Princeton, New Jersey, Princeton University Press, p. 101–112.

Emry, R. J., Bjork, P. R., and Russell, L. S., 1987, The Chadronian, Orellan, and Whitneyan land mammal ages, *in* Woodburne, M. O., ed., Cenozoic mammals of North America, geochronology and biostratigraphy: Berkeley, University of California Press, p. 118–152.

Evanoff, E., Prothero, D. R., and Lander, R. H., 1992, Eocene-Oligocene climatic change in North America: the White River Formation near Douglas, east-central Wyoming, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 116–130.

Ferrusquia-Villafranca, I., 1969, Rancho Gaitan local fauna, early Chadronian, northeastern Chihuahua: Boletin de Sociedad Geologica Mexicana, v. 30, p. 99–138.

Fisher, R. A., 1953, Dispersion on a sphere: Proceedings of the Royal Society, v. A217, p. 295–305.

Galbreath, E. C., 1953, A contribution to the Tertiary geology and paleontology of northeastern Colorado: University of Kansas Paleontological Contributions, Vertebrata, v. 4, p. 1–119.

Gingerich, P. D., 1983, Paleocene-Eocene faunal zones and a preliminary analysis of Laramide structural deformation in the Clarks Fork Basin, Wyoming: Wyoming Geological Association Guidebook, v. 34, p. 185–195.

Gustafson, E. P., 1986, Preliminary biostratigraphy of the White River Group (Oligocene, Chadron and Brule Formations) in the vicinity of Chadron, Nebraska: Transactions of the Nebraska Academy of Sciences, v. 14, p. 7–19.

Haake, M. E., and Galbreath, E. C., 1979, Dimensions of molar teeth of *Hypisodus minimus* (Artiodactyla) from the Oligocene deposits of northeastern Colorado: Transactions of the Illinois Academy of Sciences, v. 72, p. 30–36.

Harksen, J. C., and Macdonald, J. R., 1969a, Type sections for the Chadron and Brule Formations fo the White River Oligocene in the Big Badlands, South Dakota: South Dakota Geological Survey Report of Investigations, v. 99, p. 1–23.

Harksen, J. C., and Macdonald, J. R., 1969b, Guidebook to the major Cenozoic deposits of southwestern South Dakota: South Dakota Geological Survey Guidebook, v. 2, p. 1–103.

Hartl, P., Tauxe, L., and Constable, C., 1993, Early Oligocene geomagnetic field behavior from Deep Sea Drilling Project site 522: Journal of Geophysical Research, v. 98, n. B11, p. 19649–19665.

Hatcher, J. B., 1893, The *Titanotherium* beds: American Naturalist, v. 27, p. 204–221.

Heaton, T. H., 1993, The Oligocene rodent *Ischyromys* of the Great Plains: replacement mistaken for anagenesis: Journal of Paleontology, v. 67, p. 297–308.

Heaton, T. H., 1996, Ischyromyidae, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 373–398.

Heaton, T. H., and Emry, R. J., 1996, Leptomerycidae, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 581–608.

Howe, J. A., 1966, The Oligocene rodent *Ischyromys* in Nebraska: Journal of Paleontology, v. 40, p. 1200–1210.

Hunt, R. M., Jr., 1996, Amphicyonidae, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 476–485.

Johnson, H. P., Lowrie, W., and Kent, D. V., 1975, Stability of anhysteretic remanent magnetization in fine and coarse magnetite and maghemite particles: Geophysical Journal of the Royal Astronomical Society, v. 41, p. 1–10.

Korth, W. W., 1981, New Oligocene rodents from western North America: Carnegie Museum Annals, v. 50, p. 289–318.

Korth, W. W., 1989, Stratigraphic occurrence of rodents and lagomorphs in the Orella Member, Brule Formation (Oligocene), northwestern Nebraska: Contributions to Geology, University of Wyoming, v. 27, n. 1, p. 15–20.

Korth, W. W., and Hageman, J., 1988, Lagomorphs (Mammalia) from the Oligocene (Orellan and Whitneyan) Brule Formation, Nebraska: Transactions of the Nebraska Academy of Sciences, v. 16, p. 141–152.

Krishtalka, L., and 10 others, 1987, Eocene (Wasatchian through Duchesnean) biochronology of North America, *in* Woodburne, M. O., ed., Cenozoic mammals of North America, geochronology and biostratigraphy: Berkeley, University of California Press, p. 77–117.

Lander, E. B., 1977, A review of the Oreodonta (Mammalia, Artiodactyla) [Ph.D. thesis]: Berkeley, University of California, 502 p.

Lillegraven, J. A., 1970, Stratigraphy, structure, and vertebrate fossils of the Oligocene Brule Formation, Slim Buttes, northwestern South Dakota: Geological Society of America Bulletin, v. 81, p. 831–850.

Lillegraven, J. A., McKenna, M. C., and Krishtalka, L., 1981, Evolutionary relationships of middle Eocene and younger species of *Centetodon* (Mammalia, Insectivora, Geolabididae) with a description of the dentition of *Ankylodon* (Adapisoricidae): University of Wyoming Publications, v. 41, p. 1–115.

Mader, B., 1989, The Brontotheriidae: a systematic revision and preliminary phylogeny of North American genera, *in* Prothero, D. R., and Schoch, R. M., eds., The evolution of perissodactyls: New York, Oxford University Press, p. 458–484.

Martin, L. D., 1980, The early evolution of the Cricetidae in North America: University of Kansas Paleontological Contributions, v. 102, p. 1–42.

Matthew, W. D., 1899, A provisional classification of the fresh-water Tertiary of the West: Bulletin of the American Museum of Natural History, v. 12, F

Mellett, J. S., 1977, Paleobiology of North American *Hyaenodon* (Mammalia, Creodonta): Contributions to Vertebrate Evolution, v. 1, p. 1–134.

Murphy, M. A., 1977, On chronostratigraphic units: Journal of Paleontology, v. 51, p. 123–219.

NACSN (North American Commission on Stratigraphic Nomenclature), 1983, North American Stratigraphic Code: American Association of Petroleum Geologists Bulletin, v. 67, no. 5, p. 841–875.

Obradovich, J. D., Evanoff, E., and Larson, E. E., 1995, Revised single-crystal laser-fusion $^{40}Ar/^{39}Ar$ ages of Chadronian tuffs in the White River Formation of Wyoming: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. 77–78.

Opdyke, N. D., Lindsay, E. H., Johnson, N. M., and Downs, T., 1977, The paleomagnetism and magnetic polarity stratigraphy of the mammal-bearing section of Anza-Borrego State Park, California: Journal of Quaternary Research, v. 7, p. 316–329.

Osborn, H. F., 1907, Tertiary mammal horizons of North America: Bulletin of the American Museum of Natural History, v. 23, p. 237–254.

Osborn, H. F., 1910, The age of mammals in Europe, Asia, and North America: New York, MacMillan and Co., 635 p.

Osborn, H. F., 1929, The titanotheres of ancient Wyoming, Dakota, and Nebraska: U.S. Geological Survey Monograph, v. 55, p. 1–701.

Osborn, H. F., and Matthew, W. D., 1909, Cenozoic mammal horizons of western North America: U.S. Geological Survey Bulletin, v. 361, p. 1–138.

Ostrander, G. E., 1985, Correlation of the early Oligocene (Chadronian) in northwestern Nebraska, *in* Martin, J. E., ed., Fossiliferous Cenozoic deposits of western South Dakota and northwestern Nebraska: Dakoterra, v. 2, p. 205–231.

Pluhar, C. J., Kirschvink, J. L., and Adams, R. W., 1991, Magnetostratigraphy and clockwise rotation of the Plio-Pleistocene Mojave River Formation, central Mojave Desert, California: San Bernardino County Museum Association Quarterly, v. 38, no. 2, p. 31–42.

Prothero, D. R., 1982, Medial Oligocene magnetostratigraphy and mammalian biostratigraphy: testing the isochroneity of mammalian biostratigraphic events [Ph.D. thesis]: New York, Columbia University, 284 p.

Prothero, D. R., 1985a, Chadronian (early Oligocene) magnetostratigraphy of eastern Wyoming: implications for the Eocene-Oligocene boundary: Journal of Geology, v. 93, p. 555–565.

Prothero, D. R., 1985b, Correlation of the White River group by magnetostratigraphy, *in* Martin, J. E., ed., Fossiliferous Cenozoic deposits of western South Dakota and northwestern Nebraska: Dakoterra, v. 2, p. 265–276.

Prothero, D. R., 1994, The Eocene-Oligocene transition: paradise lost: New York, Columbia University Press, 291 p.

Prothero, D. R., 1995, Geochronology and magnetostratigraphy of Paleogene North American land mammal "ages": an update: SEPM Special Publication 54, p. 305–315.

Prothero, D. R., 1996a, Magnetostratigraphy of the White River group in the High Plains, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 262–277.

Prothero, D. R., 1996b, Camelidae, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 609–651.

Prothero, D. R., 1996c, Hyracodontidae, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 652–663.

Prothero, D. R., and Emry, R. J., 1996a, The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, 690 p.

Prothero, D. R., and Emry, R. J., 1996b, Summary, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 664–682.

Prothero, D. R., and Heaton, T. H., 1996, Faunal stability during the early Oligocene climatic crash: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 127, p. 257–284.

Prothero, D. R., and Shubin, N., 1989, The evolution of Oligocene horses, *in* Prothero, D. R., and Schoch, R. M., eds., The evolution of perissodactyls: New York, Oxford University Press, p. 142–175.

Prothero, D. R., and Swisher, C. C., III, 1992, Magnetostratigraphy and geochronology of the terrestrial Eocene-Oligocene transition in North America, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 46–74.

Prothero, D. R., Denham, C. R., and Farmer, H. G., 1982, Oligocene calibration of the magnetic polarity timescale: Geology, v. 10, p. 650–653.

Prothero, D. R., Denham, C. R., and Farmer, H. G., 1983, Magnetostratigraphy of the White River Group and its implications for Oligocene geochronology: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 42, p. 151–166.

Prothero, D. R., Guérin, C., and Manning, E., 1989, The history of the Rhinocerotoidea, *in* Prothero, D. R., and Schoch, R. M., eds., The evolution of perissodactyls: New York, Oxford University Press, p. 322–340.

Retallack, G., 1983, Late Eocene and Oligocene fossil paleosols from Badlands National Park, South Dakota: Geological Society of America Special Paper 193, 82 p.

Savage, D. E., 1955, Nonmarine lower Pliocene sediments in California, geochronologic-stratigraphic classification: University of California Publications in Geological Sciences, v. 31, no. 1, p. 1–26.

Savage, D. E., 1962, Cenozoic geochronology of the fossil mammals of the Western Hemisphere: Revista Museo Argentino Ciencias Naturales, v. 8, p. 53–67.

Savage, D. E., 1977, Aspects of vertebrate paleontological stratigraphy and geochronology, *in* Kaufman, E. G., and Hazel, J. E., eds., Concepts and methods in biostratigraphy: Stroudsburg, Pennsylvania, Dowden, Hutchinson, and Ross, p. 427–442.

Savage, D. E., and Russell, D. E., 1983, Mammalian paleofaunas of the world: Reading, Massachusetts, Addison-Wesley, 432 p.

Schlaikjer, E. M., 1935, Contributions to the stratigraphy and paleontology of the Goshen Hole area, Wyoming. IV. New vertebrates and the stratigraphy of the Oligocene and early Miocene: Bulletin of the Museum of Comparative Zoology, Harvard, v. 76, p. 97–189.

Schultz, C. B., and Falkenbach, C. H., 1956, Miniochoerinae and Oreonetinae, two new subfamilies of oreodonts: Bulletin of the American Museum of Natural History, v. 109, p. 377–482.

Schultz, C. B., and Falkenbach, C. H., 1968, The phylogeny of the oreodonts, Parts 1 and 2: Bulletin of the American Museum of Natural History, v. 139, p. 1–498.

Schultz, C. B., and Stout, T. M., 1938, Preliminary remarks on the Oligocene of Nebraska [abs.]: Bulletin of the Geological Society of America, v. 49, p. 1921.

Schultz, C. B., and Stout, T. M., 1955, Classification of the Oligocene sediments in Nebraska: Bulletin of the University of Nebraska State Museum, v. 4, p. 17–52.

Schultz, C. B., and Stout, T. M., 1961, Field conference on the Tertiary and Pleistocene of western Nebraska: Special Publications of the University of Nebraska State Museum, v. 2, p. 1-54.

Scott, W. B., Jepsen, G. L., and Wood, A. E., eds., 1937–1941, The mammalian fauna of the White River Oligocene, parts I–V: Transactions of the American Philosophical Society, v. 28, p. 1–980.

Sinclair, W. J., 1921, The "Turtle-Oreodon" layer or "Red Layer," a contribution to the stratigraphy of the White River Oligocene: Proceedings of the American Philosophical Society, v. 60, p. 457–466.

Singler, C. R., and Picard, M. D., 1980, Stratigraphic review of the Oligocene beds in the northern Great Plains: Wyoming Geological Association and Earth Science Bulletin, v. 13, no. 1, p. 1–18.

Stevens, M. S., and Stevens, J. B., 1996, Merycoidodontinae and Miniochoeri-

nae, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 498–573.

Storer, J. E., 1994, A latest Chadronian (late Eocene) mammalian fauna from the Cypress Hills, Saskatchewan: Canadian Journal of Earth Science, v. 31, p. 1335–1341.

Storer, J. E., and Bryant, H. R., 1993, Biostratigraphy of the Cypress Hills Formation (Eocene to Miocene), Saskatchewan: equid types (Mammalia, Perissodactyla) and associated faunal assemblages: Journal of Paleontology, v. 67, p. 660–669.

Swisher, C. C., III, and Prothero, D. R., 1990, Single-crystal $^{40}Ar/^{39}Ar$ dating of the Eocene-Oligocene transition in North America: Science, v. 249, p. 760–762.

Tabrum, A. R., Prothero, D. R., and Garcia, D., 1996, The Eocene-Oligocene transition in southwestern Montana, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 278–311.

Tedford, R. H., 1970, Principles and practices of mammalian geochronology in North America: Proceedings of the North American Paleontological Convention, v. 2F, p. 666–703.

Tedford, R. H., Swinehart, J., Prothero, D. R., Swisher, C. C., III, King, S. A., and Tierney, T. E., 1996, The Whitneyan-Arikareean transition in the High Plains, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 312–334.

Thorpe, M. R., 1937, The Merycoidodontidae, an extinct group of ruminant mammals: Memoirs of the Peabody Museum of Natural History, v. 3, no. 4, p. 1–428.

Wang, X., 1994, Phylogenetic systematics of the Hesperocyoninae (Carnivora: Canidae): Bulletin of the American Museum of Natural History, v. 221, p. 1–207.

Wang, X., and Tedford, R. H., 1996, Canidae, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, United Kingdom, Cambridge University Press, p. 433–452.

Wanless, H. R., 1923, The stratigraphy of the White River beds of South Dakota: Proceedings of the American Philosophical Society, v. 62, p. 190–269.

Wells, W. B., Terry, D. O., and LaGarry, H. E., 1995, Stratigraphic implications of a fluvial origin for the "nodular zones," Brule Formation (Orella Member), northwestern Nebraska: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. 95.

Williams, H. S., 1901, The discrimination of time-values in geology: Journal of Geology, v. 9, p. 570–585.

Wood, A. E., 1980, The Oligocene rodents of North America: Transactions of the American Philosophical Society, v. 70, no. 5, p. 1–68.

Wood, H. E., Chaney, R. W., Clark, J., Colbert, E. H., Jepsen, G. L., Reeside, J. B. Jr., and Stock, C., 1941, Nomenclature and correlation of the North American continental Tertiary: Geological Society of America Bulletin, v. 52, p. 1–48.

Woodburne, M. O., 1977, Definition and characterization in mammalian chronostratigraphy: Journal of Paleontology, v. 51, p. 220–234.

Woodburne, M. O., 1987, Principles, classification, and recommendations, *in* Woodburne, M. O., ed., Cenozoic mammals of North America, geochronology and biostratigraphy: Berkeley, University of California Press, p. 9–17.

Wortman, J., 1893, On the divisions of the White River or lower Miocene of Dakota: Bulletin of the American Museum of Natural History, v. 5, p. 95–105.

MANUSCRIPT ACCEPTED BY THE SOCIETY DECEMBER 5, 1997

Printed in U.S.A.

Geological Society of America
Special Paper 325
1998

Lithostratigraphic revision and redescription of the Brule Formation (White River Group) of northwestern Nebraska

Hannan E. LaGarry
Conservation and Survey Division (IANR) and State Museum, University of Nebraska–Lincoln, Lincoln, Nebraska 68588-0514

ABSTRACT

Based on geologic mapping, measured sections, and lithologic correlations, the boundaries of the Brule Formation are revised in the type area in northwestern Nebraska. Marker beds used to subdivide and correlate the Brule Formation are recognized as local features of the type area, and their use as stratigraphic boundaries is abandoned in favor of boundaries corresponding to mapped changes in lithology. Using revised lithostratigraphic boundaries, the Brule Formation and its members are redescribed as a series of lithotopes. The revised upper boundary of the White River Group is the unconformable lithologic contact between massive volcaniclastic siltstones and sandy siltstones of the Brule Formation and brown and grayish-brown fluvial volcaniclastic sandstones, cross-bedded conglomeratic sandstones, and massive tabular volcaniclastic sandstones of the overlying Arikaree Group. These massive siltstones and sandy siltstones were correlated to the Gering Formation of the Arikaree Group, but are now recognized as the brown siltstone member of the Brule Formation. They are redescribed as a lithotope consisting of pale brown and brown indistinctly bedded, nodular, and massive or weakly cross-bedded eolian volcaniclastic siltstones, eolian volcaniclastic sandy siltstones, minor amounts of fluvially reworked eolian volcaniclastic siltstone, and volcanic ash (the Nonpareil ash zone). The revised boundary between the brown siltstone and the underlying Whitney Members is the gradational or sharp and undulating contact between the brown siltstone and Whitney lithotopes. This contact is an erosional unconformity, where the brown siltstone member fills valleys and depressions in the Whitney Member. The Whitney Member is redescribed as two lithotopes. The first consists of pale brown, massive, and typically nodular eolian siltstones with occasional thin interbeds of brown and bluish-green sandstone, and volcanic ash (the upper and lower Whitney ashes). The second consists of white or green laminated fluvial siltstones, sheet sandstones, and channel sandstones restricted to the lowest 10 m of the Whitney Member. The revised boundary between the Whitney and underlying Orella Members is moved downward 2.5 to 5 m from the "white bed" to the lithologic contact between the Whitney and Orella lithotopes. This contact is intertonguing except where siltstones and sandstones of the Whitney Member incise the Orella Member. The Orella Member is redescribed as two lithotopes. The first consists of thinly interbedded and slightly pedogenically modified pale brown, brown, and brownish-orange volcaniclastic overbank clayey siltstones and silty claystones, brown and bluish-green overbank sheet sandstones, and volcanic ash. The second consists of single and multistoried channel sandstones occurring throughout the Orella Member. The revised boundary between the Brule and underlying Chadron

LaGarry, H. E., 1998, Lithostratigraphic revision and redescription of the Brule Formation (White River Group) of northwestern Nebraska, *in* Terry, D. O., Jr., LaGarry, H. E., and Hunt, R. M., Jr., eds., Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America): Boulder, Colorado, Geological Society of America Special Paper 325.

Formations is recognized 9 to 10 m above its previous location, the "upper purplish-white layer." This contact is intertonguing except where channel sandstones of the Orella Member cut into the underlying Chadron Formation.

INTRODUCTION

Vertebrate fossil inventories of the Oglala National Grassland from 1991 to 1993 (see LaGarry and Hunt, 1994, 1995) produced preliminary geologic maps of the White River in the type area of the Chadron and Brule Formations near Toadstool Park, Sioux County, Nebraska (Fig. 1). This mapping indicated that the stratigraphic boundaries of the White River Group in northwestern Nebraska were based on marker beds that could not be traced beyond Toadstool Park, and, unless these boundaries were revised to correspond to lithologic changes rather than local marker beds, they could not be mapped or correlated with adjacent areas. Concurrently, independent work revising the basal strata of the White River Group in South Dakota began when Evans and Terry (1994) defined the Chamberlain Pass Formation.

Geologic mapping of northwestern Nebraska continued from 1994 to 1995 within the University of Nebraska Conservation and Survey Division state map project (LaGarry and LaGarry, 1997). This work began the revision of the White River Group to meet U.S. Geological Survey (USGS) standards for the mappability of units and conformity to the North American Stratigraphic Code (NACSN, 1983). This work led to further changes to the White River Group in northwestern Nebraska (Fig. 2). These changes include the recognition of the Chamberlain Pass Formation and the Peanut Peak Member of the Chadron Formation in northwestern Nebraska (Terry, this volume), the definition of the Big Cottonwood Creek Member of the Chadron Formation (Terry and LaGarry, this volume), and this revision of the Brule Formation.

This chapter revises and redescribes the lithostratigraphic boundaries and contents of the Brule Formation in northwestern Nebraska. Although the Brule Formation, or its lithologic and temporal equivalents, is also exposed in South Dakota, Colorado, Wyoming, and North Dakota, discussion of lithostratigraphic boundaries and lithologies in these areas is beyond the scope of this chapter. Biostratigraphic subdivisions of the Brule Formation are not addressed in detail here, but are summarized in Schultz

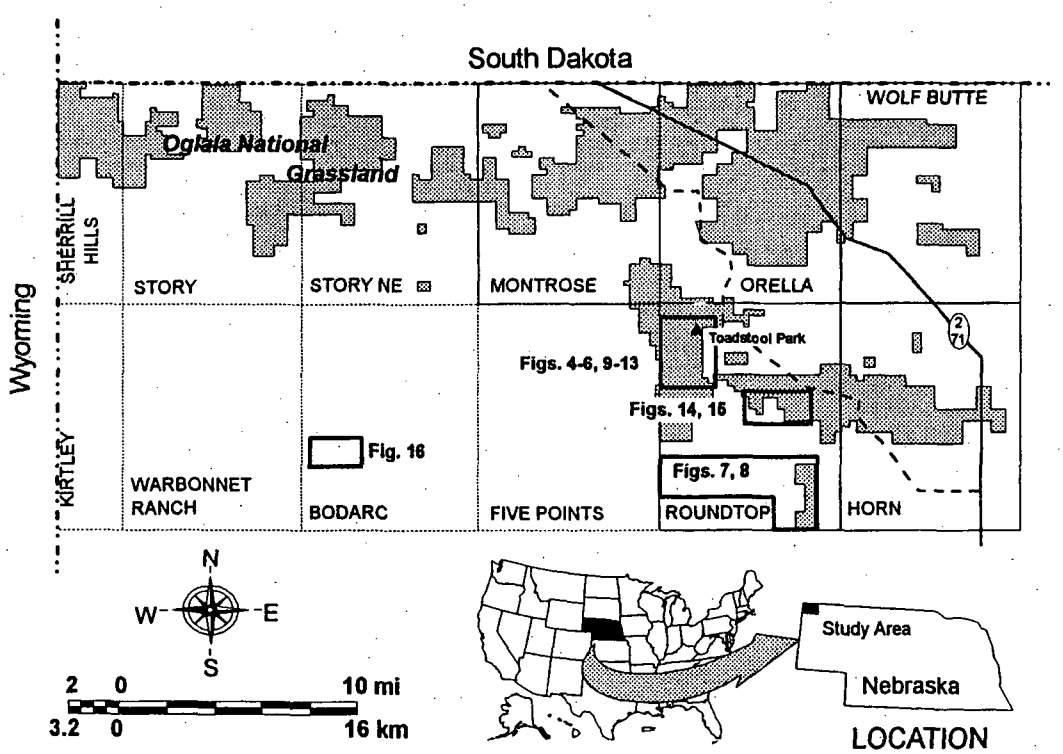


Figure 1. Locations of U.S. Geological Survey 7.5′ quadrangles, geologic maps, the Oglala National Grassland, and figured sections discussed in this chapter.

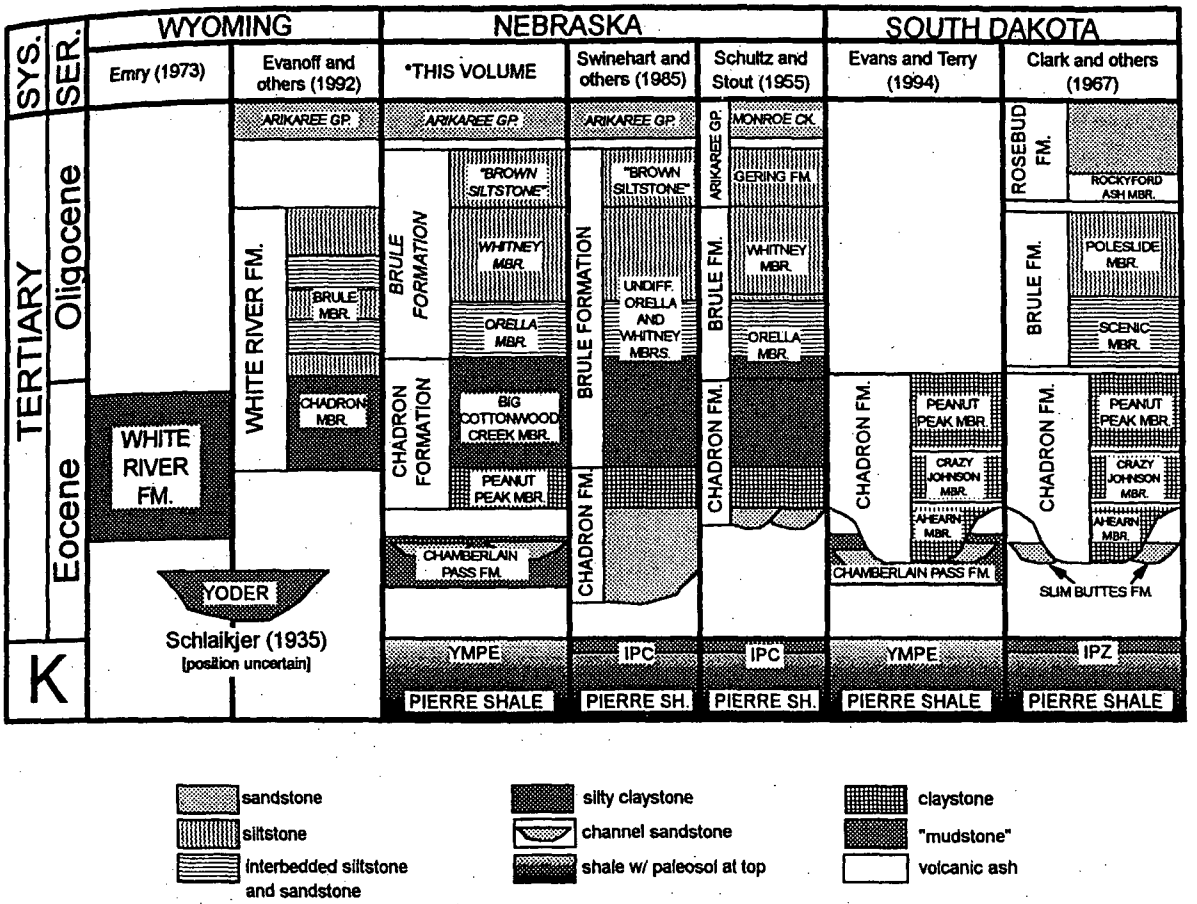


Figure 2. Proposed lithostratigraphic equivalency of the White River Group and adjacent lithostratigraphic units in Nebraska, South Dakota, and Wyoming. (*) includes data from Terry (this volume), Terry and LaGarry (this volume), and this chapter. Italicized units indicate those revised herein. YMPE = Yellow Mounds Paleosol equivalent (see Terry, this volume); IPC = Interior Paleosol Complex of Schultz and Stout (1955); IPZ = Interior Paleosol Zone of Clark et al. (1967).

and Falkenbach (1968), Emry et al. (1987), Tedford et al. (1996), and Prothero and Whittlesey (this volume).

BRULE FORMATION: HISTORICAL OVERVIEW

The type area for the White River Group is the Big Badlands of South Dakota (Meek and Hayden, 1857, 1861). However, Darton (1899) was the first to apply formation names to the White River Group when describing lithologies in northwestern Nebraska. In his "Roundtop to Adelia" measured section, Darton (1899, Fig. 226) divided the White River Group into the Chadron Formation and the Brule Clays, which were overlain by his Gering and Arikaree Formations. Since Darton's work, controversy has surrounded the type localities and sections of the White River Group in Nebraska and South Dakota. Darton (1899, p. 736) named the Chadron Formation and the Brule Clays "to afford distinct definitions for the members in Nebraska" without designating specific type sections. In the same discussion, Darton (p. 758) referred to "the typical section in the Big Bad Lands of South Dakota." Consequently, type sections in both Nebraska and South Dakota were designated based on these accounts (see Schultz and Stout, 1955; Clark et al., 1967). This has led to independent, competing, and often conflicting stratigraphic classifications in the two regions. Details of the type section controversies and stratigraphic confusion surrounding the White River Group in Nebraska and South Dakota prior to and following Darton's (1899) work have been summarized by Harksen and Macdonald (1969) and Singler and Picard (1980).

Schultz and Stout (1955) revised the White River Group in northwestern Nebraska and designated a "type" area. They con-

sidered Darton's Roundtop to Adelia measured section to be the "type" section of the Chadron and Brule Formations, and described additional sections at nearby Toadstool Park. However, Schultz and Stout differed from Darton in the lithologies that they assigned to each formation (Fig. 3); they combined Darton's Chadron Formation with the lower 52 m of Darton's Brule Clays (the basal "greenish sands and sandy clays") to form their Chadron Formation. Schultz and Stout (1955) divided their Chadron Formation into the "Chadron A," or "Yoder," "Chadron B," and "Chadron C" (see also Terry and LaGarry, this volume). Schultz and Stout's (1955) Brule Formation, which they divided into the Orella and Whitney Members, included the remainder of Darton's Brule Clays (Fig. 3). Schultz (1938) had previously divided Darton's Arikaree Formation into the Harrison and Monroe Creek Formations, and erected the Arikaree Group. Subsequently, Schultz and Stout's Brule Formation was overlain by the Gering Formation of the Arikaree Group (Figs. 2, 3).

Schultz and Stout's (1955) formal and informal White River Group boundaries were based on their model for Quaternary terrace deposition (summarized by Stout, 1978), and made the following assumptions (see also Schultz and Stout, 1945, 1948, 1955, 1977, 1980; Stout 1971, 1973, 1977): (1) sediments are deposited during climate-driven cycles, or "megacyclothems"; (2) cycles of sedimentation begin with basal gravels or conglomerates, fine upward, and terminate in event horizons or unconformities (e.g., their "terminal paleosols"); and (3) these event horizons and unconformities can be correlated across great distances. In their view, sedimentary deposits consist of a worldwide series of stacked, isochronous (*sensu* NACSN, 1983, p. 849) layers of sediment, with synchronous boundaries and erosion or nondeposition between depositional cycles. They further assumed that rock, time, and faunal units were isochronous and controlled by paleoclimatic cycles, and used lithologic, paleontologic, and paleopedologic data to rec-

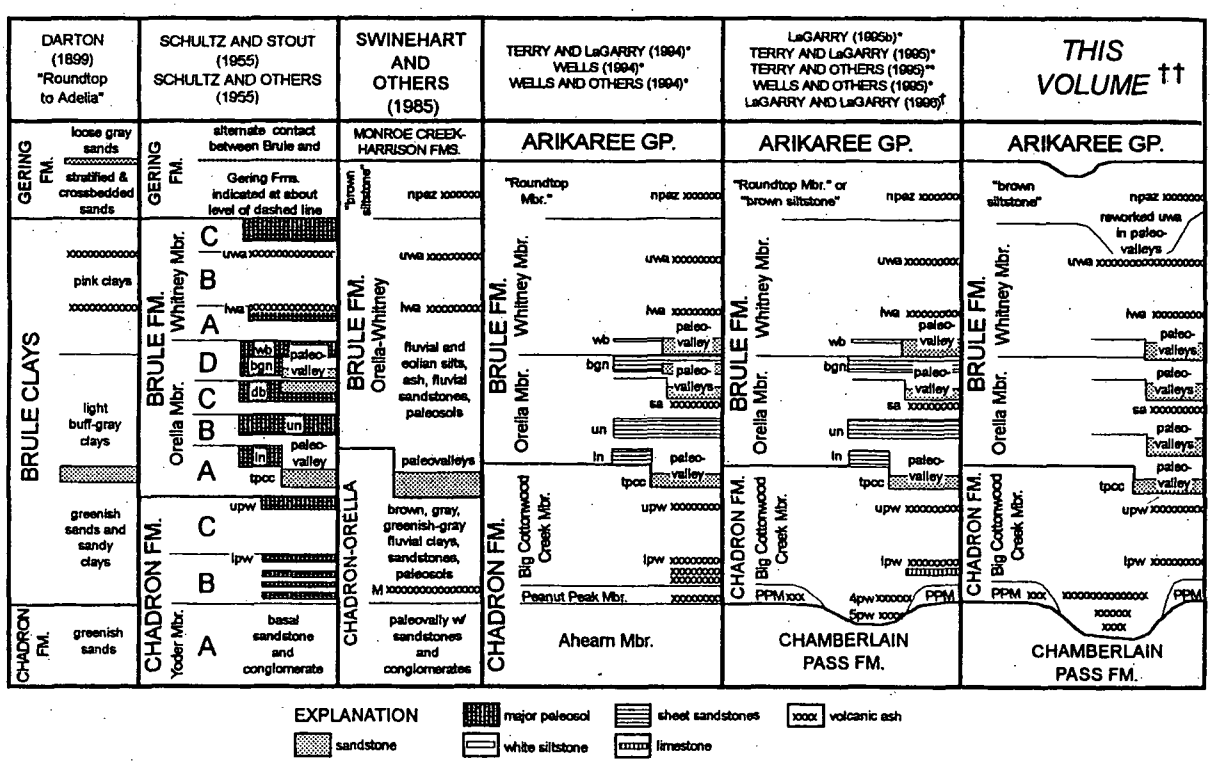


Figure 3. History of marker bed interpretation and terminology within the White River Group of northwestern Nebraska. Key: (*) refers to published abstract; (**) refers to field guide; (†) refers to an open-file report; (††) refers to data from Terry (this volume), Terry and LaGarry (this volume), and this chapter. PPM = Peanut Peak Member of the Chadron Formation of Terry et al. (1995) and Terry (this volume). Other abbreviations: uwa = upper Whitney ash; lwa = lower Whitney ash; wb = white bed; bgn = bluish-green nodules; db = *Diplolophus insolens* bench; un = upper nodules; ln = lower nodules; tpcc = Toadstool Park channel complex; upw = upper purplish-white layer; lpw = lower purplish-white layer; 4pw, 5pw = fourth purplish-white layer and fifth purplish white layer, both of Schultz and Stout (1955); npaz = Nonpareil ash zone of Swinehart et al. (1985); sa = serendipity ash of Swinehart and Evanoff (personal communication, 1993).

ognize event horizons marking the beginning and end of individual cycles and synchronous unit boundaries (Stout, 1978; Martin, 1994). Schultz and Stout (1955) recognized three depositional cycles beginning with sandstone or gravel and ending in terminal paleosols, corresponding to the Chadron Formation, and the Orella and Whitney Members of the Brule Formation, respectively, along with at least nine additional subcycle terminal paleosols, or their equivalents, within the White River Group in northwestern Nebraska (Fig. 3).

Concurrent with Schultz and Stout's (1955) classification of the White River Group in Nebraska, lithostratigraphic units within the White River Group were defined in the Big Badlands of South Dakota (Clark, 1937, 1954; Bump, 1956). These lithostratigraphic units are based on lithogenetic units (*sensu* Clark et al., 1967), rather than presumed climate-driven cyclicity (Fig. 2). Bump (1956) divided the Brule Formation in the Big Badlands into the Scenic and Poleslide Members, and designated type sections in Pennington and Shannon Counties, South Dakota (subsequently moved by Harksen and Macdonald, 1969). The Scenic Member consists of pedogenically modified tan, red, and ochre laminated volcaniclastic mudstones and siltstones interbedded with channel sandstones (Clark et al., 1967; Retallack, 1983). Nodules are common throughout the Scenic Member (Retallack, 1983). Two dense stratigraphic occurrences of these are the upper and lower nodular zones (Bump, 1956). Schultz and Stout (1955) correlated their "upper" and "lower nodules" (Fig. 3) to Bump's (1956) upper and lower nodular zones, and suggested that these sets of nodules represented temporally and lithologically equivalent event horizons. The Scenic Member intertongues with the overlying Poleslide Member, which consists primarily of pedogenically modified volcaniclastic mudstones and siltstones, and lenticular channel sandstones (Bump, 1956; Clark et al., 1967; Retallack, 1983). Schultz and Stout (1955) considered the Whitney and Poleslide Members to be temporal equivalents, but did not specify lithologic equivalents in the Big Badlands for marker beds within the Whitney Member. Until recently, the upper boundary of the Brule Formation, and the White River Group in the Big Badlands, was the base of the Rockyford Ash Member of the Sharps Formation (Nicknish and Macdonald, 1962). However, Tedford et al. (1996) recognized two lithotopes within the Sharps Formation. They assigned the lower of these, which consists primarily of nodular volcaniclastic eolian sandy siltstones, massive volcaniclastic eolian silty sandstones, and the basal Rockyford Ash Member, to the White River Group.

Based on the thickness and lateral extent of White River Group rocks in northwestern Nebraska, Singler and Picard (1980) elevated the Brule Formation to a subgroup and the Orella and Whitney Members to formations. However, these revisions were not adopted and Schultz and Stout's (1955) White River Group stratigraphy remained untested until Swinehart et al. (1985) compiled subsurface data from more than 12,000 drill holes across the Nebraska panhandle. Swinehart et al. (1985) recognized three depositional sequences within the White River Group in Nebraska, corresponding to the Chadron–lower Orella, upper Orella–Whitney, and the newly recognized brown siltstone (Figs. 2, 3). Based on regional unconformities recognized in the subsurface, Swinehart et al. restricted Darton's (1899) and Schultz and Stout's (1955) Gering Formation to a paleovalley in the southern panhandle of Nebraska. Based on outcrop data, Swinehart and Diffendal (1995) and Tedford et al. (1996) showed that beds previously considered to be within the basal Gering Formation in the southern and central panhandle of Nebraska belonged to the brown siltstone of Swinehart et al. (1985), and recognized a Gering Formation restricted to the southern Nebraska panhandle. The brown siltstone is now widely recognized in exposures along the Wildcat Ridge in the southern Nebraska panhandle (Swinehart and Diffendal, 1995, 1997; Tedford et al., 1996), but correlative exposures in northwestern Nebraska were still considered to be the Gering Formation by Meehan (1994).

LITHOGENETIC UNITS AND THE NORTH AMERICAN STRATIGRAPHIC CODE

The North American Stratigraphic Code (NACSN, 1983; Article 23e) states that lithostratigraphic and lithogenetic units should correspond such that lithogenetic units do not straddle formal lithostratigraphic boundaries. The code also specifies that lithostratigraphic units be defined by uniform lithologic content (NACSN, 1983; Article 22a). This approach promotes lithostratigraphic units that correspond to three-dimensional bodies of rock of uniform lithology and sedimentation. However, there is confusion and disagreement about the terminology and use of lithogenetic units (see Krumbein and Sloss, 1963). This confusion has resulted from efforts to develop terms for bodies of rock (e.g., lithotope, lithostrome, lithosome, lithofacies; see below) separate from the formal terms applied to lithostratigraphic units (e.g., group, formation, member). Although resolution of the problem is beyond the scope of this chapter, discussion of these terms will clarify their use here.

Most designations of *lithostratigraphic units* are based on vertical characterizations of lithology within two-dimensional representations of the rock record (e.g., stratigraphic columns). In contrast, "lithofacies" distinguish between lateral aspects, or variations in lithology, within adjacent, contemporaneous depositional environments (see Krumbein and Sloss, 1963; Bates and Jackson, 1984; p. 298, definition 1). Lithofacies are also typically characterized two-dimensionally (e.g., lithofacies maps). However, it is widely recognized that *lithogenetic units* represent particular depositional environments through time, and consist of volumes of rock. Wells (1947) used the term "lithotope" to describe three-dimensional "rock records of environment." Wells's (1947, p. 121) lithotopes are of generally uniform lithology, but include small lithologic variations if no significant change in depositional environment occurred. Subsequently, Krumbein and Sloss (1951) used lithotope and lithofacies interchangeably for rock records of depositional environments, while Teichert (1958) recognized lithotopes as

"isopic lithofacies." Wheeler and Mallory (1956) recognized the lithotope as a genetic unit of uniform lithology and sedimentation, but considered it to represent a two-dimensional subset of the lithofacies. They proposed the term "lithostrome" for uniform, vertical (two-dimensional) sequences of rock within a stratigraphic unit. Later, Krumbein and Sloss (1963) adopted Wheeler and Mallory's (1956) use of the terms lithotope and lithostrome, and suggested the term lithosome for three-dimensional bodies of rock. As currently used, the term lithosome represents bodies of rock deposited under uniform physiochemical conditions, and lithotope and lithofacies are synonyms (Bates and Jackson, 1984; p. 298 [definitions 1 and 2; p. 299]).

As stated above, the North American Stratigraphic Code (NACSN, 1983) advocates that lithostratigraphic units consist of three dimensional bodies of uniform lithology. The terms lithofacies and lithostrome denote horizontal and vertical lithologic variation in two dimensions, respectively, whereas the terms lithotope and lithosome clearly denote lithologic uniformity in three-dimensions. The term lithotope, as it was first applied by Wells (1947), best meets the provisions of the North American Stratigraphic Code (1983) while encompassing the minor variations in lithology typical within terrestrial sedimentary lithostratigraphic units. Not all sedimentary lithostratigraphic units are likely to have a consistent character in three dimensions, and lithotopes *sensu strictu* are not valid formal lithostratigraphic units (NACSN, 1983, Article 22l). However, formally defined lithostratigraphic units can be composed of one or more genetically related lithotopes.

REVISION AND REDESCRIPTION OF THE BRULE FORMATION

The following revision and redescription of the Brule Formation in northwestern Nebraska is based on lithologies (lithotopes *sensu* Wells, 1947) mapped on the Montrose, Orella, Wolf Butte, Roundtop, Horn, and Bodarc 7.5′ Quadrangles (Fig. 1) from 1991 through 1995 (see LaGarry and LaGarry, 1997). Descriptions of these lithotopes are based on measured sections in the Roundtop and Bodarc 7.5′ Quadrangles. Each of the following lithotope descriptions contains a review of Schultz and Stout's (1955) classification, a discussion of the revised lithostratigraphy, and a summary of the usefulness of the named marker beds. Preliminary accounts of these lithostratigraphic boundary revisions and lithologic redescriptions have appeared in abstracts and a field guide (Fig. 3), and although insufficient as formal treatments (NACSN, 1983), they demonstrate the evolution of the concepts applied here and are mentioned where appropriate. A description of the regional distribution of Brule Formation lithotopes is in preparation (H. E. LaGarry and L. A. LaGarry, unpublished data, 1997). For methods and conventions used during the 1991–1995 mapping, and supplemental descriptions of several measured sections used here, see LaGarry and LaGarry (1997). Legal descriptions of measured section localities are in the Appendix.

Brule Formation stratotypes

Darton (1899) did not designate a stratotype for the Brule Clays, but Schultz and Stout's (1955) revision did refer to specific sections. Schultz and Stout's Brule Formation is based on 6 measured sections at Toadstool Park, and 15 measured sections in Scottsbluff County, Nebraska, and adjacent Goshen County, Wyoming. Schultz and Stout (1955, p. 38) designated Darton's (1899) Roundtop to Adelia measured section as the type section for the Brule Formation, and used nearby Toadstool Park as their "typical" locality for their Orella and Whitney Members. Schultz and Stout did not designate stratotypes for their Orella and Whitney Members, but used the six sections of the Brule Formation at Toadstool Park as a composite stratotype. The North American Stratigraphic Code (NACSN, 1983, Article 22c) stipulates that once a type section or locality is established it cannot be changed or supplanted. Therefore, Darton's (1899) Roundtop to Adelia measured section and Schultz and Stout's (1955, Fig. 3, sections 4–9) sections at Toadstool Park remain as the type sections for the Brule Formation and its members, respectively. Measured sections described here, unless a redescription of one of Schultz and Stout's (1955) sections, are supplemental reference sections. Figures and sections within the parentheses of a citation can be found within the cited work. Formal designation and description of a stratotype of the brown siltstone member of the Brule Formation will be presented elsewhere (Swinehart, personal communication, 1996).

Basal Arikaree Group and the "brown siltstone" member of the Brule Formation

Darton (1899) named the Gering Formation to describe exposures of predominately fluvial sandstones and conglomerates in the southern panhandle of Nebraska. Subsequently, Vondra et al. (1969) formally defined the Gering Formation in the North Platte River Valley, but included within it rocks now assigned to the brown siltstone and Whitney Members of the Brule Formation (Tedford et al., 1996). In northwestern Nebraska, Darton (1899, Fig. 226) described the Gering Formation (Fig. 3) in his Roundtop to Adelia section as consisting of "stratified and cross-bedded sands," a "hard white bed," and "loose gray sands with clay and pebbly streaks." Schultz and Stout (1955) also assigned the beds overlying the Whitney Member of the Brule Formation near Toadstool Park to the Gering Formation (Figs. 2, 3). However, Swinehart et al. (1985) and Swinehart and Diffendal (1995, 1997) argued, as mentioned previously, that the Gering Formation is restricted to pumaceous sandstones and associated strata confined to a paleovalley exposed along Wildcat Ridge in the southern Nebraska panhandle. Based on this work, basal Arikaree beds in northwestern Nebraska previously recognized by Darton and Schultz and Stout as the Gering Formation must be redescribed.

The brown siltstone of Swinehart et al. (1985) is a distinct lithologic unit composed of pale brown and brown volcaniclastic

sandy siltstones and volcaniclastic silty sandstone, very fine grained sandstones, and thinly bedded mudstone. In the subsurface, this unit is bounded by regional unconformities. In the southern panhandle of Nebraska, it overlies the Whitney Member of the Brule Formation and underlies the Gering Formation (*sensu* Swinehart and Diffendal, 1995, 1997). The brown siltstone, like the underlying Whitney Member of the Brule Formation, consists of 50 to 60% glass shards, compared to the 20 to 30% typical of the overlying sandstones of the Arikaree Group (Swinehart et al., 1985; Tedford et al., 1996). The Nonpareil ash zone (NP_{1-3}) (Tedford et al., 1996) occurs within the lower 10 to 30 m of this unit, and is composed of as much as three discrete, closely spaced ash beds (Swinehart et al., 1985, Fig. 5). Along Wildcat Ridge in the southern panhandle of Nebraska, as much as 135 m of the brown siltstone is exposed. Locally, streams that deposited fluvial mudstones and sandstones within the brown siltstone incised through the Nonpareil ash zone and the underlying upper Whitney ash (Tedford et al., 1996). Based on extensive bore-hole data, Swinehart et al. (1985, Fig. 5) recognized their brown siltstone in exposures at the base of the Pine Ridge in central Dawes County, Nebraska, ~48 km southeast of Toadstool Park. At Toadstool Park, Terry et al. (1995) referred to correlative rocks as the "Roundtop member" (Fig. 3) of the Brule Formation. On their geologic map of the Roundtop 7.5′ Quadrangle, LaGarry and LaGarry (1997) referred to the brown siltstone as the "Horn member" of the Brule Formation (Fig. 3) anticipating its formal description and definition by Swinehart (personal communication, 1996). The Nonpareil ash zone is the only named marker bed within this unit.

Observations. In northwestern Nebraska the Arikaree Group crops out as 100 to 140 m of massive, grayish-brown sandstones in the vertical cliffs of the Pine Ridge escarpment. In most areas, the basal Arikaree Group and underlying strata are covered by Quaternary colluvium and slope wash. However, in the Roundtop 7.5′ Quadrangle these strata are exposed in a series of hills that form a northeastward extension of the Pine Ridge. One of these hills is Roundtop, a prominent landmark 4 km south of Toadstool Park and the top of Darton's (1899) Roundtop to Adelia measured section (Fig. 4). A composite section of the rocks containing the contact between the Arikaree and White River Groups can be observed here and in canyons to the north (Fig. 5).

Exposures at Wright Hill (Fig. 6A) and in a hill to the northeast (Fig. 6B) consist of brownish-gray medium to thickly bedded volcaniclastic sandstones (Fig. 4, sections 1, 2). In section 1 (Fig. 5), the sandstones are tabular, massive, contain abundant "pipey" and "potato" concretions (see Schultz, 1941), calcareous root traces, and rhizoliths, and vary from well cemented to poorly indurated. The beds are from 0.2 to 4.0 m thick: the majority of the thinner beds contain fewer concretions, are well cemented, and stand out as ledges. The thicker sandstone beds are poorly cemented and weather more readily. In section 2 (Fig. 5) the concretions, root traces, and rhizoliths are abundant at the top of the section and become less frequent near the basal sandy siltstone. The bases of these sections are covered by vegetation and slope wash.

The concretionary sandstones in sections 1 and 2 intertongue eastward with sandstones exposed on a hill west of Roundtop (Fig. 6C) and on the upper flanks of Roundtop (Fig. 6D). In section 3 (Fig. 5), the exposures are of dark brownish-gray, medium-bedded, well-cemented, trough and cross-bedded conglomeratic sandstones interbedded with medium- to thickly bedded, poorly indurated, grayish- or yellowish-brown volcaniclastic sandstones. The poorly indurated sandstones show differential compaction that likely indicates weakly expressed bedding. In section 6 (Fig. 5) this sequence consists of 28 m of medium-bedded, well-indurated, dark grayish-brown conglomeratic sandstones interbedded with brownish-gray, massive, and poorly indurated volcaniclastic sandstones. The well-indurated sandstones are thicker (up to 1.2 m) than in section 3, and coarse to medium grained with less gravel and fewer cobbles. The massive volcaniclastic sandstones are from 3 to 5 m thick and contain calcareous rhizoliths and a few small potato-shaped concretions. The well-indurated sandstones become less common toward the base of section 6, and only the poorly indurated sandstones remain. Toward the base of section 6 the massive sandstones contain interbeds of medium-bedded, poorly indurated, grayish-brown, laminated and cross-laminated volcaniclastic sandstones.

Several small hills at the base of Roundtop and to the north (Fig. 4) are capped by the cross-bedded conglomeratic sandstones mentioned above (Fig. 5, sections 4, 5, 7) or a thin, tabular, white or gray, well-cemented fine-grained sandstone (Fig. 5, section 8). In sections 4 and 5 the conglomeratic sandstone is underlain by poorly indurated sandstones similar to those in section 6. However, below the sandstones in sections 7 and 8 are sequences of massive, pale brown volcaniclastic siltstones and sandy siltstones (Fig. 6E). Zones of differential compaction or cementation within the volcaniclastic siltstones appears to be weakly defined cross-bedding.

Sandstones similar to those in sections 3 and 6 are exposed in sections 16 and 17 along Little Cottonwood Creek (Fig. 7). These grayish-brown, massive, poorly indurated volcaniclastic sandstones are interbedded with dark grayish-brown, well-indurated, trough and planar cross-bedded medium- and coarse-grained sandstones. These sandstones contain very few calcareous rhizoliths and root traces, and no concretions. The cross-bedded sandstones occur in lenticular and tabular beds 0.4 to 0.6 m thick. At the base of this sequence is 1.5 to 2 m of dark brownish-gray, well-indurated, planar and trough cross-bedded conglomeratic sandstone. This sandstone is coarser grained than in those in sections 3 and 6, and contains well-rounded pebbles and cobbles <6 cm in diameter. Most of the larger clasts are composed of various igneous and metamorphic rocks and chert. No vitric shards or lithic clasts were observed in these cross-bedded sandstones. Below this sandstone is about 2.5 m of massive pale brown siltstone that is partly covered by vegetation and slope wash.

The conglomeratic sandstone observed in sections 3–7 and 16–17 also occurs as caprock atop several small hills along Little

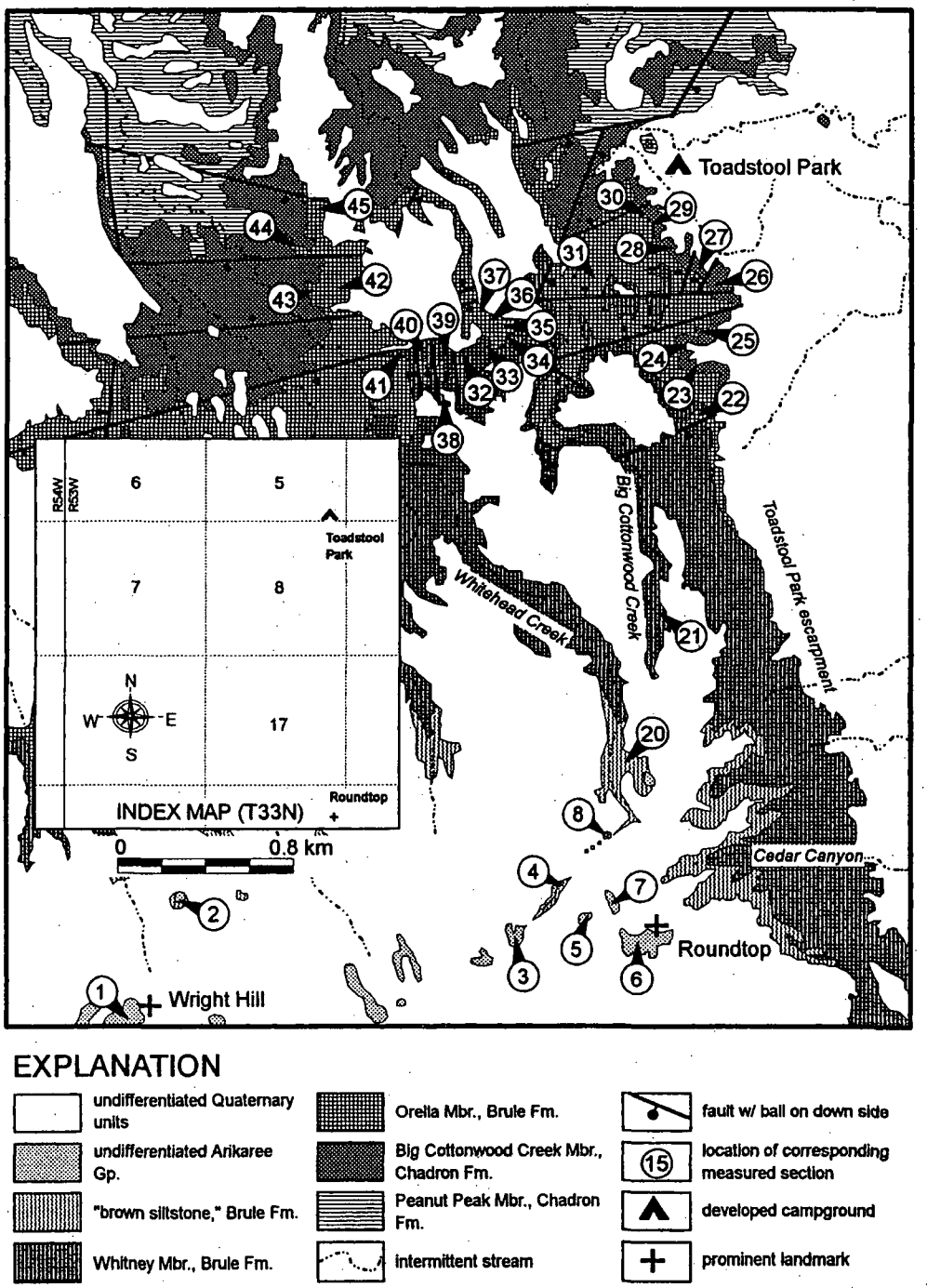


Figure 4. Geologic map of the Roundtop/Toadstool Park vicinity (Roundtop 7.5′ Quadrangle, Sioux County, Nebraska), including the type areas of Darton's (1899) Brule Formation (Brule Clay) and Schultz and Stout's (1955) Whitney and Orella Members. See Figure 1 for map location.

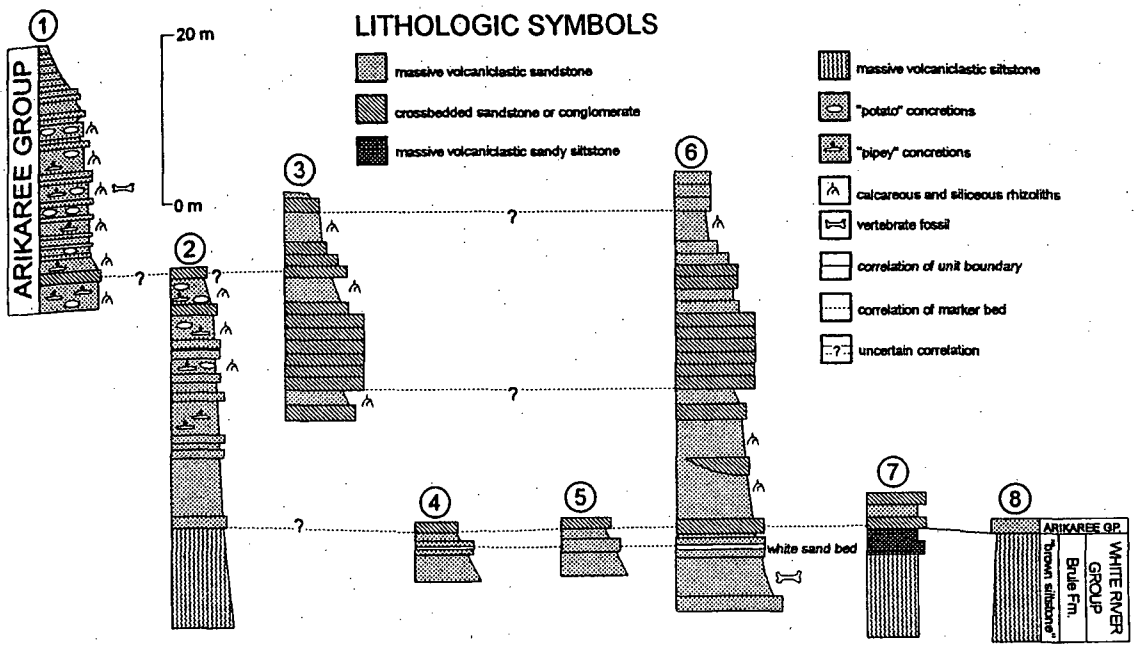


Figure 5. Measured sections near Roundtop. See Figure 4 for locations. Sections 1 and 3–8 modified from LaGarry and LaGarry (1997).

Cottonwood Creek (Fig. 7, sections 12, 13, 18, 19). Below these sandstones (Fig. 8A–E) are the same massive, pale brown volcaniclastic sandstones seen in sections 7 and 8. Section 19 contains massive volcaniclastic siltstones interbedded with well-indurated, weakly bedded pale brown and brown volcaniclastic siltstones, lenses of well-indurated, cross-bedded brown and pale brown volcaniclastic sandy siltstones, and thinly bedded (0.05 and 0.12 m thick), dark reddish-brown laminated claystones (Fig. 8F). A prominent volcanic ash 0.3 to 0.4 m thick occurs within sections 13 and 16 (Fig. 7).

Several exposures of the massive volcaniclastic siltstones and sandy siltstones mentioned above occur along Little Cottonwood Creek (Fig. 7, sections 9–11, 14, 15). In sections 9 and 10 these massive pale brown volcaniclastic siltstones and sandy siltstones contain a 2.0 to 2.5 m interval of small (2–5 cm diameter), irregularly shaped calcareous nodules. In section 11 the nodules occur ~5 m above a lithologic contact with massive, pale brown, nodular siltstones. This contact is partially covered by vegetation, but appears sharp and undulating. Two intervals measured along Little Cottonwood Creek (Fig. 7, sections 14 and 15) show a gradational contact between the brown and pale brown massive volcaniclastic siltstones and sandy siltstones and underlying pale brown, massive, nodular volcaniclastic siltstone.

Section 13 (Fig. 7) contains an interval of the volcaniclastic siltstone and sandy siltstone having contacts with overlying and underlying lithologies. Overlying the volcaniclastic siltstones and sandy siltstones are grayish-brown, planar cross-bedded volcaniclastic sandstones. This contact is likely an erosional unconformity, in that the overlying cross-bedded sandstones appear to be a fluvial valley fill incised into the volcaniclastic siltstone. Underlying the volcaniclastic siltstone and sandy siltstones are pale brown, massive, nodular volcaniclastic siltstones. Based on the appearance of this lower contact, the volcaniclastic siltstones and sandy siltstones likely filled a valley or depression in the underlying volcaniclastic nodular siltstone.

These volcaniclastic siltstones and sandy siltstones also occur along Whitehead Creek (Fig. 4). In section 20 (Fig. 9) the volcaniclastic siltstones and sandy siltstones are overlain by a thin, tabular, hard, fine-grained white sandstone as at section 8 (Fig. 6E,F). Underlying the volcaniclastic siltstones and sandy siltstones are pale brown, massive, nodular (3–6 cm in diameter), slope-forming, volcaniclastic siltstones. This contact is abrupt and steeply dipping, and appears to follow the walls of the arroyo containing Whitehead Creek. A 0.1- to 0.5-m-thick volcanic ash bed occurs in section 20 at about the same stratigraphic position as the ash in sections 13 and 16 (Fig. 7). An ash at this stratigraphic position is also visible in the walls of Cedar Canyon east of Roundtop (Fig. 10B). Section 20 contains an ash bed in the lower 5 m of the volcaniclastic siltstones and sandy siltstones that is apparently continuous with an ash in the underlying lithology exposed in the canyon walls (Fig. 10A). This ash is 0.1 to 0.2 m thick and is highly intermixed with epiclastic sediment.

Interpretations. LaGarry and LaGarry (1997) did not divide the Arikaree Group in these areas into formations but mapped the

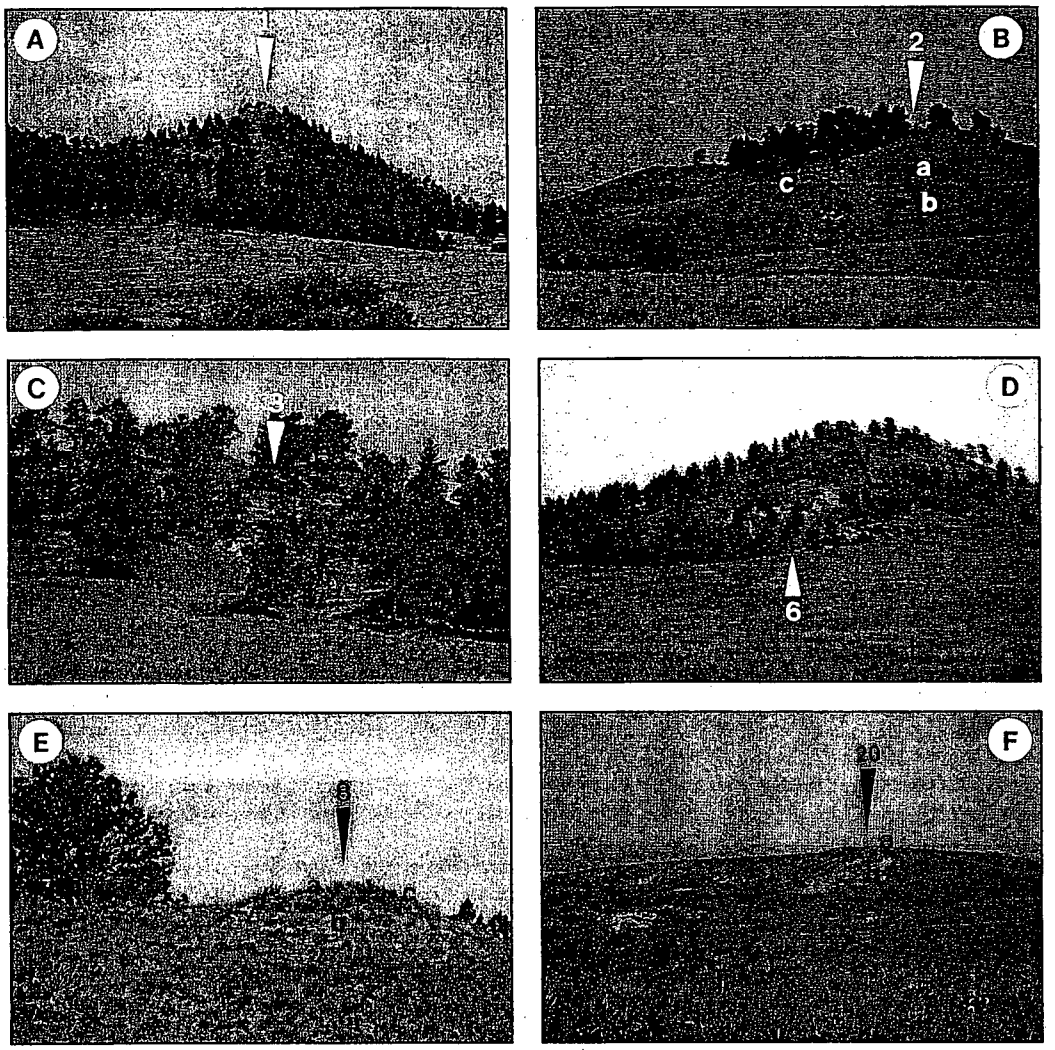


Figure 6. Photographs of outcrops and measured sections (arrows) near Roundtop, Sioux County, Nebraska (c = contact; a= Arikaree Group; b = White River Group). A, Wright Hill and section 1; B, hill to the northeast of A and section 2; C, Little Roundtop and section 3; D, Roundtop and section 6; E and F, small hills north of Roundtop and sections 8 and 20, respectively. See Figure 4 for locations.

various sandstones as undifferentiated Arikaree Group. However, to fully describe the contact between the White River and Arikaree Groups, the lithologies adjacent to the contact must be clearly defined. Based on the preceding observations, four lithotopes are recognized in the sections described above. The stratigraphically highest lithotope consists of the medium-bedded, grayish-brown fine- to medium-grained tabular sandstones having abundant rhizoliths and concretions exposed at Wright Hill (Fig. 5, sections 1, 2). This lithotope is undoubtedly part of the Arikaree Group, and likely corresponds to Darton's (1899, Fig. 226) "gray sand with pipey concretions" (Fig. 3) and the Monroe Creek Formation of Swinehart et al. (1985). However, the abundance of resistant, tabular interbeds suggests that this lithotope intertongues with the next described lithotope.

The second and next lower lithotope consists of interbedded, tabular, and massive grayish-brown fine- to medium-grained sandstones, tabular and lenticular cross-bedded dark brown conglomeratic sandstones, and interbedded grayish-brown, medium-grained, poorly indurated volcaniclastic sandstones (Fig. 5, sections 3–8; Fig. 7, sections 12–19). The poorly indurated sandstones of this lithotope likely correspond to Darton's (1899, Fig. 226) "loose gray sands with clay and pebbly streaks." A

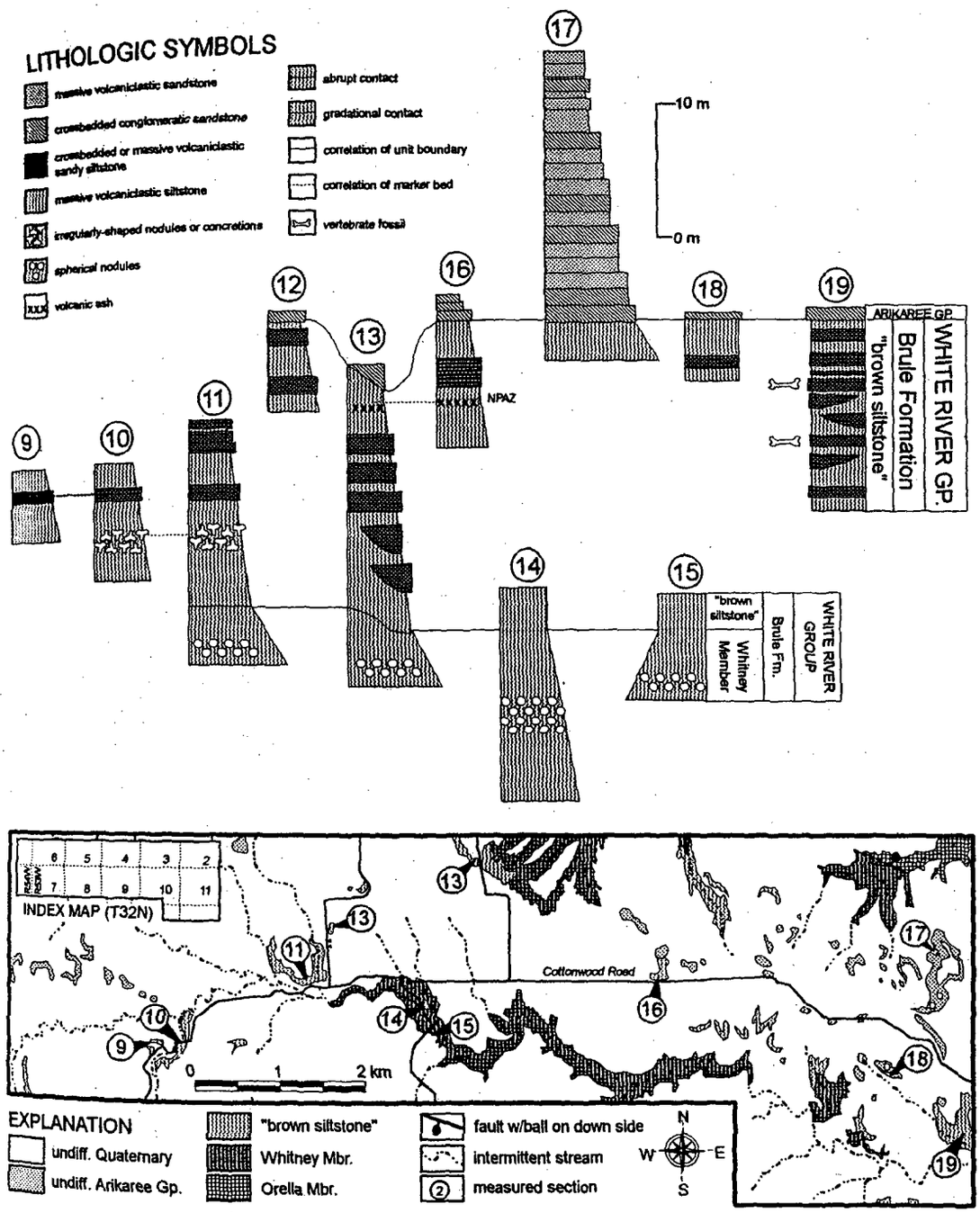


Figure 7. Geologic map and measured sections along Little Cottonwood Creek (Roundtop 7.5′ Quadrangle, Sioux County, Nebraska). NPAZ = Nonpareil ash zone of Swinehart et al. (1985). See Figure 1 for map location. Sections 12 and 16–19 and geologic map after LaGarry and LaGarry (1997).

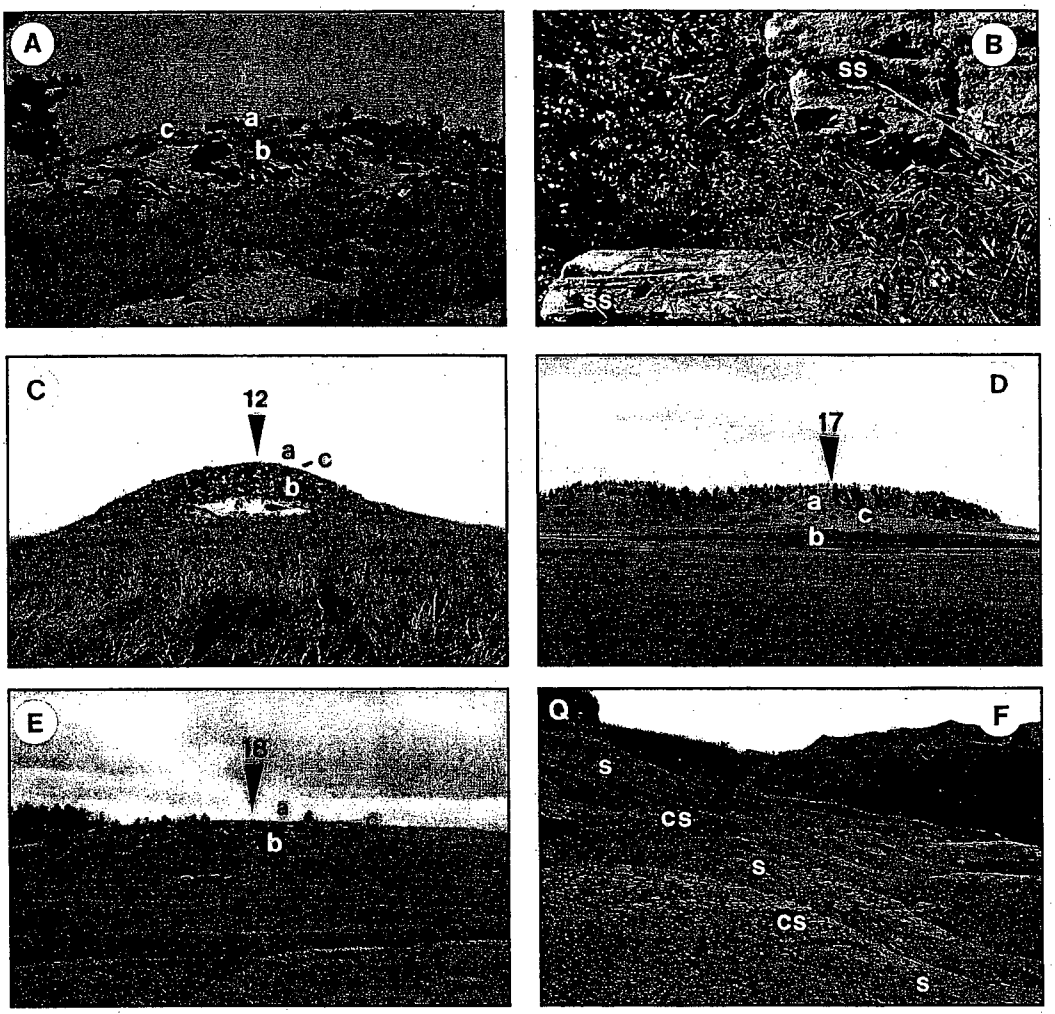


Figure 8. Photographs of outcrops and measured sections along Little Cottonwood Creek, Sioux County, Nebraska (c = contact; a= Arikaree Group; b = White River Group): A, Typical outcrop showing contact between Arikaree and White River Groups; B, close-up of basal Arikaree sandstone (ss) in A; C, D, and E, sections 12, 17, and 18, respectively; F, section 19, showing contact with overlying Quaternary (Q) deposits, massive and/or laminated siltstone (s), and cross-bedded sandy siltstone (cs). See Figure 4 for locations.

resistant tabular sandstone bed of this lithotope occurs at the contact with the next lower lithotope in sections 8 and 20 (Fig. 6E,F), and likely corresponds to Darton's "hard white bed." Darton (1899), Schultz and Stout (1955), and many later workers considered sandstones of this lithotope to represent the Gering Formation. These rocks are similar to the Gering Formation in that both consist of fluvial sandstones at the base of the Arikaree Group. However, the Gering Formation is restricted to a paleovalley in the southern Nebraska panhandle (Swinehart and Diffendal, 1995, 1997). The basal Arikaree sandstones of northwestern Nebraska likely represent an unnamed, primarily fluvial interval of the Arikaree Group that is coeval with the Gering Formation as suggested by Tedford et al. (1996) and MacFadden and Hunt (this volume).

The third and next lower lithotope consists of volcaniclastic siltstones and sandy siltstones of the brown siltstone of Swinehart et al. (1985). These beds were also referred by Schultz and Stout (1955) to the Gering Formation (Fig. 3). However, this lithotope was likely deposited as windblown silt and fine sand (see Swinehart, 1995; Souders and Swinehart, 1996; Tedford et al., 1996), and shows no features consistent with fluvial deposition, with the exception of the fluvial reworking of these sedi-

ments seen in section 19 (Figs. 7, 8F). The ash within this lithotope exposed in the canyons east and north Roundtop (Fig. 10B) has been correlated to the Nonpareil ash zone (NP_3) (Tedford et al., 1996), further supporting the assignment of these beds to the brown siltstone. Because this lithotope is similar in lithology to underlying Brule lithotopes (see the following discussion), the assignment of the brown siltstone to the brule formation by Swinehart et al. (1985) is followed here.

The fourth and stratigraphically lowest lithotope displays all of the distinguishing characteristics of Schultz and Stout's (1955) lithologically uniform Whitney Member and is discussed later within the context of more complete measured sections (Figs. 10–14).

Upper contact/boundary. The upper lithostratigraphic boundary of the brown siltstone member of the Brule Formation, and the revised upper boundary of the White River Group, is herein recognized at the lithologic contact between fluvial sandstones (Arikaree Group) and eolian siltstones (White River Group). The tabular fine-grained sandstone above the contact in sections 8 and 20 was not traced beyond the small hills north of Roundtop (Fig. 6E,F). However, the well-indurated, cross-bedded, and conglomeratic fluvial sandstones capping sections 12–16 and 10–11 (Fig. 7; Fig. 8A–D) can be traced east into central Dawes County to the tops of several prominent hills north of the Pine Ridge escarpment, and west into Niobrara County, Wyoming, where it caps the Sherrill Hills. Isolated exposures of the conglomeratic sandstones above this contact have been mistaken for sandstones within the White River Group (Swinehart, personal communication, 1996). However, sandstones within the White River Group of northwestern Nebraska are typically restricted to a few well-known channels within the lower Brule, Chadron, and Chamberlain Pass Formations (LaGarry and LaGarry, 1997; Terry, this volume; Terry and LaGarry, this volume), are typically finer grained and poorly indurated, and often contain sand grains supported by a siltstone or claystone matrix (Evans and Terry, 1994; LaGarry et al., 1996).

Nonpareil ash zone. Near Roundtop and along Little Cottonwood Creek, the brown siltstone member has a maximum thickness of 33 m. The volcanic ashes in sections 13 and 16, and the upper ash in section 20 (Figs. 5, 7), as in the subsurface and exposed at Wildcat Ridge, occur within the first 30 m above the lower contact. The ash at the base of section 20 (Fig. 10A) is likely reworked from the underlying Whitney Member rather than an air-fall ash within the brown siltstone. The Nonpareil ash zone (NP_3) (Tedford et al., 1996) exposed in the canyon east of Roundtop (Fig. 10B) was $^{40}AR/^{39}AR$ dated at 30.050 ± 0.19 Ma (Swisher and Prothero, 1990). There are no widespread marker beds that can be used to determine an approximate stratigraphic position within this lithotope, which means that most local and regional correlations require knowledge of the upper or lower contact and the position of the Nonpareil ash zone.

Lower contact/boundary. The lower lithostratigraphic boundary of the brown siltstone member is the lithologic contact between volcaniclastic siltstones and sandy siltstones of the brown siltstone member and nodular volcaniclastic siltstones of the Whitney Member. This change in lithotope is usually accompanied by a pronounced change in outcrop slope and appearance (Fig. 10C,D). The contact is an erosional unconformity as in sections 11, 13, and 20 (Figs. 7, 9), or gradational as in sections 14 and 15 (Fig. 7). The overall impression is that the volcaniclastic siltstones and sandy siltstones of the brown siltstone member filled paleovalleys in the underlying lithology, and that these paleovalleys are being exhumed by erosion. The brown siltstone member was also deposited within a complex system of paleovalleys in the Wildcat Ridge (Tedford et al., 1996). Schultz et al. (1955) reported a minor paleosol in the brown siltstone (their Gering Formation) above the lower contact with the Whitney Member, but no paleosols were recognized within the brown siltstone during this study.

Whitney Member of the Brule Formation

Schultz and Stout's (1955) Whitney Member consisted of 85 to 95 m of massive, pink siltstone and claystone (Figs. 2, 3). The loess-like character of the Whitney Member has been discussed by Tychsen (1954), Harvey (1956), Schultz and Falkenbach (1968), Swinehart et al. (1985), Meehan (1994), Souders and Swinehart (1996), and Tedford et al. (1996). Schultz and Stout (1955) subdivided the Whitney Member into the "Whitney A," "B," and "C" at the "upper" and "lower Whitney ashes." Their upper boundary was a "fractoconformity" at the contact with their overlying Gering Formation, and their lower boundary was the "white bed" (Fig. 3).

In addition to the upper Whitney Member intervals described along Little Cottonwood Creek (see preceding discussion), the Whitney Member is prominently exposed along Big Cottonwood and Whitehead Creeks, and in the Toadstool Park escarpment (Fig. 4). During the 1991–1995 mapping, several sections of the Whitney Member were measured at Toadstool Park (Fig. 9, sections 20–23, 31, 32–36, 38–40) and along Dirty Creek (Fig. 14, sections 46–52). The Whitney Member is poorly exposed eastward, but is well exposed to the west, especially in the areas of Lady in a Shoe (Fig. 15, sections 54–57), Warbonnet Buttes, and the Sherrill Hills.

Observations. The only complete measured section through the pale brown, massive, nodular volcaniclastic siltstone described here is section 21 (Figs. 4, 9). The upper 5 m of section 21 consists of pale brown, massive, highly fractured, cliff-forming volcaniclastic siltstone. At the base of the cliff is a sharp, undulating contact with 83 m of underlying pale brown, massive, nodular, slope-forming volcaniclastic siltstone. The nodules (3–10 cm in diameter) consist of calcite-cemented volcaniclastic siltstone. They are typically horizontally aligned, and occur in discontinuous zones from 0.1 to 6.5 m thick at irregular intervals down the section (Fig. 11A). Two prominent volcanic ashes occur at 30 and 53 m down-section (Fig. 10C,D). The upper and lower ashes are 1.1 and 1.6 m thick, respectively. Both are diffuse zones of ash intermixed with volcaniclastic sediment, and have a

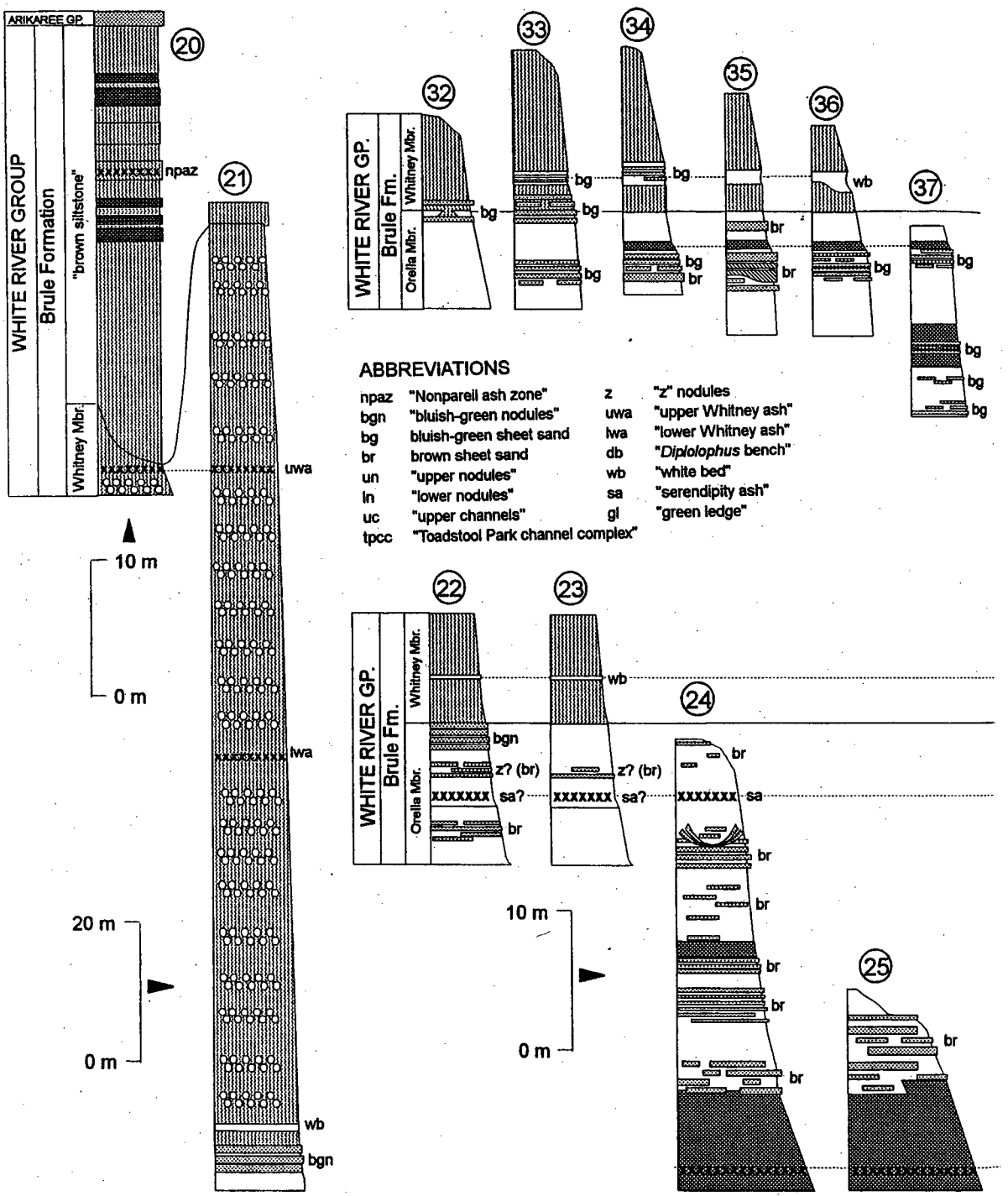
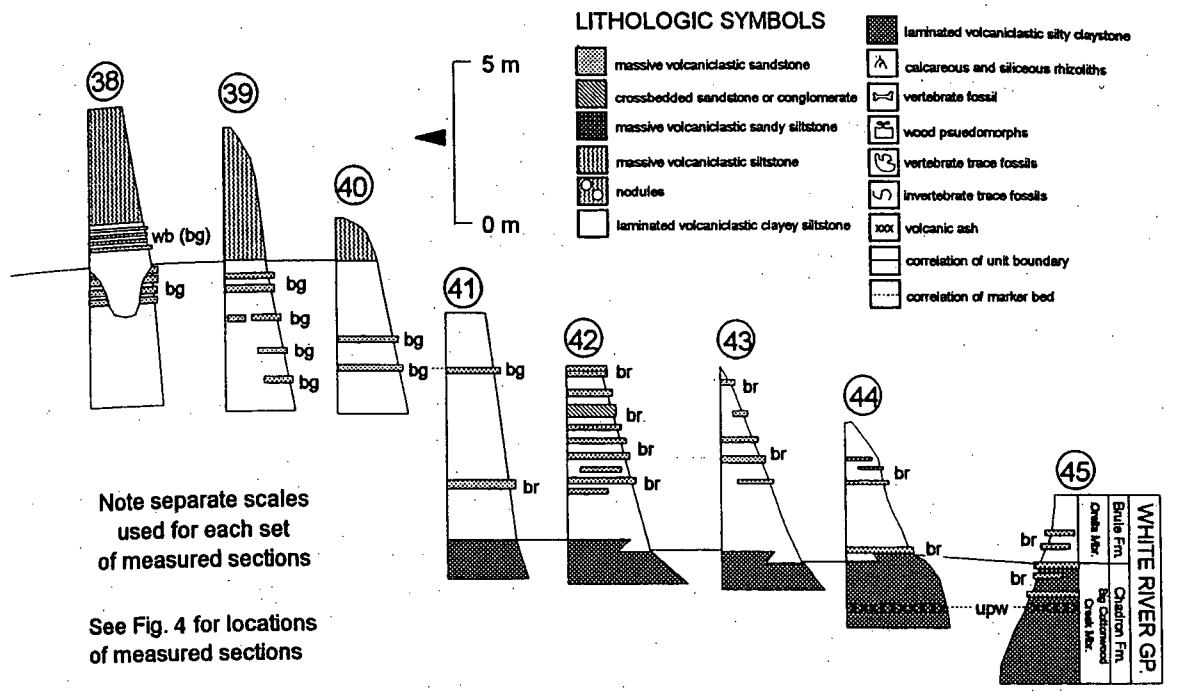
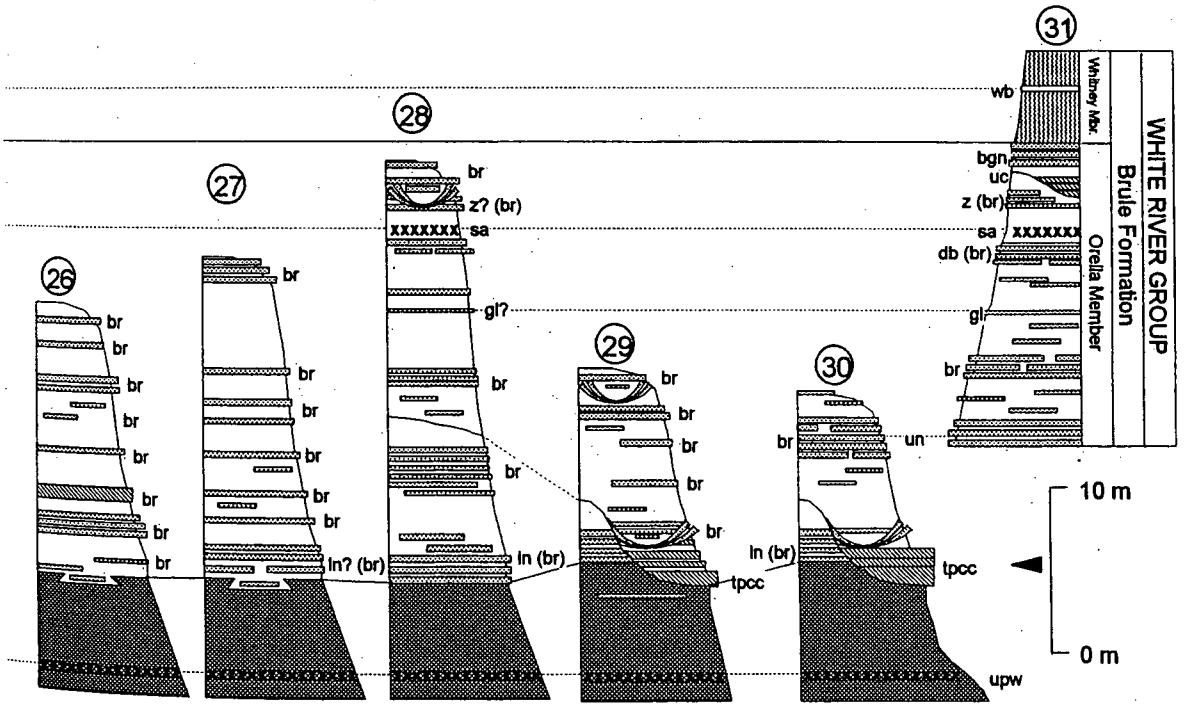


Figure 9 (above and facing page). Measured sections along Big Cottonwood and Whitehead Creeks north of Roundtop, and the Toadstool Park escarpment. tpcc = Toadstool Park channel complex of Schultz and Stout (1955). Sections 20 and 21 modified from LaGarry and LaGarry (1997).


LITHOLOGIC SYMBOLS
massive volcaniclastic sandstone
crossbedded sandstone or conglomerate
massive volcaniclastic sandy siltstone
massive volcaniclastic siltstone
nodules
laminated volcaniclastic clayey siltstone
laminated volcaniclastic silty claystone
calcareous and siliceous rhizoliths
vertebrate fossil
wood psuedomorphs
vertebrate trace fossils
invertebrate trace fossils
volcanic ash
correlation of unit boundary
correlation of marker bed
5 m
0 m
Note separate scales used for each set of measured sections
See Fig. 4 for locations of measured sections
WHITE RIVER GP.
Chadron Fm.
Big Cottonwood Creek Mbr.
Brule Fm.
Orella Mbr.


WHITE RIVER GROUP
Brule Formation
Orella Member
Whitney Mbr.
10 m
0 m

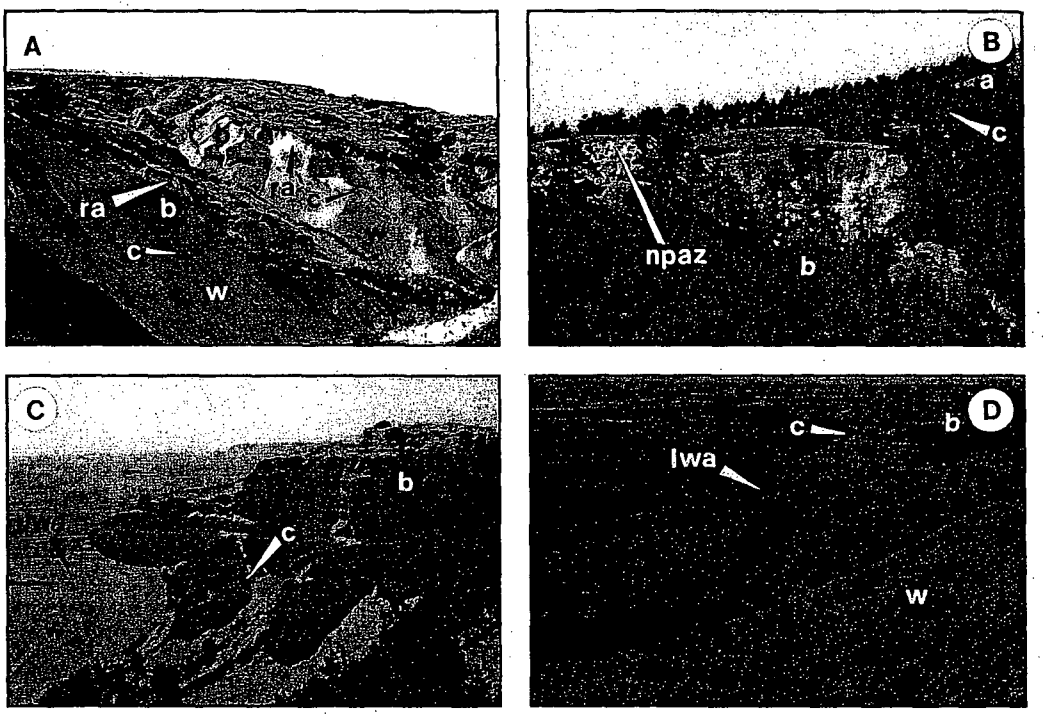


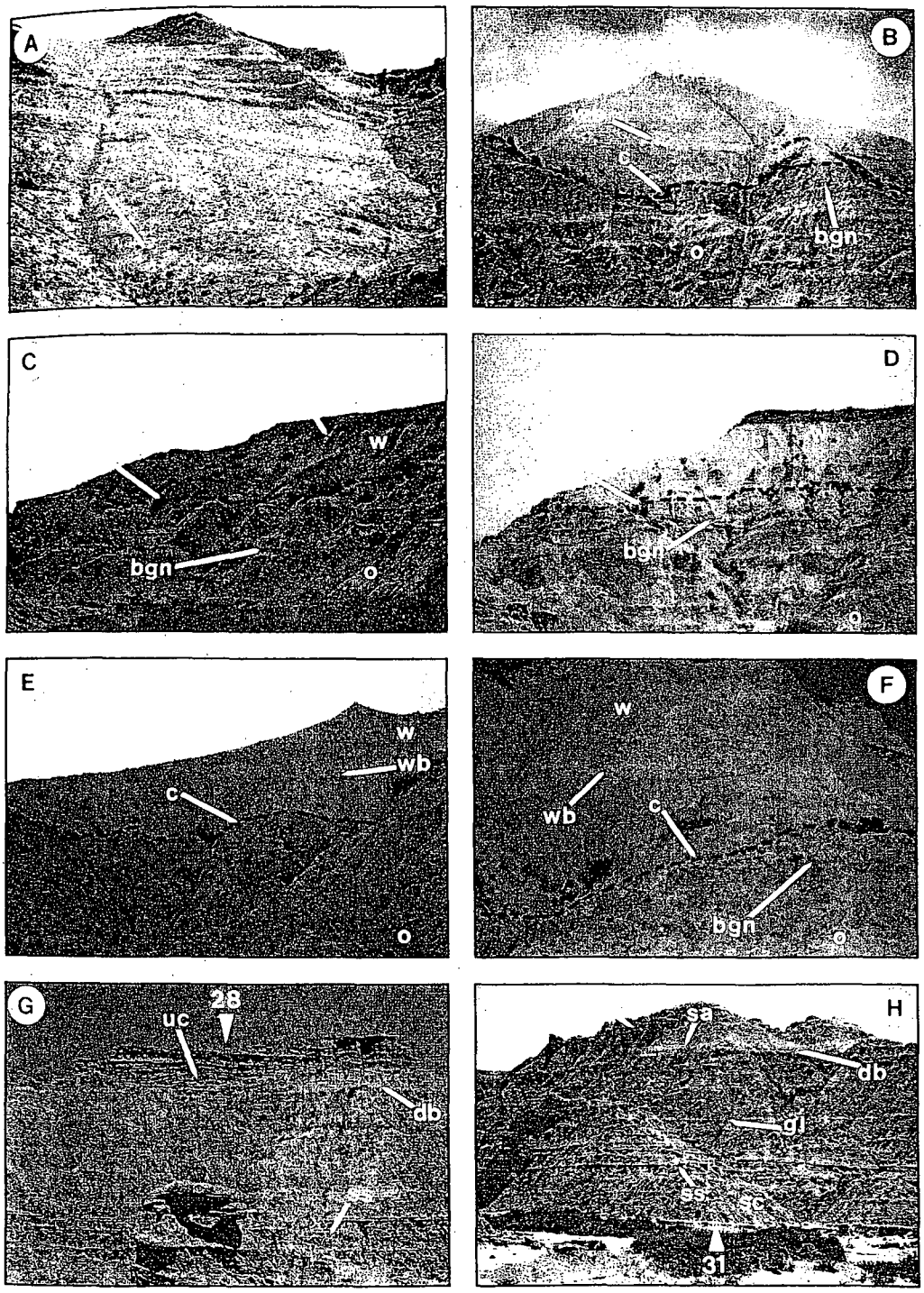
Figure 10. Outcrops and measured sections in canyons north and east of Roundtop showing the upper (uwa) and lower Whitney (lwa) ashes, and lithologic contacts (c) between the Arikaree Group (a), the brown siltstone (b), and Whitney (w) Members of the Brule Formation. A, measured section 20 along Whitehead Creek, showing the reworked upper Whitney ash (ra) within the brown siltstone member; B–D, sequence of photographs (east to west) of the Nonpareil ash zone (npaz) sample locality of Swisher and Prothero (1990).

higher vitric content at their base. The lower ash is surrounded and overprinted by pale green mottles. At the base of the section is a thin (0.2–0.3 m) green, gray, or grayish-white siltstone bed. This bed is locally continuous along the canyon wall, and stands out in shallow relief. Below 84.5 m the uniform, nodular siltstone intertongues with underlying thinly interbedded sandstones and clayey siltstones. The sandstones are bluish-green nearest the nodular siltstone or brown down-section. They are thin (0.05–0.2 m thick), laterally discontinuous, fine to medium grained, and massive, laminated, or cross-laminated.

The sequence of pale brown, massive, nodular volcaniclastic siltstone is uniform in composition with two exceptions. The first is the green, gray, or grayish-white siltstone bed (Fig. 11B–F). This bed is as described above in exposures in the Toadstool Park escarpment (Fig. 9, sections 22, 23, 31) and along Dirty Creek (Figs. 13,14, sections 46–52). Along Whitehead Creek it is more discontinuous and occurs as 3 m of laminated siltstone filling trenches in the underlying siltstone (Fig. 9, sections 32–36, 38–40; Fig. 13B–F). In sections 33, 34, and 38 the laminated siltstones are interbedded with thin (<3 cm), discontinuous lenticular sheets of bluish-green fine grained sandstone (Figs. 9, 13A). In the westernmost part of the study area (Fig. 1), this stratigraphic position within the Whitney Member is occupied by a discontinuous multistoried channel sandstone (Fig. 15, section 57). The channel sandstone is predominately medium to coarse grained, and contains trough and planar cross-bedding, plane bed lamination, scour marks, tool marks, and invertebrate and vertebrate trace fossils.

The second exception to the lithologic uniformity of the nodular volcaniclastic siltstone is the position of the sheet sandstones near its base. In the Toadstool Park escarpment the thin

Figure 11. Outcrops and measured sections along Big Cottonwood Creek and the Toadstool Park escarpment. A, Whitney Member in section 21, showing nodules (n) and lower Whitney ash (lwa); B–F, view of sections 22 (east face), 23, Schultz and Stout's (1955) section 5 (S+S5), and 22 (distant and detailed view of west face), respectively, showing the contact (c) between the Whitney (w) and Orella (o) Members of the Brule Formation, the white bed (wb), and the bluish-green nodules (bgn); G and H, sections 28 and 31, showing "serendipity ash" (sa), sheet sandstones (ss), laminated silty claystone (sc), and the upper channels (uc), *Diplolophus insolens* bench (db), and green ledge (gl) of Schultz and Stout (1955). See Figure 4 for locations.


A
B
c
o
bgn
C
w
bgn
o
D
w
bgn
o
E
w
wb
c
o
F
w
wb
c
bgn
o
G
uc
28
db
H
sa
db
gl
ss
sc
31

green and brown sandstones are restricted to the clayey and silty claystones underlying the massive nodular (Fig. 9, sections 22, 23, 31; Fig. 11B–F). Along White- (Fig. 9, sections 32–34, 38; Fig. 13A–G) and Dirty Creeks (Fig. 14, section 51), these discontinuous sheet sandstones occur the lowest 2 to 5 m of nodular volcaniclastic siltstone.

Interpretations. Section 21 is one of Schultz and Stout's (1955, Fig. 3, section 8) type sections for the Whitney Member. Their other type section (Schultz and Stout, 1955, Fig. 3, section 9) was described along Whitehead Creek, and parts of it are shown in sections 32–36 and 38–40 (Fig. 9). Based on the above observations, the Whitney Member is redescribed as two lithotopes. The first Whitney lithotope consists of pale brown, massive, nodular, eolian volcaniclastic siltstone. This lithotope comprises most of the volume of the Whitney Member, and includes all strata below the lithologic contact with the brown siltstone member and above the lithologic contact with interbedded sheet sandstones and volcaniclastic clayey siltstones.

The second Whitney lithotope consists of the discontinuous green, gray, or grayish-white siltstones, laminated siltstones, laminated siltstones interbedded with thin lenticular sheets of bluish-green, fine-grained, discontinuous sheet sandstones, and multistoried channel sandstones at the stratigraphic position of Schultz and Stout's (1955) white bed. Schultz and Stout (1955), Schultz et al. (1955), and Stout (1978, Fig. 5) interpreted the white bed to represent the terminal Orella D paleosol, and the channel sandstones to be the basal channels of the following Whitney A cycle. Singler and Picard's (1981) detailed study of this bed showed that it is not a paleosol. Based on the constant stratigraphic position of the white bed and its laterally equivalent siltstones and channel sandstones, their deposition was probably synchronous. These laminated siltstones, interbedded laminated siltstones and sheet sandstones, and channel siltstones and sandstones likely represent different facies within a small, isolated fluvial system. This lithotope disappears to the east of Toadstool Park (H. E. LaGarry and L. A. LaGarry, unpublished data, 1997) and becomes sandier to the west (Fig. 16, section 57). It likely represents a fluvial system that became sand starved as flow progessed eastward, or a disappearing stream that sank into the underlying loess of the dominant Whitney lithotope.

Upper contact/boundary. As mentioned previously, Schultz and Stout (1955) described the contact of the Whitney Member with the overlying brown siltstone member (their Gering Formation) as a fractoconformity. This term was likely coined to describe the highly jointed appearance of the upper parts of the exposed Whitney Member. Schultz and Stout (1955) and Schultz et al. (1955) described a major paleosol complex within the Whitney Member below this contact (Fig. 3). However, no clear evidence of paleosols was observed within this interval.

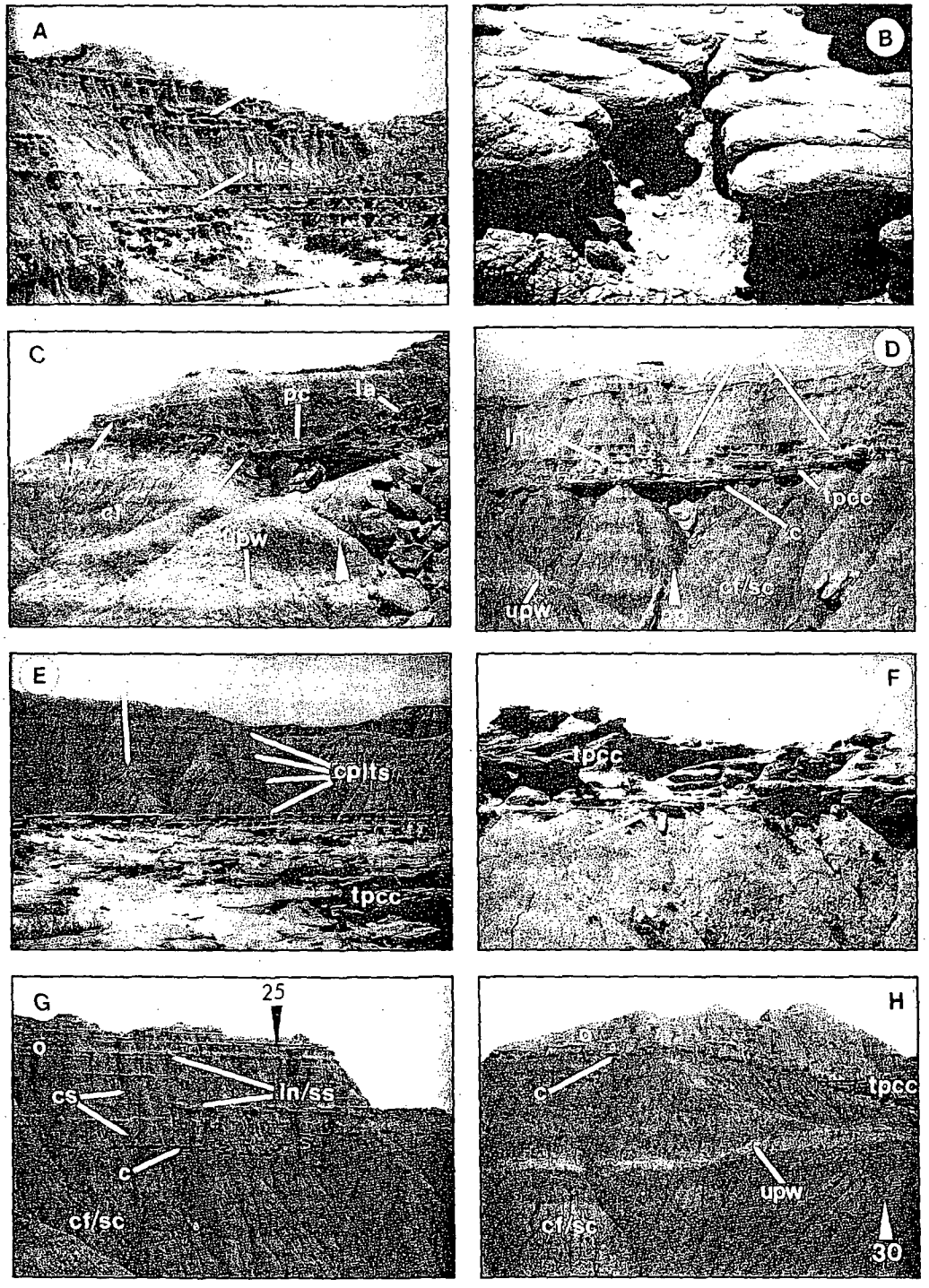
Upper and lower Whitney ashes. Although Schultz and Stout (1955) recognized 4 or 5 volcanic ashes within the Whitney Member, the upper and lower Whitney ashes are the most prominent and the only ones named (Fig. 3). Prothero and Swisher (1992) radiometrically dated the upper Whitney ash at 31.85 ± 0.02 Ma. The lower Whitney ash has yielded dates of 31.846 Ma (biotite), 31.811 Ma (anorthoclase), and 30.673 Ma (plagioclase). According to Deino (personal communication, 1996), new $^{40}Ar/^{39}Ar$ ages of 31.24 ± 0.06 Ma and 31.29 ± 0.06 Ma support geochemical and magnetopolarity data indicating a correlation of the lower Whitney ash to the Windous Butte Formation of east-central Nevada and western Utah.

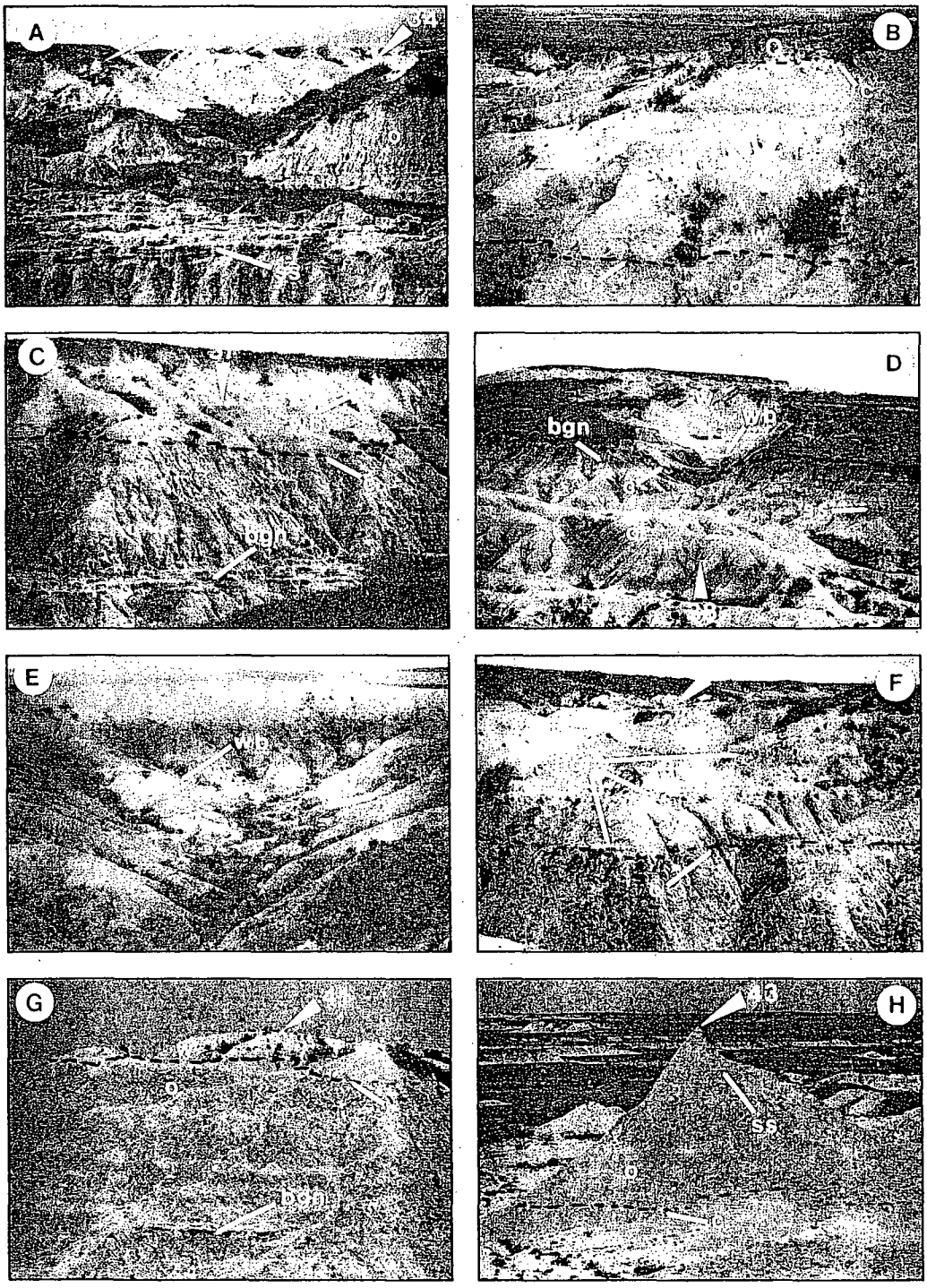
Schultz and Stout (1955) and Schultz et al. (1955) indicated that a prominent, regional paleosol was present below the lower ash (Figs. 3, 11A). This conclusion was supported by Singler and Picard (1981). However, based on detailed petrographic study of the Whitney Member, Meehan (1994) concluded that there was no paleosol present below the lower ash, and that little or no clear evidence of paleosol development was present within the Whitney Member. In the vicinity of Chimney Rock in the southern Nebraska panhandle, Schwartman (1989) reported trace fossils from the lower Whitney ash, and interpreted this ash to represent six discreet, closely spaced eruptions. Meehan (1994) reported trace fossils in the form of earthworm burrows (the green mottles reported here) from the lower ash indicating a stable land surface, but found no other evidence of pedogenesis.

White bed (X) and associated paleosols and channels. Schultz and Stout (1955) placed the lower boundary of their Whitney Member at the white bed (Fig. 3). Schultz and Stout (1955, Fig. 3, sections 8, 9) recognized this bed in only one (section 9) of their two type sections (the photograph in Fig. 11F was taken at the base of their section 8 along Big Cottonwood Creek). The white bed is local, of variable lithology, difficult to recognize, and does not constitute a valid lithostratigraphic boundary in that there is no change in lithology across it (NACSN, 1983). This discontinuous and lithologically variable horizon has only limited use for correlation in the vicinity of Toadstool Park.

Lower contact/boundary. The revised lower lithostratigraphic boundary of the Whitney Member is the lithologic contact between the Whitney lithotopes and underlying interbedded sandstones and clayey siltstones. This contact is intertonguing, such that the contact resembles interlocking dovetails, and the pale brown, massive, nodular volcaniclastic siltstone is intercalated with the underlying lithotope (Fig. 11B–F; Fig. 13A–G). This zone of intercalation is generally about 2 m thick, and is likely the result of the gradual, but episodic, burial of the depositional environment of the Orella lithotopes by those of the Whit-

Figure 12. Outcrops and measured sections along Big Cottonwood Creek and the Toadstool Park escarpment. A and B, Orella Member (Brule Formation) within Toadstool Park showing upper and lower nodules (un, ln), sheet sandstones (ss), and laminated clayey siltstones (cs); C–D, sections 30 and 29, respectively, showing the lower contact (c) with the Chadron Formation (cf), the upper purplish-white layer (upw), silty claystones (sc), the upper and lower nodules (un, ln), sheet sandstones (ss), the Toadstool Park Channel complex (tpcc), paleochannels (pc), and lateral accretion surfaces (la); E and F, Orella paleovalley fill within Toadstool Park and incision into underlying Chadron Formation; G and H, sections 25 and 30, respectively, showing intertonguing contact (c) between Brule and Chadron Formations. See Figure 4 for locations.


A
B
C
pc
lb
cf
upw
D
tpcc
c
cf/sc
upw
E
cplts
tpcc
F
tpcc
G
25
o
cs
ln/ss
c
cf/sc
H
o
c
tpcc
upw
cf/sc
30


A
B
C
D
bgn
E
F
G
bgn
H
ss

ney Member. Clark et al. (1967) described a similarly intertonguing contact between the predominately fluvial Scenic and the predominately eolian Poleslide Members of the Brule Formation in the Big Badlands of South Dakota.

Orella Member of the Brule Formation

Schultz and Stout's (1955) Orella Member consisted of 56 to 85 m of interbedded massive pink silty clays, laminated green silty clays, laminated pink, brown, and buff silty clays, nodular sandstone, siltstone, claystone, gypsum, gypsiferous paleosols, and channel sandstones (Fig. 2). The Orella Member was subdivided (Orella A, B, C, and D) based on the presence of regional paleosols and associated unconformities, and valley incision and backfilling (Fig. 3). The upper boundary was the white bed, and the lower boundary was the upper purplish-white layer. The Orella Member contains the majority of Schultz and Stout's (1955) named marker beds within the White River Group.

The Orella Member is well exposed in the Toadstool Park escarpment, in badlands surrounding Toadstool Park, along Whitehead Creek (Fig. 4), along Sand and Dirty Creeks (Fig. 14), and north of Lady in a Shoe (Fig. 15). Like the overlying Whitney Member, the Orella Member is only intermittently exposed to the east, but is well exposed in areas to the west.

Observations. Below the pale brown, massive, nodular volcaniclastic siltstone in section 21 (Fig. 9) is 26 to 29 m of thinly interbedded sheet sandstones, volcaniclastic clayey siltstones and silty claystones, and single and multistoried cross-bedded sandstones. The thickest and most prominent of these sandstones is at the base of the sequence (Fig. 9, sections 36, 37).

The sheet sandstones are thin (<0.2 m), fine to medium grained, and discontinuous (Fig. 12A). Most are brown, while others are a striking bluish-green that is likely the result of secondary mineralization in the silt and clay fractions. They are laminated, cross-laminated, cross-bedded, or massive, and contain dewatering veins, worm burrows, and mudcracks. Many contain chalcedony-filled root traces. Most exposed sheet sandstones are jointed every 10 to 30 cm, which causes the sandstones to weather into rounded masses (Fig. 13A,B). Some of this jointing is the result of fractures propagated through adjacent strata. Others follow what appear to be mudcracks on the upper surfaces of many of the thin sandstone beds. Many of the sheet sandstone beds are not jointed or mudcracked, and retain a tabular appearance in weathered profile. There are at least seven "zones," or dense stratigraphic sequences, of these sheet sandstones in vertical succession within Toadstool Park (Fig. 9, sections 22–31). The uppermost of these zones is often, but not always, the striking bluish-green mentioned above. These zones of sheet sands do not persist more than 50 to 100 m laterally. In the stratigraphic intervals between zones, sheet sandstones occur intermittently, are thinner (< 0.05 to 0.15 m), less continuous, and extend laterally for only a few tens of meters.

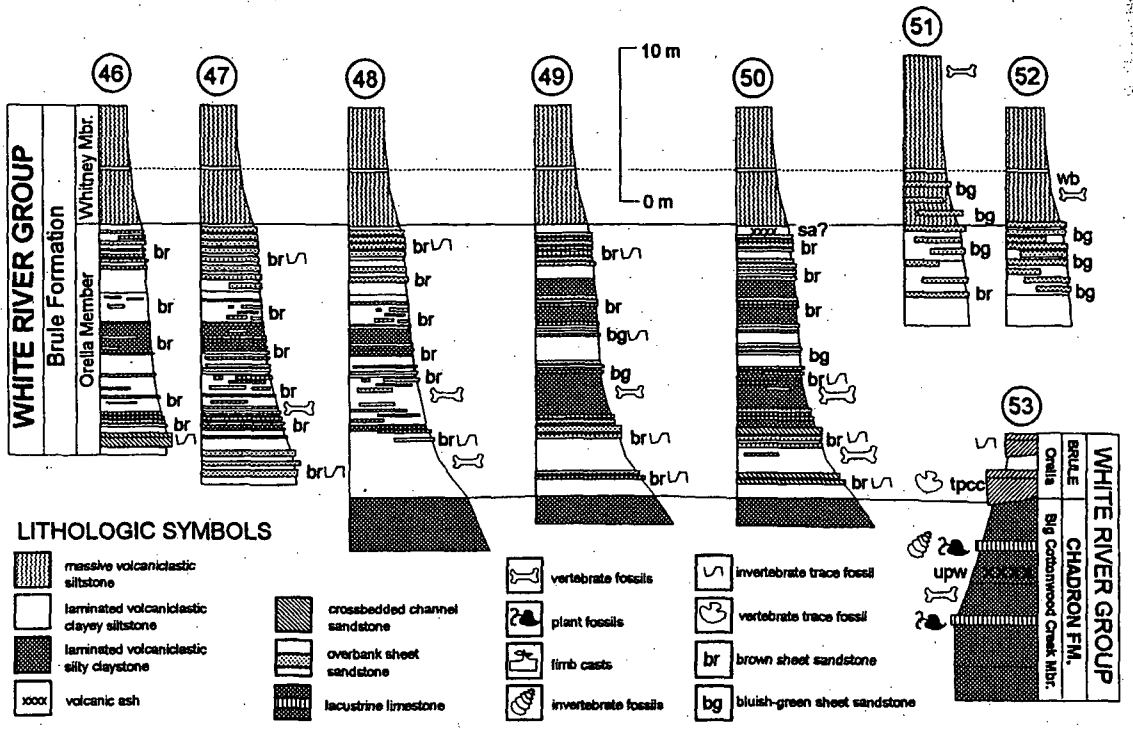
The sheet sandstones are separated by one or more beds of brownish-orange, pale brown, and brown laminated or massive volcaniclastic clayey siltstone, and near the base of the sequence by minor amounts of volcaniclastic silty claystone. The volcaniclastic clayey siltstones and silty claystones contain vertebrate fossils, chalcedony-filled root traces, chalcedony wood pseudomorphs, and chalcedony or calcite-filled structures interpreted to be insect egg chambers. Like the sandstone interbeds, the clayey siltstones and silty claystones are discontinuous and lenticular, but are generally more laterally persistent than the sheet sands.

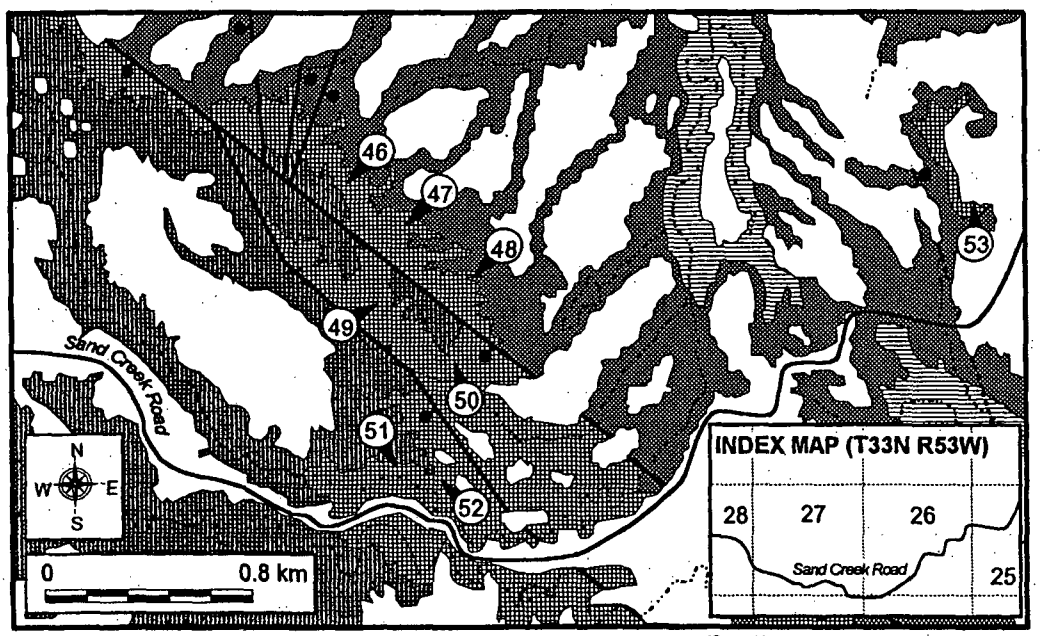
At various levels within the sequence are single storied channel sandstone bodies within small, meandering paleochannels (Fig. 9, sections 24, 28–31). These paleochannels are 2 to 5 m deep and 5 to 10 m wide, and are well defined by lateral accretion surfaces (Fig. 11G,H; Fig. 12C,D). In cross section these sandstone bodies are cross-bedded, cross-laminated, and lenticularly bedded with clay and silt drapes. A small paleovalley begins within the middle of the sequence and incises through underlying interbedded brown sheet sandstones and brownish-orange, pale brown, and brown volcaniclastic clayey siltstones and silty claystones, and into underlying green and pinkish-gray silty claystones (Fig. 9, sections 28–30; Fig. 13D,F). This small paleovalley is 12 to 14 m deep. There is a volcanic ash ~20 m above the base of the paleovalley (Fig. 9, sections 22–24, 28, 31; Fig. 11H). Individual beds thin toward paleochannel walls forming wedges.

Within the bottom of the small paleovalley is a multistoried cross-bedded channel sandstone (Fig. 12F). Individual beds within this channel sandstone are grayish-brown and 0.05 to 1.5 m thick. The sandstones are fine to coarse grained, are occasionally siltstone or claystone matrix supported, and contain a suite of fluvial sedimentary structures, including trough, planar, and climbing ripple cross-beds, lateral accretion surfaces, plane bed laminations, scour and tool marks, gravel lag and pebble stringers, and invertebrate and vertebrate trace fossils. Thinner beds are typically laminated or cross-laminated, and fine to medium grained. Thicker beds typically contain planar or trough cross-beds, climbing ripples, or plane bed laminations. Many of the cross-bedded sandstone beds have gravel lags and stringers. Within some of the basal lags and stringers are small, weathered fragments of fossil bone and tortoise shell. The basal sandstone of the paleovalley fill defines the greatest thickness of the interbedded sandstone and clayey siltstones in the study area.

Underlying the sequence of thinly interbedded sandstones, laminated volcaniclastic clayey siltstones, and volcaniclastic silty claystones are green, pale brown, and pinkish-gray laminated volcaniclastic silty claystones (Fig. 9, sections 24–30, 41–45; Fig. 14, sections 48–53; Fig. 15, 69, 60). The contact of

Figure 13. Outcrops and measured sections along Whitehead Creek showing the contacts between the Whitney and Orella Members (A-G) and the Brule and Chadron Formations (H). Abbreviations: unconsolidated Quaternary deposits (Q), Whitney (w) and Orella (o) Members of the Brule Formation, Chadron Formation (cf), bluish-green nodules (bgn) and white bed (wb) of Schultz and Stout (1955), lithologic contact (c), fault (f), and sheet sandstone (ss). See Figure 4 for locations.





MAP EXPLANATION

- undifferentiated Quaternary units
- Whitney Mbr., Brule Fm.
- Orella Mbr., Brule Fm.
- Big Cottonwood Creek Mbr., Chadron Fm.
- Peanut Peak Mbr., Chadron Fm.
- fault w/ ball on down side
- (15) location of corresponding measured section
- intermittent stream

the interbedded sandstones, clayey siltstones, and silty claystones with the underlying laminated volcaniclastic silty claystones is intertonguing, and resembles interlocking dovetails (Fig. 12C,G,H; Fig. 13H). In some areas the volcaniclastic clayey siltstones and volcaniclastic silty claystones above this contact are intercalated within a zone ~2 m thick. At several locations the intertonguing contact and zone of intercalation are incised by the channel sandstones in the base of the prominent valley fill (Fig. 9, sections 29, 30; Fig. 14, section 53; Fig. 15, sections 59, 60). Where incised by the paleovalley, the contact is a local erosional unconformity (Fig. 12C,D,F).

Interpretations. Sections 30 and 31 (Fig. 9) are parts of Schultz and Stout's (1955, Fig. 3, sections 4–7) composite type section for the Orella Member. Based on a lithologic reinterpretation, the Orella Member consists of two lithotopes. The first consists of interbedded brown and bluish-green sheet sandstones and brownish-orange, pale brown, and brown laminated and massive volcaniclastic clayey siltstones, infrequent volcaniclastic silty claystones, and volcanic ash (the "serendipity ash"). This lithotope is generally similar to the lithologies within Clark et al.'s (1967) Scenic Member of the Brule Formation in the Big Badlands of South Dakota, except that the Scenic Member contains abundant nodules (Retallack, 1983; Wells et al., 1995) and the Orella Member does not. Based on this redescription, Schultz and Stout's (1955) Orella D, which consists of pale brown, massive, nodular volcaniclastic siltstones (Fig. 11B), is part of the dominant Whitney Member lithotope (Fig. 3). Likewise, Terry and LaGarry (this volume) redefined Schultz and Stout's (1955) Orella A, which consists of 8 to 10 m of green, pale brown, and pinkish-gray volcaniclastic silty claystones, as part of their Big Cottonwood Creek Member of the Chadron Formation. The second lithotope within the Orella Member consists of the single and multistoried sandstone bodies occurring throughout the unit, and includes Schultz and Stout's (1955) upper channels and Toadstool Park channel complex along with several unnamed channel fills (Fig. 11G,H; Fig. 12C,D,F).

LaGarry and LaGarry (1997) measured a total thickness of 26 to 29 m for the Orella Member, which corresponds to the maximum topographic relief for their measured sections. In contrast, Schultz and Stout (1955) reported a thickness of 45 to 49 m for the same interval. LaGarry and LaGarry (1997) attributed this disparity to the imbricate faulting prevalent in this area, and argued that Schultz and Stout (1955) may have inadvertently measured parts of this stratigraphic interval several times.

Wells (1994), Wells et al. (1994, 1995), and Terry et al. (1995) interpreted the Orella Member to represent a stacked sequence of overbank sheet sandstones and mudstones. They followed Harvey (1960) in using the term "couplets" to describe these packages (Fig. 12E), which were likely produced by repeated flood events that eventually filled and overtopped a complex system of multiple cut-and-fill paleochannel sequences. Terry et al. (1995) reported slopes ranging from 5° to 36° degrees along the exhumed walls of the small paleovalley. According to Harvey (1960) and Terry et al. (1995), deposition of the Orella Member was the result of two or more separate and alternating periods of incision and backfilling. Harvey's interpretation was of repeated, cyclic deposition within nested paleovalleys. The recent work of Terry et al. (1995) suggested a more complex system of overlapping and episodic cutting and filling. The multiple stratigraphic occurrences of paleochannels reported here supports the conclusion of the latter workers. Unlike Schultz and Stout (1955) and Schultz et al. (1955), Clark et al. (1967) argued that there was no definitive evidence of paleosols within the Orella Member, and that the Orella Member was dominated by overbank sheet deposition rather than paleosol development. Terry et al. (1995) described the overbank clayey siltstones and silty claystones within each couplet as containing "entisols" based on the presence of root traces and very limited soil structure or horizonation.

Bluish-green, Y, and Z nodules. The uppermost of Schultz and Stout's (1955) zones of nodules are the "bluish-green nodules" (Fig. 3). Like all of their nodules, these were assumed to be the result of the pedogenic accumulation of minerals. These nodules are actually jointed overbank sheet sandstones that weather into rounded masses (Fig. 12B). Also, the bluish-green color occurs in sheet sandstones throughout the overbank lithotope, and a stratigraphic concentration of these sheet sandstones is not always present at any particular level (e.g., Fig. 9, sections 22, 23; Fig. 13G). These same conclusions apply to Schultz and Stout's (1955) Y and Z nodules (Fig. 9, sections 22, 23, 28, 31). These beds have no value as markers except in the specific sections figured by Schultz and Stout (1955).

Upper channels. Schultz and Stout's (1955) upper channels were considered to represent the basal sandstones of their Whitney A. In their discussion of these channels, they apparently considered the paleochannels in section 31 (Fig. 9) and the channel sandstones and siltstones within the white bed to be equivalent (see Schultz and Stout, 1955, Fig. 3). Their upper channels, if correctly identified as those in section 31 (Fig. 12H), represent paleochannels observed at several stratigraphic positions within the Orella Member (Fig. 9, sections 24, 28–31), and are useless as stratigraphic markers, even within Toadstool Park.

Diplolophus insolens bench. Schultz and Stout (1955) considered this bench to represent a gypsiferous or calcareous paleosol, from which they collected several specimens of *Diplolophus insolens* (Fig. 12G,H). The white horizon that they recognized as a gypsum or calcrete is the "serendipity ash" of Swinehart and Evanoff (personal communication, 1993). No evidence of paleosol formation was observed within this ash, although it has not been studied in detail. This ash bed occurs in the same stratigraphic position along the Toadstool Park escarpment (Fig. 9, sections 22, 23, 28, 31), and because it is the only ash recognized within surface exposures of the Orella Member, it is an important datum within the Brule Formation. The serendipity ash has not been dated.

Figure 14. Geologic map and measured sections along Dirty Creek and its tributaries, Sioux County, Nebraska. See Figure 1 for location of mapped area. Abbreviations as in previous figures.

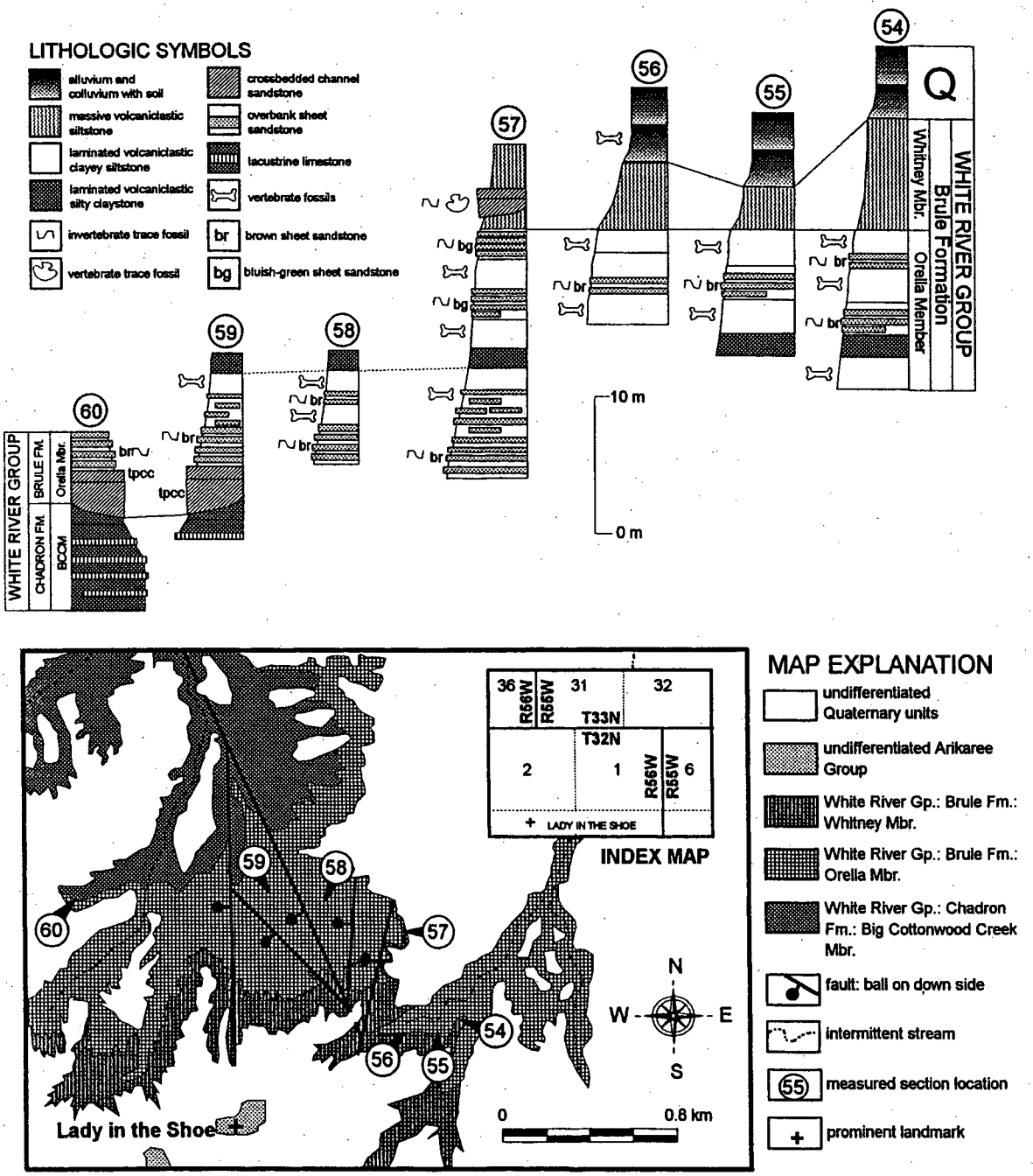


Figure 15. Geologic map and measured sections at University of Nebraska State Museum localities Sx-6 and Sx-7, the "Old Floyd Hall Place." See Figure 1 for location of mapped area. tpcc = Toadstool Park channel complex of Shultz and Stout (1955). Geologic map modified after LaGarry (1995a).

Green ledge. Schultz and Stout's (1955) green ledge is a thin (0.03 m) greenish-blue sheet sandstone bed (Fig. 9, sections 28, 31). This bed was traceable no more than 160 m through Toadstool Park (Fig. 11H). It has no value as a stratigraphic marker bed except within the sections figured by Schultz and Stout (1955).

Upper and lower nodules. Schultz and Stout (1955) described two dense stratigraphic concentrations of the sheet sands as their upper and lower nodules (Fig. 12A). These nodules were thought to be pedogenic nodules formed below regional unconformities (Schultz et al., 1955), and Schultz and Stout (1955) correlated them to the Bump's (1956) upper and lower nodular zones. Harvey (1960), Vondra (1960), Wells (1994), Wells et al. (1994, 1995), and Terry et al. (1995) have mentioned that these beds are overbank sheet sands rather than pedogenic nodules, and therefore are not lithologically correlative to nodular zones in the Big Badlands of South Dakota. There is no evidence for a lengthy or widespread break in sedimentation above, or associated with, these sheet sandstones, and because as many as seven of these zones were observed throughout the overbank lithotope in variable stratigraphic positions (see preceding discussion), they likely have no value for correlation. Calcareous siltstone nodules have been observed in the westernmost part of the study area, but they are rare and localized occurrences (H. E. LaGarry, unpublished data, 1997).

Toadstool Park channel complex. Schultz and Stout (1955) recognized that their Toadstool Park channel complex (Fig. 3) was formed late in deposition of the Orella Member and incised into the underlying Chadron Formation (as defined by Terry and LaGarry, this volume). The Toadstool Park channel complex and its associated overbank deposits represent one of several episodes of fluvial incision and backfilling within the Orella Member (Wells et al., 1994). The Toadstool Park channel complex contains two main facies: a channel sandstone facies and an overbank sandstone and mudstone facies. Terry et al. (1995) reported trough cross-bedding, planar tabular cross-bedding, cross-lamination, plane bed lamination, tabular bedding, slumping, load casting, sand volcanoes, dewatering veins, lateral-accretion surfaces, and mudstone blocks and chips derived from bank failure within the channel sandstone facies. Based on the vertical and lateral arrangement of these structures, and on comparisons to sedimentary structures and facies in modern rivers, Terry et al. (1995) concluded that the Toadstool Park channel complex was likely deposited in a seasonal mixed-load river system.

The depositional environments of the Orella Member are currently being restudied (W. B. Wells, unpublished data, 1997) because of the recent recognition of an important vertebrate trackway site at Toadstool Park. Most of the vertebrate and invertebrate trackways have been observed on the uppermost surface of the Toadstool Park channel complex, although trackways occur at multiple levels within the multistoried basal sandstones. Trackways of at least 3 invertebrate and 11 vertebrate species, including insects, worms, tadpole shrimp, shore birds, ducks, oreodonts, entelodonts, camels, rhinoceroses, and carnivores have been observed along more than 1 km of paleostreambed (Nixon, 1991; Terry et al., 1995; LaGarry et al., 1996).

Lower contact/boundary. The revised lithostratigraphic boundary between the Chadron and Brule Formations is the intertonguing contact (Fig. 12G,H; Fig. 13H) between the interbedded sandstones and volcaniclastic clayey siltstones of the Orella Member of the Brule Formation and the volcaniclastic silty claystones of the Chadron Formation (Fig. 9, sections 24–31, 41–45; Fig. 14, sections 48–53; Fig. 16, section 59, 60). In the Toadstool Park area, Schultz and Stout's (1955) lower nodular zone occurs directly above this contact (Fig. 12G), while along Whitehead Creek this contact has no marker bed associated with it (Fig. 13H).

SUMMARY AND DISCUSSION

Schultz and Stout's (1955) revision of Darton's (1899) Brule Clay consisted of stratigraphic units defined by lithology, faunas, and inferred geologic history, and used local marker beds as unit boundaries (Fig. 3). Based on recent geologic mapping of the White River Group in the type area of the Brule and Chadron Formations at Toadstool Park, Sioux County, Nebraska, Schultz and Stout's marker beds were discovered to be restricted to the type area. Revised lithostratigraphic units (Figs. 2, 3) were defined based on lithologic and lithogenetic criteria conforming to the North American Stratigraphic Code (NACSN, 1983). In addition to the revision and redescription of the Brule Formation reported here, Terry (this volume) and Terry and LaGarry (this volume) revised, redefined, and redescribed various parts of the Chadron Formation to meet standards established by the North American Stratigraphic Code (NACSN, 1983).

Revision and redescription of the Brule Formation

Previous versions of the North American Stratigraphic Code (ACSN, 1970, p. 31) included geologic-climate units that would have likely included Schultz and Stout's (1955) climatic cycle–based White River Group stratigraphic classification. However, geologic-climate units were abandoned (NACSN, 1983, p. 849) because paleoclimatic inferences were too subjective and tenuous to provide a basis for formal geologic units. Based on guidelines within the current code, Schultz and Stout's unconformity bounded units would likely be considered *chronostratigraphic units* (geochrons) equivalent to material referents for the Chadronian, Orellan, and Whitneyan North American land-mammal ages of Wood et al. (1941), but not *lithostratigraphic units* (NACSN, 1983, Articles 22–30, 62, 63, 77). However, the preceding redescription of lithologies within the Brule Formation demonstrates that Schultz and Stout's (1955) inferred geologic history of the Brule Formation, and the premise for their stratigraphic classification, is not supported by lithologic evidence. The Brule Formation is revised herein to conform to North American Stratigraphic Code (NACSN, 1983) guidelines for the designation and use of *lithostratigraphic units.*

Category and rank. The revised and redescribed Brule Formation in the Toadstool Park area is a lithostratigraphic unit composed of the lithologically defined brown siltstone, Whitney, and Orella Members (NACSN, 1983, Articles 22–25).

Type sections. Once established, the type sections and locality of a lithostratigraphic unit cannot be changed (NACSN, 1983, Articles 8e, 22c). Therefore, Darton's (1899) type section and locality for the Brule Formation, and Schultz and Stout's (1955, Fig. 3) type sections and locality for the Whitney and Orella Members, are retained. Formal definition of the brown siltstone member and designation of a type section are in preparation (Swinehart, personal communication, 1996).

Redescription of units and boundaries. The upper and lower boundaries of the Brule Formation are revised to correspond to lithologic contacts, and the members of the Brule Formation are redescribed as lithogenetic units, or lithotopes, as specified by the North American Stratigraphic Code (NACSN, 1983, Article 23a,e).

Upper boundary of the Brule Formation and the White River Group. The revised lithostratigraphic boundary between the Brule Formation of the White River Group and the overlying Arikaree Group is placed within Darton's (1899) Gering Formation at the abrupt, unconformable lithologic contact between overlying basal Arikaree sandstones and underlying siltstones of the brown siltstone member of the Brule Formation (Fig. 3).

Brown siltstone member lithotope. The lower part of Darton's (1899) Gering Formation below the lithostratigraphic upper boundary of the White River Group is assigned to the brown siltstone member of the Brule Formation of Swinehart et al. (1985). The brown siltstone member is redescribed as a lithotope consisting of pale brown and brown, cliff-forming, indistinctly bedded, occasionally nodular, and massive or weakly cross-bedded siltstones, sandy siltstones, minor amounts of fluvially reworked volcaniclastic siltstone, and volcanic ash (Nonpareil ash zone). The revised lithostratigraphic boundary between the brown siltstone and underlying Whitney Members is the contact between the brown siltstone and Whitney lithotopes, which varies from gradational to sharp and undulating depending on where it is observed. This contact is also an erosional unconformity where the brown siltstone member fills valleys and depressions in the underlying Whitney Member (Fig. 3).

Whitney Member lithotopes. The Whitney Member is redescribed as two lithotopes. The first consists of pale brown, massive, and typically nodular siltstones with occasional thin interbeds of grayish-brown and bluish-green sandstone, and volcanic ash (upper and lower Whitney ashes). The second consists of white or green laminated fluvial siltstones, channel siltstones, and channel sandstones. The revised lithostratigraphic boundary between the Whitney and Orella Members is moved downward 2.5 to 5 m from the white bed to the lithologic contact between the Whitney and Orella lithotopes. This contact is intertonguing except where channel siltstones and sandstones of the Whitney Member cut into the underlying Orella Member (Fig. 3).

Orella Member lithotopes and the lower boundary of the Brule Formation. The Orella Member is redescribed as two lithotopes. The first consists of thinly interbedded and slightly pedogenically modified brownish-orange, pale brown, and brown volcaniclastic clayey siltstones and silty claystones, brown and bluish-green sheet sandstones, and volcanic ash (the serendipity ash). The second consists of single and multistoried channel sandstones. The revised lithostratigraphic boundary between the Brule and underlying Chadron Formations in the Toadstool Park area is recognized 9 to 10 m above its previous location, the upper purplish-white layer. This lithologic contact is intertonguing except where paleovalleys of the Orella Member cut into the underlying Chadron Formation (Fig. 3).

ACKNOWLEDGMENTS

I thank D. O. Terry, Jr., J. B. Swinehart, L. A. LaGarry, R. M. Hunt, Jr., W. B. Wells, and J. E. Evans for months of informative discussion, and I thank Terry, Hunt, LaGarry, R. J. Zakrzewski, M. R. Voorhies, A. Joern, and J. B. Swinehart for reviewing the manuscript. I thank the staff of Nebraska National Forest in Chadron, Nebraska, for the use of work space and office equipment during the 1994 and 1995 field seasons. I thank the field crews who participated in the vertebrate fossil inventories of the Oglala National Grassland; without their assistance, preliminary drafts of the surficial geologic maps could not have been completed. Preliminary geologic mapping during 1991–1993 was supported by Challenge Cost-Share Agreements 02-07-91-013 (1991), 02-07-92-025 (1992), and 02-07-93-011 (1993) between the Nebraska National Forest and the University of Nebraska–Lincoln (University of Nebraska State Museum), and was directed by R. M. Hunt, Jr. Mapping during the 1994 and 1995 field seasons was funded by the U.S. Geological Survey to the University of Nebraska Conservation and Survey Division for mapping of the Alliance 1° × 2° Quadrangle.

APPENDIX: LEGAL DESCRIPTIONS OF MEASURED SECTIONS

All legal descriptions are in Sioux County, Nebraska. All sections on Roundtop 7.5′ quadrangle, except sections marked with an asterisk on Bodarc 7.5′ Quadrangle.

sec.	Legal Description	References
1.	NW1/4NW1/4SW1/4 sec. 19, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)
2.	SW1/4SE1/4SE1/4 sec. 18, T. 33 N., R. 53 W.	
3.	NW1/4SE 1/4NW1/4 sec. 20, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)
4.	SE 1/4NW1/4NE 1/4 sec. 20, T. 33 N., R 53 W.	LaGarry and LaGarry (1997)
5.	NW1/4SW1/4NE1/4 sec. 20, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)
6.	SE1/4NE1/4 sec. 20, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)
7.	SE1/4NE1/4NW1/4 sec. 20, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)

8. NE1/4NW1/4NE1/4 sec. 20, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)	
9. SW1/4SW1/4NE1/4 sec. 12, T. 32 N., R. 54 W.		
10. NE1/4SE1/4NE1/4 sec. 12, T. 32 N., R. 54 W.		
11. SW1/4SE1/4 sec. 6, T. 32 N., R. 53 W.		
12. SW1/4NE1/4SE1/4 sec. 6, T. 32 N., R. 53 W.	LaGarry and LaGarry (1997)	
13. NE1/4SE1/4NE1/4 sec. 5, T. 32 N., R. 53 W.		
14. NE1/4NW1/4 sec. 8, T. 32 N., R. 53 W.		
15. NE1/4SE1/4NW1/4 sec. 8, T. 32 N., R. 53 W.		
16. SE1/4SE1/4SE1/4 sec. 4, T. 32 N., R. 53 W.	LaGarry and LaGarry (1997)	
17. SE1/4SW1/4 SE1/4 sec. 2, T. 32 N., R. 53 W.	LaGarry and LaGarry (1997)	
18. NE1/4NE1/4 SW1/4 sec. 11, T. 32 N., R. 53 W.	LaGarry and LaGarry (1997)	
19. SE1/4SE1/4SE1/4 sec. 11, T. 32 N., R. 53 W.	LaGarry and LaGarry (1997)	
20. E1/2 sec. 17, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)	
21. NE1/4 sec. 17, and SE1/4 sec. 8, T. 33 N., R. 53W.	Schultz and Stout (1955) LaGarry and LaGarry (1997)	
22. NW1/4SW1/4SW1/4 sec. 9, T. 33 N., R. 53 W.		
23. SW1/4NW1/4SW1/4 sec. 9, T. 33 N., R. 53 W.		
24. NE1/4NE1/4SE1/4 sec. 8, T. 33 N., R. 53 W.		
25. NW1/4NW1/4SW1/4 sec. 9, T. 33 N., R. 53 W.		
26. NW1/4SW1/4NW1/4 sec. 9, T. 33 N., R. 53 W.		
27. NW1/4SW1/4NW1/4 sec. 9, T. 33 N., R. 53 W.		
28. SE1/4NE1/4NE1/4 sec. 8, T. 33 N., R. 53 W.		
29. SE1/4NE1/4NE1/4 sec. 8, T. 33 N., R. 53 W.		
30. NW1/4NE1/4NE1/4 sec. 8, T. 33 N., R. 53 W.	Schultz and Stout (1955) LaGarry and LaGarry (1997)	
31. NE1/4SW1/4NE1/4 sec. 8, T. 33 N., R. 53 W.	Schultz and Stout (1955) LaGarry and LaGarry (1997)	
32. NE1/4NW1/4SW1/4 sec. 8, T. 33 N., R. 53 W.		
33. SW1/4SE1/4NW1/4 sec. 8, T. 33 N., R. 53 W.		
34. SE1/4SE1/4NW1/4 sec. 8, T. 33 N., R. 53 W.		
35. SE1/4SE1/4NW1/4 sec. 8, T. 33 N., R. 53 W.		
36. NW1/4SE 1/4NW1/4 sec. 8, T. 33 N., R. 53 W.		
37. NW1/4SE1/4NW1/4 sec. 8, T. 33 N., R. 53 W.		
38. S1/2NW1/ SW1/4 sec. 8, T. 33 N., R. 53 W.		
39. NE1/4NW1/4SW1/4 sec. 8, T. 33 N., R. 53 W.		
40. NW1/4NW1/4SW1/4 sec. 8, T. 33 N., R. 53 W.		
41. SE1/4 SE1/4NE1/4 sec. 7, T. 33 N., R. 53 W.		
42. NW1/4SE1/4NE1/4 sec. 7, T. 33 N., R. 53 W.		
43. NE1/4SW1/4NE1/4 sec. 7, T. 33 N., R. 53 W.		
44. SE1/4NW1/4NE1/4 sec. 7, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)	
45. NW1/4NE1/4NE1/4 sec. 7, T. 33 N., R. 53 W.		
46. NE1/4NW1/4NE1/4 sec. 27, T. 33 N., R. 53 W.		Harvey (1960) in part
47. NW1/4NE1/4NE1/4 sec. 27, T. 33 N., R. 53 W.		Harvey (1960) in part
48. SW1/4NW1/4NW1/4 sec. 26, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)	Harvey (1960) in part
49. NW1/4SE1/4NE1/4 sec. 27, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)	
50. SW1/4SW1/4NW1/4 sec. 26, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)	
51. SW1/4NE1/4SE1/4 sec. 27, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)	
52. SE1/4NE1/4SE1/4 sec. 27, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)	
53. NE1/4NE1/4NW1/4 sec. 25, T. 33 N., R. 53 W.	LaGarry and LaGarry (1997)	Evans and Welzenbach (this volume)
54. NW1/4NW1/4NW1/4 sec. 7, T. 32 N., R. 55 W.*		
55. NE1/4NE1/4NE1/4 sec. 12, T. 32 N., R. 56 W.*		
56. SW1/4NE1/4NE1/4 sec. 12, T. 32 N., R. 56 W.*		
57. NE1/4NW1/4NE1/4 sec. 1, T. 32 N., R. 56 W.*		
58. SE1/4SW1/4SE1/4 sec. 31, T. 33 N., R. 55 W.*		
59. SE1/4SW1/4SW1/4 sec. 31, T. 33 N., R. 55 W.*		
60. SW1/4SE1/4 sec. 36, T. 33 N., R. 56 W.*		

REFERENCES CITED

ACSN (American Commission on Stratigraphic Nomenclature), 1970, Code of stratigraphic nomenclature (second edition): Tulsa, American Association of Petroleum Geologists, 45 p.

Bates, R. L., and Jackson, J. A., eds., American Geologic Institute, 1984, Dictionary of geologic terms (third edition): New York, Anchor Press/Doubleday, 571 p.

Bump, J. D., 1956, Geographic names for members of the Brule Formation of the Big Badlands of South Dakota: American Journal of Science, v. 254, p. 429–432.

Clark, J., 1937, The stratigraphy and paleontology of the Chadron Formation in the Big Badlands of South Dakota: Carnegie Museum Annals, v. 25, p. 261–350.

Clark, J., 1954, Geographic designation of the members of the Chadron Formation in South Dakota: Carnegie Museum Annals, v. 33, p. 197–198.

Clark, J., Beerbower, J. R., and Kietzke, K. K., 1967, Oligocene sedimentation, stratigraphy, paleoecology, and paleoclimatology in the Big Badlands of South Dakota: Fieldiana Geology Memoirs, v. 5, 158 p.

Darton, H. H., 1899, Preliminary report on the geology and water resources of Nebraska west of the one hundred and third meridian: U.S. Geological Survey 19th Annual Report 1897–1898, pt. 4, p. 719–814.

Emry, R. J., 1973, Stratigraphy and preliminary biostratigraphy of the Flagstaff Rim area, Natrona County, Wyoming: Smithsonian Contributions to Paleobiology, v. 18, p. 1–43.

Emry, R. J., Bjork, P. R., and Russell, L. S., 1987, The Chadronian, Orellan, and Whitneyan North American land mammal ages, *in* Woodburne, M. O., ed., Cenozoic mammals of North America: geochronology and biostratigraphy: Berkeley, University of California Press, p. 118–152

Evanoff, E., Prothero, D. R., and Lander, R. H., 1992, Eocene-Oligocene climatic change in North America: the White River Formation near Douglas, east-central Wyoming, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 116–130.

Evans, J. E., and Terry, D. O., 1994, The significance of incision and fluvial sedimentation in the basal White River Group (Eocene-Oligocene), Badlands of South Dakota, U.S.A.: Sedimentary Geology, v. 90, p. 137–152.

Harksen, J. C., and Macdonald, J. R., 1969, Type sections for the Chadron and Brule Formations of the White River Oligocene in the Big Badlands of South Dakota: South Dakota Geological Survey Report of Investigations 99, 23 p.

Harvey, C., 1956, A paleoecological interpretation of the White River faunas of Sioux County, Nebraska [M.S. thesis]: Lincoln, University of Nebraska, 107 p.

Harvey, C., 1960, Stratigraphy, sedimentation, and environment of the White River Group of the Oligocene of northern Sioux County, Nebraska [Ph.D. thesis]: Lincoln, University of Nebraska, 151 p.

Krumbein, W. C., and Sloss, L. L., 1951, Stratigraphy and sedimentation: San Francisco, W. H. Freeman, 497 p.

Krumbein, W. C., and Sloss, L. L., 1963, Stratigraphy and sedimentation (second edition): San Francisco, W. H. Freeman, 660 p.

LaGarry, H. E., 1995a, Geology of the "Old Floyd Hall Place" (UNSM localities Sx-6 and Sx-7), Sioux County, Nebraska: Proceedings, 105th Annual Meeting, Nebraska Academy of Science, p. 55.

LaGarry, H. E., 1995b, Provisional revision of lithostratigraphic boundaries of the Brule Formation in northwestern Nebraska: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A66–67.

LaGarry, H. E., and Hunt, R. M., Jr., 1994, 1993 Paleontological resource survey of the Oglala National Grassland (Montrose, Orella, Wolf Butte, Five Points, Roundtop, and Horn 7.5′ Quadrangles), Sioux and Dawes Counties, Nebraska—Part 1: Inventory: USDA Forest Service/University of Nebraska–Lincoln (Challenge Cost-Share Agreement No. 02-07-93-011), 81 p.

LaGarry, H. E., and Hunt, R. M., Jr., 1995, 1994 Paleontological resource survey of the Oglala National Grassland (Sherrill Hills, Story, Story NE, Montrose, Orella, Wolf Butte, Five Points, Roundtop, and Horn 7.5′ Quadrangles), Sioux and Dawes Counties, Nebraska: USDA Forest Service/University of Nebraska–Lincoln (Challenge Cost-Share Agreement No. 02-07-94-015), 64 p.

LaGarry, H. E., and LaGarry, L. A., 1997, Geology of the Montrose, Orella, Wolf Butte, Roundtop, and Horn 7.5′ USGS Quadrangles, Sioux and Dawes Counties, Nebraska: University of Nebraska–Lincoln Conservation and Survey Division Open-File Report 48, 159 p.

LaGarry, H. E., Wells, W. B., and Terry, D. O., Jr., 1996, A giant notostracan(?) from the Brule Formation (Oligocene) at Toadstool Park, Sioux County, Nebraska: Geological Society of America Abstracts with Programs, v. 28, no. 4, p. A14.

Martin, L. D., 1994, Cenozoic climatic history from a biological perspective, *in* Dort, W., ed., Nebraska Academy of Sciences TER-QUA Symposium Series, v. 2, p. 39–56.

Meehan, T. J., 1994, Sediment analysis of the middle Whitney Member: climatic implications for the Upper Oligocene of northwestern Nebraska, *in* Dort, W., ed., Nebraska Academy of Sciences TER-QUA Symposium Series, v. 2, p. 57–88.

Meek, F. B., and Hayden, F. V., 1857, Descriptions of new species and genera of fossils, collected by Dr. F. V. Hayden in Nebraska Territory; with some remarks on the Tertiary and Cretaceous formations of the northwest, and the parallelism of the latter with those of other portions of the United States and Territories: Proceedings of the Academy of Natural Science of Philadelphia for 1857, p. 117–148.

Meek, F. B., and Hayden, F. V., 1861, Descriptions of new Lower Silurian (primordial), Jurassic, Cretaceous, and Tertiary fossils, collected in Nebraska, by the exploring expedition under the command of Capt. Wm. F. Raynolds, U.S. topographical engineer; with some remarks on the rocks from which they were obtained: Proceedings of the Academy of Natural Science of Philadelphia, v. 13, p. 415–447.

NACSN (North American Commission on Stratigraphic Nomenclature), 1983, North American Stratigraphic Code: American Association of Petroleum Geologists Bulletin, v. 67, p. 841–875.

Nicknish, J. M., and Macdonald, J. R., 1962, Basal Miocene ash in White River Badlands, South Dakota: American Association of Petroleum Geologists Bulletin, v. 46, p. 685–690.

Nixon, D. A., 1991, The presence of mammalian and avian trackways in the Oligocene Toadstool Park Channel sandstones in northern Sioux County, Nebraska: *in* Proceedings, 101st Annual Meeting, Nebraska Academy of Science, p. 60–61.

Prothero, D. R., and Swisher, C. C., III, 1992, Magnetostratigraphy and geochronology of the terrestrial Eocene-Oligocene transition in North America, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 46–73.

Retallack, G. J., 1983, Late Eocene and Oligocene paleosols from Badlands National Park, South Dakota: Geological Society of America Special Paper 193, 82 p.

Schaikjer, E. M., 1935, Contributions to the stratigraphy and paleontology of the Goshen Hole area, Wyoming, part III, A new basal Oligocene formation: Bulletin of the Museum of Comparative Zoology, v. 76, no. 3, p. 71–93.

Schultz, C. B., 1938, The Miocene of western Nebraska: American Journal of Science, v. 35, p. 441–444.

Schultz, C. B., 1941, The pipey concretions of the Arikaree: University of Nebraska State Museum Bulletin, v. 2, p. 69–82.

Schultz, C. B., and Falkenbach, C. H., 1968, The phylogeny of the oreodonts. Parts 1–2: Bulletin of the American Museum of Natural History, v. 39, 498 p.

Schultz, C. B. and Stout, T. M., 1945, Pleistocene loess deposits of Nebraska: American Journal of Science, v. 243, no. 5, p. 231–244.

Schultz, C. B., and Stout, T. M., 1948, Pleistocene mammals and terraces in the Great Plains: Geological Society of America Bulletin, v. 59, no. 6, p. 553–588.

Schultz, C. B., and Stout, T. M., 1955, Classification of Oligocene sediments in Nebraska: Bulletin of the University of Nebraska State Museum, v. 4, p. 17–52.

Schultz, C. B., and Stout, T. M., 1977, Drought and the model of a Quaternary terrace cycle: Nebraska Academy of Science Transactions, v. 4, p. 191–201.

Schultz, C. B., and Stout, T. M., 1980, Ancient soils and climatic changes in the central Great Plains: Nebraska Academy of Science Transactions, v. 8, p. 187–205.

Schultz, C. B., Tanner, L. G., and Harvey, C., 1955, Paleosols of the Oligocene of Nebraska: Bulletin of the University of Nebraska State Museum, v. 4, p. 1–16.

Schwartman, K. J., 1989, A geologic study of the bioturbated lower Whitney ash bed (Oligocene), near Chimney Rock Historic Site [M.S. thesis]: Lincoln, University of Nebraska, 82 p.

Singler, C. R., and Picard, M. D., 1980, Stratigraphic review of Oligocene beds in northern Great Plains: Wyoming Geological Association Earth Science Bulletin, v. 13, no. 1, p. 1–18.

Singler, C. R., and Picard, M. D., 1981, Paleosols in the Oligocene of northwest

Nebraska: University of Wyoming Contributions to Geology, v. 20, no. 1, p. 57–68.

Souders, V. L., and Swinehart, J. B., 1996, Cenozoic loesses of Nebraska: Proceedings, 106th Annual Meeting, Nebraska Academy of Sciences, p. 49.

Stout, T. M., 1971, Quaternary classification and sedimentary model; *in* Union internationale pour l'étude du Quaternaire (INQUA), VIIIe Congrès: Paris, 1969, Études sur le Quaternaire dans le Monde, v. 2, p. 790.

Stout, T. M., 1973, Revised classification of the Quaternary and Medial to Late Tertiary, based on the standard for the Great Plains (USA) and intercontinental correlations; *in* 9th Congress, International Union for Quaternary Research: Christchurch, New Zealand, INQUA, p. 347–348.

Stout, T. M., 1977, Frontiers of Nebraska geology: Nebraska Academy of Science Transactions, v. 4, p. 117–120.

Stout, T. M., 1978, The comparative method in stratigraphy: the beginning and end of an Ice Age: Nebraska Academy of Science Transactions, v. 6, p. 1–18.

Swinehart, J. B., 1995, Surface clarification of outcrop confusion: a history of the Brown Siltstone (White River Group) of Nebraska: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A89.

Swinehart, J. B., and Diffendal, R. F., 1995, Geologic map of Morrill County, Nebraska: U.S. Geological Survey Miscellaneous Investigations Series Map I–2496, scale 1:62,500.

Swinehart, J. B., and Diffendal, R. F., 1997, Geologic map of the Scottsbluff 1° × 2° quadrangle, Nebraska and Colorado: U.S. Geological Survey, Miscellaneous Investigations Series Map I–2545, scale 1:250,000.

Swinehart, J. B., Souders, V. L., DeGraw, H. M., and Diffendal, R. F., Jr., 1985, Cenozoic paleogeography of western Nebraska, *in* Flores, R. M., and Kaplan, S. S., eds., Cenozoic paleogeography of west-central United States: Rocky Mountain Paleogeography Symposium 3, Rocky Mountain Section: Tulsa, Society of Economic Paleontologists and Mineralogists, p. 209–229.

Swisher, C. C., III, and Prothero, D. D., 1990, Single-crystal $^{40}AR/^{39}AR$ dating of the Eocene-Oligocene transition in North America: Science, v. 249, p. 760–762.

Tedford, R. H., Swinehart, J., Prothero, D. R., Swisher, C. C., III, King, S. A., and Tierney, T. E., 1996, The Whitneyan-Arikareean transition in the High Plains, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, U.K., Cambridge University Press, p. 312–334.

Teichert, C., 1958, Concepts of facies: American Association of Petroleum Geologists Bulletin, v. 42, no. 11, p. 2718–2744.

Terry, D. O., Jr., and LaGarry, H. E., 1994, Lithostratigraphic correlation and provisional redefinition of the Eocene-Oligocene White River Group, Nebraska and South Dakota: Geological Society of America Abstracts with Programs, v. 26, no. 7, p. A249.

Terry, D. O., Jr., and LaGarry, H. E., 1995, Provisional revision and redescription of the upper part of the Chadron Formation in northwestern Nebraska: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A89.

Terry D. O., Jr., LaGarry, H. E., and Wells, W. B., 1995, The White River Group revisited: vertebrate trackways, ecosystems, and lithostratigraphic revision, redefinition, and redescription, *in* Flowerday, C. A., ed., Geologic Field Trips in Nebraska and Adjacent Parts of Kansas and South Dakota: University of Nebraska–Lincoln Conservation and Survey Division Guidebook 10, p. 43–57.

Tychsen, P. C., 1954, A sedimentation study of the Brule Formation in northwestern Nebraska [Ph.D. thesis]: Lincoln, University of Nebraska, 189 p.

Vondra, C. F., 1960, Stratigraphy of the Chadron Formation in northwestern Nebraska: Compass, v. 37, no. 2, p.73–90.

Vondra, C. F., Schultz, C. B., and Stout, T. M., 1969, New members of the Gering Formation (Miocene) in western Nebraska: Nebraska Geological Survey Paper, v. 18, p. 1–18.

Wells, J. W., 1947, Provisional paleoecological analysis of the Devonian rocks of the Columbus region: Ohio Journal of Science, v. 47, p. 119–126.

Wells, W. B., 1994, Paleoenvironmental and paleoclimatic implications of multiple cut and fill paleovalley sequences within Oligocene sediments of northwest Nebraska: *in* Proceedings, 104th Annual Meeting, Nebraska Academy of Science, p. 54.

Wells, W. B., Terry, D. O., and Nixon, D. A., 1994, Multiple cut and fill paleovalley sequences within Oligocene sediments (White River Group) of northwestern Nebraska: Geological Society of America Abstracts with Programs, v. 26, no. 7, p. A492.

Wells, W. B., Terry, D. O., and LaGarry, H. E., 1995, Stratigraphic implications of a fluvial origin for the "nodular zones," Brule Formation (Orella Member), northwestern Nebraska: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A95.

Wheeler, H. E., and Mallory, S. V., 1956, Factors in lithostratigraphy: American Association of Petroleum Geologists Bulletin, v. 40, no. 11, p. 2711–2723.

Wood, H. E., Chaney, R. W., Clark, J., Colbert, E. H., Jepsen, G. L., Reeside, J. B., Jr., and Stock, C., 1941, Nomenclature and correlation of the North American continental Tertiary: Geological Society of America Bulletin, v. 52, p. 1–48.

MANUSCRIPT ACCEPTED BY THE SOCIETY DECEMBER 5, 1997

Printed in U.S.A.

Geological Society of America
Special Paper 325
1998

Episodes of carbonate deposition in a siliciclastic-dominated fluvial sequence, Eocene-Oligocene White River Group, South Dakota and Nebraska

James E. Evans
Department of Geology, Bowling Green State University, Bowling Green, Ohio 43403
Linda C. Welzenbach
Public Programs Sector, National Museum of Natural History, Washington, D.C. 20560

ABSTRACT

At least four lacustrine limestone beds occur in the upper part of the Chadron Formation in South Dakota and Nebraska. Most limestones have limited areal extent, and represent short-lived features. Limestones typically form shallowing-upward sequences up to 2.1 m thick, consisting of (from bottom to top): marl, algal stromatolite, algal mat and oncolites, undulose-bedded limestone, and charophyte-rich massive limestone. Fossils in these limestones include Centrachidae and Amiidae fish, turtle, *Cypris* sp. ostracodes, *Galba* sp. and *Planorbis* sp. gastropods, *Lampsisus*(?) sp. bivalves, calcified roots of aquatic macrophytes, charophyte stems and gyrogonites, and algal filament ghosts. Trace fossils include *Taenidium*, *Skolithos*, *Paleophycus*, *Planolites*, and vertebrate (e.g., bird) tracks. Other carbonate features in the Chadron Formation include pedogenic calcrete, and paleogroundwater deposits (tufa, travertine, and nonpedogenic calcrete pinnacles interpreted as vertical conduits for groundwater discharge). The two most notable facts about carbonate in the Chadron Formation are that: (1) the lacustrine limestones are pure carbonate (siliciclastic content <1%), indicating that these depositional systems were isolated or protected from the adjacent, overbank-dominated, siliciclastic fluvial systems, and (2) the lacustrine limestones and paleogroundwater deposits are found adjacent to structural features (e.g., faults, fracture systems, and lineaments), suggesting that the deposits were localized at structurally controlled places of paleogroundwater discharge. Finally, it is notable that all of the lacustrine limestones (and most of the tufas) are in the Chadron Formation to the exclusion of the underlying Chamberlain Pass Formation or overlying Brule Formation. This suggests a unique paleohydrologic event in this region during the Late Eocene. It is suggested that unroofing of the Black Hills uplift led to the initiation of a regional paleogroundwater flow system during the deposition of the Chadron Formation. Such regional groundwater system was apparently not as effective a mechanism for carbonate deposition in the overlying Brule Formation, probably as a consequence of the paleoclimatic and paleohydrologic changes (increased aridity, decreased recharge) that occurred in this region during the Oligocene.

Evans, J. E., and Welzenbach, L. C., 1998, Episodes of carbonate deposition in a siliciclastic-dominated fluvial sequence, Eocene-Oligocene White River Group, South Dakota and Nebraska, *in* Terry, D. O., Jr., LaGarry, H. E., and Hunt, R. M., Jr., eds., Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America): Boulder, Colorado, Geological Society of America Special Paper 325.

INTRODUCTION

Fresh-water limestone can be an important constituent of many nonmarine depositional environments. In North America, significant fresh-water limestones formed in the central Rocky Mountains during the Paleogene as portions of Lakes Flagstaff, Gosiute, and Uinta (Bradley, 1964; Ryder et al., 1976; Stanley and Collinson, 1979; Flores, 1981). Fresh-water limestone can also be associated with fluvial channel deposits (Ordonez and Garcia del Cura, 1983), with flood-plain sequences (Demicco et al., 1987; Sanz et al., 1995), with coal sequences (Cabrera and Saez, 1987; Gierlowski-Kordesch et al., 1991), and with paleosols (Freytet and Plaziat, 1982; Platt, 1989). Information obtained from lacustrine limestones and related carbonate deposits has helped to clarify regional paleogeography, paleoclimatic changes, and tectonic controls on sedimentation (Beaumont, 1979; Yuretich, 1989).

This chapter examines the stratigraphy, lithology, and paleontology of fresh-water limestones in the Late Eocene Chadron Formation, White River Group, in Nebraska and South Dakota. Previous workers have interpreted Chadron Formation carbonate as fresh-water limestone (Wanless, 1923), cryptalgal limestone (Clark et al., 1967), marl (Harksen and Macdonald, 1969a), and pedogenic calcrete (Retallack, 1983). In fact, all of these lithologies can be present at different outcrops of the Chadron Formation, in a predictable sequence of beds.

Although lacustrine limestone is volumetrically a minor constituent in the White River Group, the presence of pure carbonate in a siliciclastic, overbank-dominated, fluvial sequence provides important insights regarding tectonics, paleoclimate, and sediment supply. The goal of this chapter is to describe the vertical and lateral stratigraphic relationships of the limestones and their related carbonate facies. The evidence will show that deposition of freshwater limestone and spatially related, paleogroundwater deposits (tufa, travertine, and nonpedogenic calcrete pinnacles) in the Chadron Formation represents a unique paleohydrologic episode in the history of the White River Group.

GEOLOGIC SETTING

Regional geology

The Precambrian basement in western South Dakota and northwest Nebraska is overlain by a thick succession of Paleozoic and Mesozoic sedimentary rocks, ending with the Late Cretaceous Pierre Shale. In this region, early Cenozoic weathering led to the formation of the Yellow Mounds Paleosol Series on exposed marine sedimentary rocks (Retallack, 1983). Subsequent Paleogene uplift and unroofing of the Black Hills, as well as incision and backfilling of adjacent areas, led to the accumulation of as much as 250 m of nonmarine sediments of the Eocene-Oligocene White River Group in the study area (Clark et al., 1967; Stanley, 1976; Evans and Terry, 1994; Terry et al., 1995). The significance of base-level changes on the deposition of the basal White River Group is discussed elsewhere (Evans and Terry, 1994).

White River Group

The study area is located in southwestern South Dakota and northwestern Nebraska (Fig. 1). In this region, the White River Group consists of the Chamberlain Pass, Chadron, and Brule Formations. The Chamberlain Pass Formation is a newly recognized unit as much as 16 m thick, which is composed of multistory channel deposits and overbank mudstones (Terry, 1991; Evans and Terry, 1994). The overbank mudstones have been modified by pedogenesis to form the Interior and Weta Paleosols (Retallack, 1983; Terry and Evans, 1994).

The overlying Chadron Formation ranges considerably in thickness, due to backfilling of paleovalleys and blanketing of a surface with paleorelief. In South Dakota the unit fills and then overtops a 53-m-deep paleovalley that is incised through the

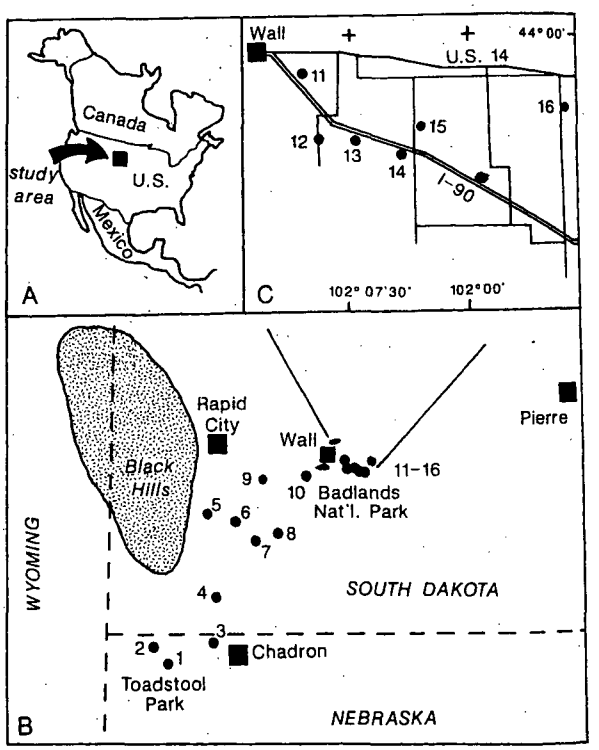


Figure 1. Location map. A; Location of the study area in central United States. B; Locations of measured sections. C; Detailed locations of sections at locations 11–14 (Bloom Basin Limestone Bed). 1 = Sand Creek Road; 2 = Whitehead Creek; 3 = Rattlesnake Butte; 4 = Limestone Butte; 5 = Fairburn; 6 = French Creek; 7 = Red Shirt Table; 8 = Cottonwood Pass; 9 = Bear Creek; 10 = Railroad Buttes; 11 = Mile post 114; 12 = Bloom Basin; 13 = Mile post 118; 14 = Mile post 121; 15 = Street Hill; and 16 = Walker Hill (see text for legal coordinates).

Chamberlain Pass Formation, Yellow Mounds Paleosol, and into the Pierre Shale (Harksen and Macdonald, 1969a; Evans and Terry, 1994). Inside the paleovalley, the Chadron Formation is subdivided into the Ahearn, Crazy Johnson, and Peanut Peak Members (Clark, 1954). Outside the paleovalley the Chadron Formation is finer grained. It was undifferentiated by Clark et al. (1967) but has recently been assigned to the Peanut Peak Member (Terry et al., 1995). In Nebraska, portions of the Chadron Formation also backfill paleovalleys, reaching a maximum thickness of about 100 m (Terry et al., 1995). These Nebraska paleovalleys are filled and overtopped by the Peanut Peak Member, which is overlain by the newly recognized Big Cottonwood Creek Member (Terry and LaGarry, this volume). The Chadron Formation is interpreted as a fluvial depositional environment with channels and overbank sequences (Wanless, 1922, 1923; Clark et al., 1967), overprinted by paleosols (Retallack, 1983).

Local limestone beds in the upper Chadron Formation have been recognized by previous workers (Wanless, 1923; Clark et al., 1967; Harksen and Macdonald, 1969b) but have not been studied in detail until recently (Welzenbach, 1992). The age of these lacustrine limestones is known in part. Detailed stratigraphic studies (this chapter) show that Harksen and Macdonald (1969b) were correct in assigning the limestones to the Chadron Formation, but, in contrast to practice by some (e.g., Terry et al., 1995), at no place do any of the limestones form the depositional upper contact of the Chadron Formation. Revised radiometric dating using $^{40}Ar/^{39}Ar$ has shown that the Chadron Formation is late Eocene to early Oligocene (about 37 to 33.7 Ma) in age (Swisher and Prothero, 1990). Magnetostratigraphy at Red Shirt Table and Cottonwood Pass (Fig. 1) indicate that the limestone beds at these locations correlate to the upper part of Chron 13R, close to the boundary between C13R and C13N at about 33.7 Ma (Prothero and Swisher, 1992; Whittlesey and Prothero, 1995).

The contact between the Chadron Formation and the overlying Brule Formation varies considerably. In the Big Badlands of South Dakota, the contact is marked by either a prominent calcareous claystone breccia (Harksen and Macdonald, 1969b) or the paleosol that contains this calcareous horizon (Retallack, 1983). In either case, the contact is marked by the change from "haystack" weathering of the Chadron Formation to "tread and riser" weathering of the Brule Formation (Harksen and Macdonald, 1969b). In Nebraska, the contact between the Chadron and Brule Formations is intertonguing, except where fluvial channels of the Orella Member cut into the Big Cottonwood Creek Member (Terry et al., 1995).

The overlying Brule Formation is about 137 m thick in South Dakota, and about 177 m thick in Nebraska. The unit consists of fluvial and eolian deposits with interstratified air-fall tuffs (Wanless, 1923; Clark et al., 1967). In South Dakota, the unit is split into the Scenic and Poleslide Members (Bump, 1956) on the basis of calcareous nodule horizons, color, and weathering features. The unit is overprinted by the Conata, Gleska, Zisa, Ohaka, Ogi, Wisangie, and Pinnacles Paleosol Series (Retallack, 1983). In Nebraska, the unit is split into the Orella Member (fluvial), Whitney Member (eolian), and "brown siltstone" member (fluvial and eolian) (Terry et al., 1995; LaGarry, this volume).

UPPER CHADRON FORMATION LIMESTONE BEDS

Facies analysis

Lacustrine limestone. Limestones in the Chadron Formation are bench-forming, tan or light gray, massive to crudely bedded or cryptalgal laminated carbonates that weather white or reddish. The limestones are either nearly pure calcite or ferroan calcite. Trace fossils, mudcracks, and macrofossils including gastropods, bivalves, fish, turtle, and charophyte stems are common. Microkarst features include solution-enhanced fractures that were filled by siliciclastic material from overlying deposits, typically angular granule- and sand-size quartz grains in a reddish clay matrix.

There are five carbonate lithofacies in the Chadron Formation lacustrine limestones (Table 1). Each represents the common field occurrence of combinations of six carbonate microfacies, as described and interpreted from thin-section study: (1) micrite, (2) clotted micrite, (3) ostracode-charophyte wackestone, (4) oncolite-peloidal packstone, (5) gastropod-charophyte-vertebrate floatstone, and (6) algal bindstone.

Charophyte-rich massive limestones (lithofacies A) mostly consist of massive ostracode-charophyte wackestone (Fig. 2). Commonly, there is a thin capping layer of fossiliferous floatstone (accumulations of charophyte stems, gastropods, fish bones, or turtle plastron plates). The upper contact is often mudcracked and rooted and has surface traces, including bird tracks. Burrows are common, and include *Taenidium*, *Skolithos*, and *Paleophycus*. The massive aspect of the limestone is probably due to rooting and bioturbation. This lithofacies is interpreted as the rooted charophyte "meadow" of the littoral zone (e.g., Platt and Wright, 1991).

Undulose bedded limestones (lithofacies B) typically pinch and swell in thickness, and are internally massive or crudely bedded. Bedding is due to accumulations of ostracodes, gastropods, bivalve shell debris, or fish bones (Fig. 3). Oncolites, coated grains, and burrows are common features. This lithofacies is interpreted as the wave-reworked portion of the littoral zone, similar to low-energy "bench" margins described elsewhere (Murphy and Wilkinson, 1980).

Laminated limestone (lithofacies C) is primarily flat or wrinkled algal bindstones (rhythmites composed of alternating dense micrite, clotted micrite, and peloidal packstone), oncolites, and coated grains (Fig. 4). Fossils are relatively rare, and consist of ostracodes and gastropods found in the peloidal packstones, and rare horizons of articulated fish fossils. Burrows are rare, and surface traces (*Planolites*) are uncommon. This lithofacies is interpreted as algal mat from the profundal zone, basinward of the zone of wave reworking, in a stratified lake (Dean and Fouch, 1983; Platt and Wright, 1991).

Stromatolitic limestone (lithofacies D) consists of laterally linked hemispheroids as tall as 10 cm. This lithofacies differs

TABLE 1. CARBONATE LITHOFACIES IN THE CHADRON FORMATION

Lithofacies and Lithology	Fossils*	Structures	Interpretation
A. Charophyte-rich, Massive Limestone			
Ostracode-charophyte wackestone	o, c	b, st, mc	Littoral zone
Fossiliferous floatstone	o, c, g, f	b, st, mc	
Micrite and clotted micrite	Rare	b, st, mc	
B. Undulose Bedded Limestone			
Ostrocode-charophyte wackestone	o, c	b, st	Zone of wave reworking
Peloidal packstone	o, g, b, f	b, st	
Fossiliferous floatstone	o, c, g; f, t	b, st, fl	
C. Laminated Limestone			
Algal bindstone	Rare	st, sc(?)	Algal mat
Oncolite-peloidal packstone	o, g, f	st	
Micrite and clotted micrite	Rare	st	
D. Stromatolite			
Algal bindstone	Rare	st	Algal bioherm
Peloidal packstone	o, g	st	
E. Marl			
Peloidal packstone	o	Rare	Deep lake
Dense micrite	Rare	Rare	
F. Tufa			
Pisolite-intraclast wackestone	Rare v, p	Rare	Spring deposits
Dense and porous micrite	Rare	Rare	
G. Travertine			
Dense and porous micrite	Rare	Rare	Spring deposits
Feather dendrites	Rare	Rare	
H. Nonpedogenic Calcrete			
All microfacies	Rare	p, f	Spring deposits

*b = bivalves; c = charophyte stems or gyrogonites; f = fish bones; g = gastropods; o = ostracodes, p = terrestrial plants (as roots, root molds, or calcified stems), v = terrestrial vertebrate bones. Rare = fossils are uncommon.

b = burrows; f = fluid escape structures; fl = fossil lag horizon; mc = mudcracks; p = primary sedimentary structures (bedding and cross bedding); r = terrestrial plant roots; sc(?) = possible synaeresis cracks; st = surface traces. Rare = structures are uncommon.

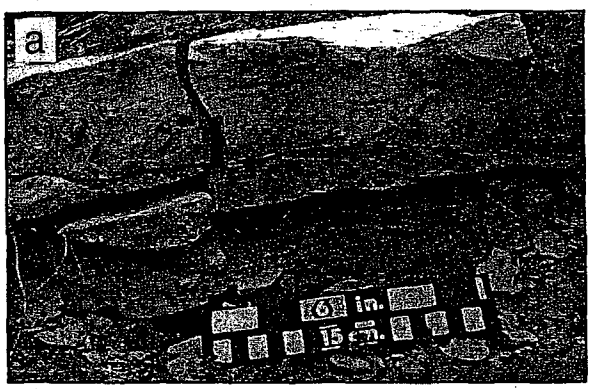


Figure 2a. Photographs of charophyte-rich massive limestone. Field photograph of lithofacies A (scale, 15 cm).

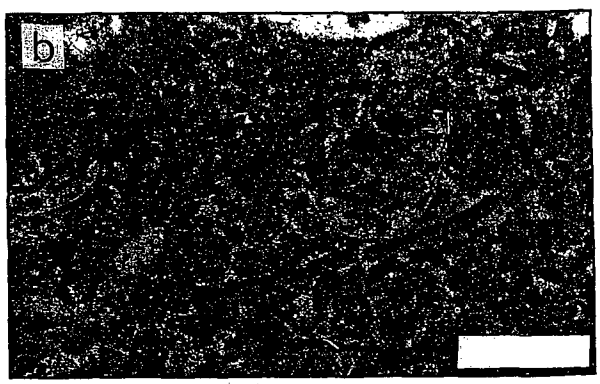


Figure 2b. Photomicrograph showing peloidal, clotted fabric and bioclasts (plane polarized light; scale bar, 50 μ; magnification ×31).

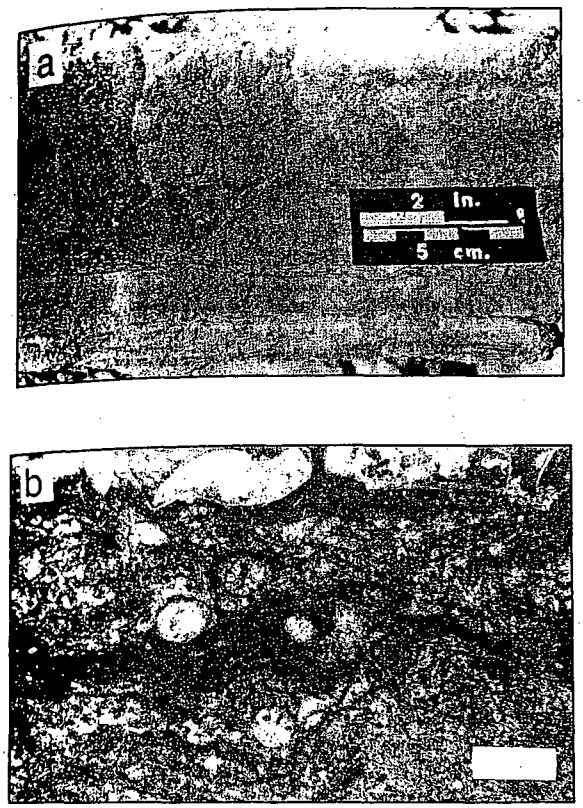


Figure 3. Photographs of undulose bedded limestone. a; Field photograph of lithofacies B (middle), overlying laminated limestone at base, with charophyte-floatstone (lithofacies A) at top (scale, 5 cm). b; Photomicrograph showing scours and bioclast accumulations (plane polarized light; scale bar, 50 μ; magnification ×31).

only in external morphology from the laminated algal mats described above, and the two lithofacies are often interstratified. This lithofacies is interpreted as deeper water algal bioherms, similar to but of smaller scale than those described from Lake Tanganyika (Cohen and Thouin, 1987).

Marl (lithofacies E) consist of alternating gray, red, and black thinly bedded carbonate mudstones and interstratified laminated dense micrite or peloidal packstone (Fig. 5). Fossils and trace fossils are generally absent. This lithofacies is interpreted as deeper lake deposits (Kelts and Hsu, 1978).

Paleogroundwater deposits. Paleogroundwater deposits in the Chadron Formation include tufas, travertines, and nonpedogenic calcrete pinnacles. These three lithofacies consist in turn of five carbonate and siliceous microfacies: (1) pisolite-intraclast wackestone, (2) dense micrite, (3) porous micrite, (4) feather dendrites, and (5) siliceous carbonate (representing a spectrum from almost pure calcrete to almost pure silcrete).

Tufas (lithofacies F) are found as localized "mounds" of offlapping sheets composed of pisolite-intraclast wackestone or porous micrite that are interstratified with siliciclastic mudstones (Fig. 6a). These sheets are laterally discontinuous on the scale of tens of meters. Each sheet pinches and swells in thickness, to fill paleotopography (Fig. 6b). Individual carbonate sheets are partly to completely draped by siliciclastic mudstones. The carbonates contain pisolites, concentrically banded mud intraclasts ("mudballs"), rootcasts, terrestrial vertebrate bones, and calcified terrestrial plant roots and stems. Internally, the tufas consist of poorly organized, alternating horizons of pisolite-intraclast wackestone, porous micrite, and dense micrite with a fenestral fabric. This lithofacies is interpreted as subaerial spring deposits (Hardie et al., 1978). These tufas represent the deposits of paleogroundwater discharge because they are interstratified with the Chadron Formation (within any tufa mound complex, each sheet of carbonate is partly to completely draped by siliciclastic mudstone).

Travertines (lithofacies G) are found in a few locations,

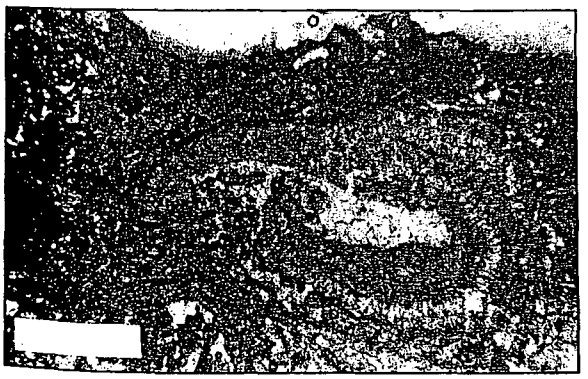


Figure 4. Photomicrograph of laminated limestone (lithofacies C), showing oncolite and coated grains (plane polarized light; scale bar, 1 mm; magnification ×31).

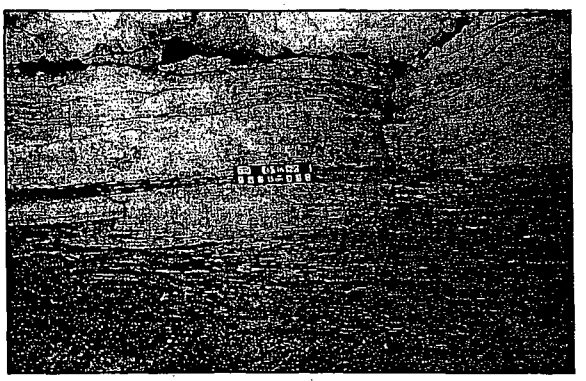


Figure 5. Photograph of two shallowing-upward sequences from gray, black, and red marl at the base (lithofacies E) to stromatolite (lithofacies D) and algal-laminated limestone (lithofacies C) at the top (scale, 15 cm).

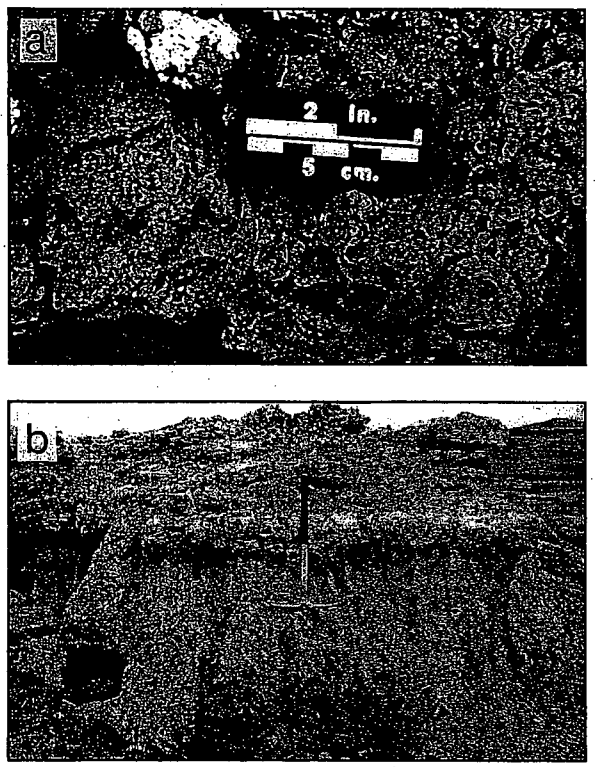


Figure 6. Photographs of tufas interbedded with the Chadron Formation (lithofacies F). a; Pisolite-intraclast wackestone (scale, 5 cm). b; Sequence of offlapping sheets with partial to complete mudstone drapes (pick handle, 65 cm).

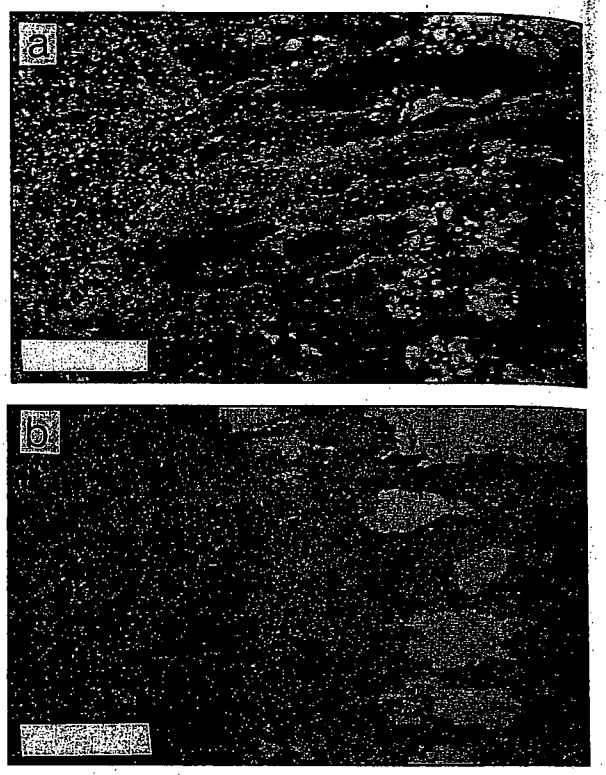


Figure 7. Photomicrographs of banded travertine interbedded with the Chadron Formation (lithofacies F), showing feather dendrites (a, crossed polars; b, plane polarized light; scale bar, 100 μ; magnification ×31).

associated with tufas, and resembling tufa mounds in external form. Their features are similar to classic descriptions of travertines (Scholl, 1960; Scholl and Taft, 1964; Hardie et al., 1978). In the Chadron Formation, travertines are internally banded carbonates that consist of alternating laminae of dense micrite, porous micrite, and "feather dendrites" (Fig. 7), which are characteristic features of hot spring deposits (Jones and Renaut, 1995). As with the tufas, the age of these travertines is clearly contemporaneous with the Chadron Formation because they are interstratified on a fine scale. In other words, the travertines are paleogroundwater deposits.

These tufa and travertine mounds are often associated with subvertical, nonpedogenic calcrete pinnacles or "tufa tubes" (lithofacies H) that stand out in erosional relief from surrounding strata due to enhanced cementation within the pinnacle. The cements within the pinnacles include porous and dense micrite, mosaic calcite, and one or several episodes of silica cementation (see Evans and Terry, 1994: one of these pinnacles is shown in their Fig. 7b and discussed as a silcrete). Internally, the pinnacles can show concentric banding and fluid escape features that crosscut primary sedimentary structures. These pinnacles are usually found within a sandy parent material (usually a fluvial channel deposit) directly below a tufa or travertine mound. Such deposits are interpreted as the conduits for vertical paleogroundwater flow toward springs (points of discharge). These features are called "pinnacles" solely because of their tendency to stand in erosional relief today, they are not directly comparable to the constructional algal-tufa pinnacles observed in modern lakes (e.g., Scholl, 1960; Scholl and Taft, 1964), although it is possible that they were the "roots" of such constructional pinnacles.

Paleontology

Fish. Specimens identified from the Chadron Formation include Centrarchidae (sunfish) and Amiidae (bowfin). Fossil material occurs as both fully and partially articulated fish skeletons, as scattered bones, as a complete impression of a fish, and as individual scales (Fig. 8). The best articulated specimens come from deeper water deposits (laminated limestone and marl). The Centrarchidae specimens are probably *Plioplarchus* (G. R. Smith, written communication, 1992), although S. Foss (written communication, 1995) has suggested the specimens might represent a new genus.

Figure 8. Photograph of articulated Centrarchidae skeleton found in algal bindstone (scale, 1 cm).

Figure 9. Photograph of the silicified plastron plate of an aquatic turtle (scale, 3 cm).

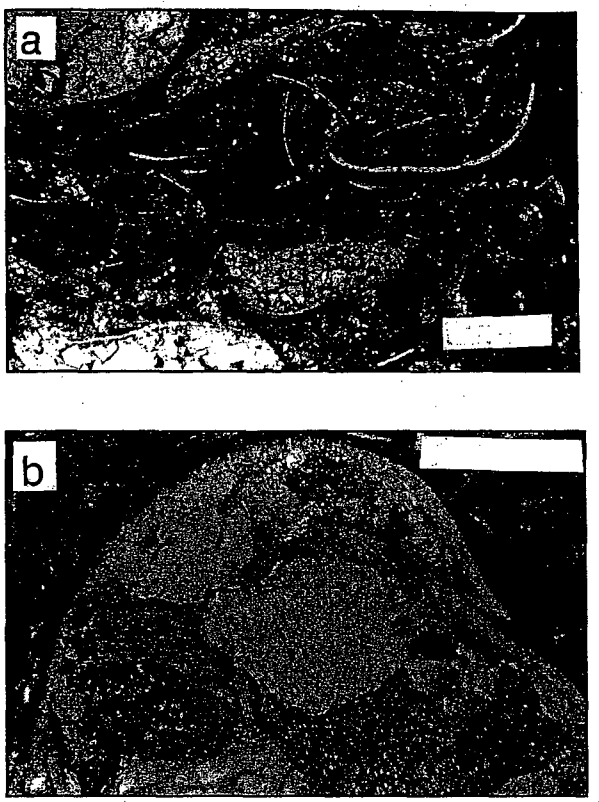


Figure 10. Photographs of ostracodes. a; Accumulation of disarticulated bioclasts (plane polarized light; scale bar, 50 μ; magnification ×31). b; Articulated specimen with an overlapping valve, and moldic porosity with mosaic calcite infilling (crossed polars; scale bar, 100 μ; magnification ×125).

They are typically 3–5 cm long (Fig. 8). One fully articulated specimen has at least 5 or more dorsal and anal spines, suggesting that it is either *Plioplarchus whitnei* (9 dorsal and 5 anal spines) or *P. sexspinosis* (10 dorsal and 6 anal spines) (Smith and Miller, 1985).

Amiidae specimens are identified on the basis of the long dorsal fin, caudal skeleton, and other features. S. Foss (unpublished data, 1995) has undertaken a systematic study of Amiidae specimens collected from the White River Group. Although sample localities for many older specimens were poorly recorded, at least several specimens were collected by John Clark (donated to the Field Museum of Natural History) from the Chadron Formation limestones described in this chapter (location 12 in Fig. 1). In addition, new Amiidae specimens have been recovered from location 14 (Fig. 1) during 1995 (S. Foss, written communication, 1995).

Turtles. Fossilized turtle remains in the Chadron limestones are in the form of silicified plastron plates (Fig. 9). These were relatively small aquatic turtles, whose remains are found in shallow-water deposits (charophyte-rich massive limestones and undulose bedded limestones). The average size of complete plastron plates from these deposits is about 5 cm, suggesting that the size of the organism was about 40 cm long and 20 cm wide (G. R. Smith, written communication, 1992). There is not sufficient material to classify the turtles from the Chadron limestones at this time. Turtle fossils recovered from deposits of similar age and environment have included specimens of *Emys* and *Trionyx* (Daley, 1972; Taylor, 1978).

Ostracodes. Articulated or disarticulated ostracodes are abundant in the Chadron Formation limestones. All specimens were determined to be of a single taxon, *Cypris* sp. (R. M. Forester, written communication, 1991; A. Smith, written communication, 1991). These have ovate carapaces, with smooth and unornamented valves, one valve being slightly longer and overlapping, and each valve being between 0.5 and 3 mm long (Fig. 10). The valve microstructure is a homogeneous, fine-grained prismatic spar.

In the Chadron limestones, ostracodes are found in the shallower water deposits (charophyte-rich massive limestone and undulose bedded limestone). This is in accord with their known habitat as benthic organisms in oxygenated, perennial,

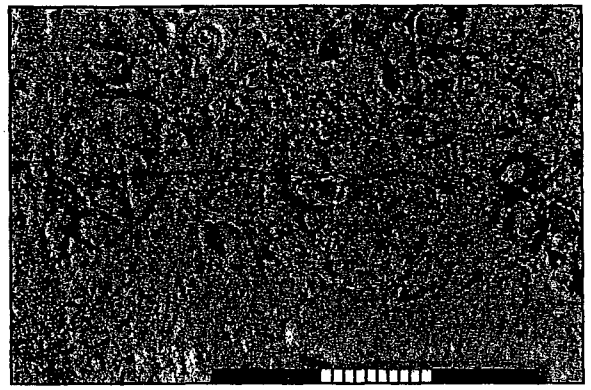


Figure 11. Photograph of casts and molds of pulmonate gastropods *Galba longiscata* and *Planorbis* sp. on bedding surface (scale, 3 cm).

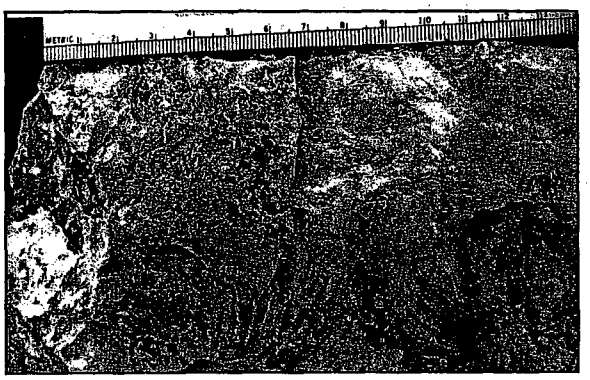


Figure 12. Photograph of disarticulated bivalve shells, identified as *Lampsisus*(?), in undulose-bedded limestone (scale, 13 cm).

relatively calm water (Forester, 1991). Some horizons of articulated ostracodes are found in life position on a surface with desiccation cracks (A. Smith, written communication, 1991). These life assemblages may be the result of sudden lowering of the water surface, emergence, desiccation, and death of the ostracode benthic community.

Gastropods. Chadron limestones contain numerous specimens of two types of pulmonate gastropods: *Galba longiscata* and *Planorbis* sp. (Fig. 11). The two taxa are found together in littoral zone deposits (charophyte-rich massive limestone) as shell lags in the undulose bedded limestone or as scattered shells in deeper water deposits (laminated limestones).

Galba longiscata was recovered as molds or impressions in limestone. Microscopically, individual specimens are moldic with a micrite rim and are commonly infilled with mosaic calcite. Geopetal structures are common. Molds of shells are 3–7 mm in length, with three or four convex whorls, thin shells, and simple apertures.

Planorbis sp. was also recovered as molds or impressions in limestone. Shells are up to 4 mm long, sinistral and planispiral, with a wide umbilicus. Microscopically, shells were identifiable from spar-filled voids, with no preservation of wall structure. Thin micrite rims surround each shell, with partial to complete infillings of mosaic calcite. Geopetal structures are common.

Pulmonate gastropods are known from a wide variety of lake-margin settings from Paleogene to Recent (e.g., Daley, 1972; Taylor, 1978). In modern lakes, pulmonate gastropods feed on algal films encrusting vegetation, and occupy a marginal position where emergent vegetation permits them to move out of the water to breathe (Daley, 1972). Modern representatives of the two taxa are rarely found basinward of the limit of rooted vegetation, except as shell lags.

Bivalves. At least two taxa of bivalves are present, but one taxon has been recovered only as small shell fragments of a bivalve that has a smooth and unornamented outer shell. Microscopically, these specimens display thicker walls than ostracodes, having a lamellar, prismatic, and commonly foliated microstructure, which may be homogeneous to granular in texture. It has not been possible to identify this taxon.

The second taxon is tentatively classified as *Lampsisus(?)* based on comparisons to published descriptions (Cook and Mansfield, 1933; Gries and Bishop, 1966). These bivalves are found as collections of numerous disarticulated valves within undulose bedded limestone (Fig. 12). As noted by Gries and Bishop (1966), these bivalves are often accompanied by gastropods and oncolites. Preservation is in the form of casts and molds, and fossils have the same light gray color as the surrounding matrix; thus, unweathered specimens are difficult to discern. On weathering, specimens turn reddish-brown in color. Prominent growth lines are visible on internal casts (Gries and Bishop, 1966)(Fig. 12). The maximum dimensions observed for these bivalves is ~10 cm long × 6 cm wide.

Vascular plants. Calcified roots and stems of aquatic macrophytes have been found associated with shallow-water limestones and with spring deposits (Fig. 13). In each instance, the plant material was apparently coated with micrite, and the resulting mold was infilled and cemented after the original organic material decayed. It has not been possible to identify the original material.

Charophytes. Stems, peloids, and gyrogonites (fruiting bodies of oogonia) are abundant in shallow-water portions of the Chadron limestones. Stems are broken vegetative structures that may be straight or branched, up to about 1 cm in length (Fig. 14a,b). Stems typically display micrite envelopes, moldic porosity, and partial to complete infilling of mosaic calcite. Geopetal structures can be common (Fig. 14a). Cross-sectional views of vegetative structures resemble a small cluster of circles or flattened circles, each representing micrite envelopes around individual charophyte stems.

Gyrogonites are spherical or flattened bodies about 50 μ in diameter. The outer margin may appear to be smooth or may show a spiral ornamentation (Fig. 14c). Typically there is moldic porosity, infilled with mosaic calcite.

Charophytes are typically found in shallow, low-salinity,

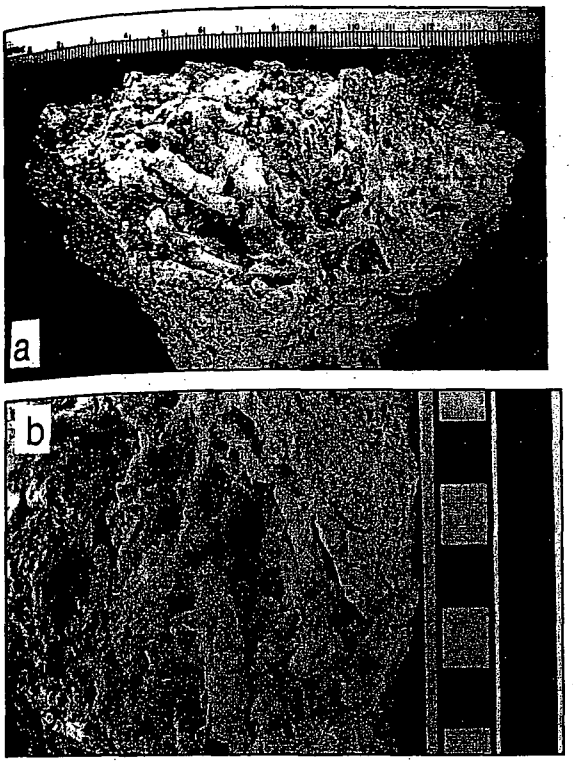


Figure 13. Photograph of calcified roots and stems of a vascular plant, found with tufas (spring deposits). a; Complete specimen (scale, 15 cm). b; Structure of rhizoliths (scale, 5 cm). Structure of rhizoliths (scale, 5 cm).

alkaline, fresh-water environments that have relatively low wave energy, where they form extensive *Chara* meadows (Wray, 1977; Flugel, 1982; Cohen and Thouin, 1987). Corroded individual stems and isolated gyrogonites indicate reworking and transport to the deepest parts of the lakes (Freytet and Plaziat, 1982).

Algae. Evidence for filamentous and nonfilamentous algae in the Chadron limestones include algal mats and algal stromatolites, oncolites and coated grains (Fig. 4), and algal filament ghosts (micritic peloids or spar-filled voids with filamentous shape). Such features are common throughout the Chadron limestones, and are especially well developed in the laminated limestone and stromatolite facies.

Trace fossils

Taenidium. These burrows are unbranched, uniform diameter, horizontal to slightly inclined tubes, with unlined and unornamented tube walls, and meniscate backfill (Fig. 15). Individual tubes are as much as 8 mm wide and 7 cm long, may intersect or interpenetrate, and show complete horizontal, inclined, and cir-

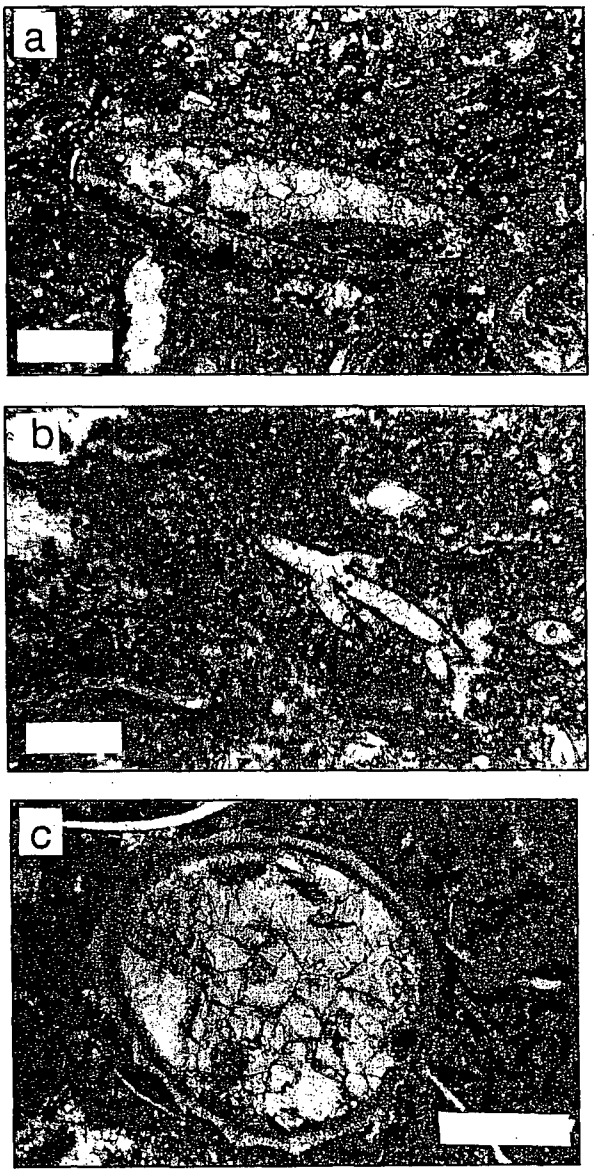


Figure 14. Photomicrographs of charophytes. a; Stems with micrite rims, mosaic calcite infilling and geopetal structure (scale, 50 μ). b; Branching stems (scale, 50 μ). c; Gyrogonite with surface (spiral) ornamentation (scale, 100 μ). (All three photomicrographs are plane polarized light, magnification ×31.)

cular terminations. Menisci are uniform crescent-shaped layers distinguished by changes in grain size. The tubes are preserved as concave and convex epireliefs. These trace fossils are identified as *Taenidium* on the basis of comparison to published studies (Ekdale et al., 1984; Squires and Advocate, 1984; D'Alessandro and Bromley, 1987). The burrows are interpreted as fodichnia (Simpson, 1975), probably of aquatic oligochaetes or insects (Chamberlain, 1975).

Skolithos. These burrows are straight, vertical, unbranched, unlined, unornamented tubes of uniform diameter. The tubes are as much as 6 mm wide and 3 cm long, and do not intersect. Individual tubes are unfilled molds, preserved as endichnia. The slight curving or "J" shape to some burrows suggests that it served as domichnia for terrestrial insects (Bromley and Asgaard, 1979; Frey and Pemberton, 1984).

Paleophycus. These burrows are cylindrical, straight to slightly curved, horizontal, lined tubes that may be ornamented or smooth (Fig. 16). The tubes are as much as 5 mm wide and 4 cm long. Most tubes are simple, but some branch or intersect. The tubes were molds that were infilled with micrite, preserved as hyporeliefs. They are interpreted as domichnia (Pemberton and Frey, 1982).

Planolites. These are epichnial surface traces, horizontal, with a slightly meandering path direction. The external boundaries of the trace are irregular, so that the trace width varies. Traces do not branch or intersect. The trackway left by *Planolites* is typically infilled by the overlying sediment, suggesting that the trackway formed in a cohesive carbonate mudstone. They are interpreted as fodichnia, probably of insects (Pemberton and Frey, 1982). In the Chadron Formation, *Planolites* is most commonly found in the laminated limestone facies.

Vertebrate tracks. Bird tracks were recovered at one locality (Fig. 17). Individual impressions showed three anterior phalanges ~2.5 cm long, and one hallux of ~1.5 cm long, giving a total foot length of ~4 cm; it is possible that the foot is webbed. Individual tracks are superimposed to form a trackway, suggestive of short stride lengths. The mud impressions are consistent with being cohesive (submerged or wet emergent substrate). The trackway is interpreted as the fodichnia of a wading bird.

Stratigraphy

Detailed stratigraphic descriptions for each site are given in the Appendix. Several general observations can be made about the stratigraphy of the Chadron Formation in the study area (Fig. 18). First, the lacustrine limestones do not form a continuous marker horizon across the area, with the exception of a relatively continuous horizon in the vicinity of Wall, South Dakota (the "Bloom Basin Limestone Bed" of Welzenbach, 1992). However, lacustrine limestone beds are concentrated toward the top of the Chadron Formation throughout the study area, suggesting a regional episode of the formation of numerous smaller carbonate ponds. It should be noted that lacustrine limestones are absent from the underlying Chamberlain Pass Formation and overlying

Figure 15. Photograph showing prominent *Taenidium* burrows with meniscate backfill in massive limestone (scale, 2 cm).

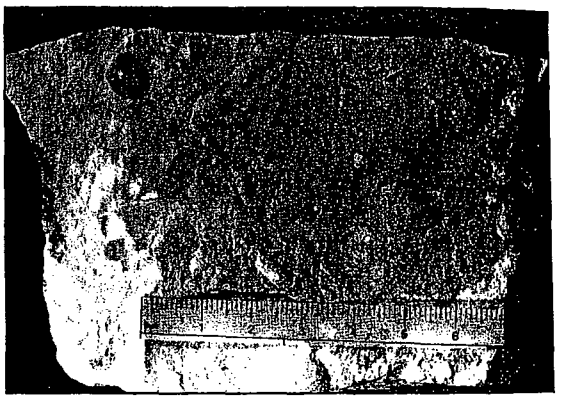


Figure 16. Photograph showing *Paleophycus* burrows in massive charophyte-rich limestone (scale, 7 cm).

Figure 17. Photograph showing vertebrate (bird) tracks in massive, charophyte-rich limestone (scale, 4 cm).

Brule Formation, bracketing this regional episode of lacustrine carbonate to late Eocene.

Second, paleogroundwater deposits (tufas, travertines, and nonpedogenic calcrete) occur throughout the Chadron Formation, and have been observed in the underlying Chamberlain Pass Formation and overlying Brule Formation. Where lacustrine limestone is found in the Chadron Formation, it is typical for paleogroundwater deposits to underlie the limestone (Fig. 18; Appendix). Tufas and travertines are commonly found as "tufa mound complexes" or as a series of thin (typically 3–40 cm thick) laterally discontinuous tufa beds that are stacked vertically, offlap one another, and are interstratified with siliciclastic mudstones. These complexes can be seen even at the scale used to construct Figure 18 and are described in more detail in the Appendix.

Third, the lacustrine limestones, tufas, and travertines are found associated with distal flood-plain deposits in the Chadron Formation. The limestones are not, for example, fluvial carbonates (i.e., they do not represent the infilling of abandoned channels). Aside from that general association, the carbonate deposits show no close association with any specific fluvial subenvironment. The only exception is that the nonpedogenic calcrete pinnacles ("tufa tubes") are best developed in sandy host materials such as fluvial channel deposits or proximal overbank deposits.

DISCUSSION

Lacustrine depositional environment

Lacustrine limestones in the Chadron Formation show a consistent facies succession of marl (lithofacies E), overlain by stromatolite (lithofacies D) and algal mat (lithofacies C), overlain by undulose bedded limestone (lithofacies B) and charophyte-rich massive limestone (lithofacies A). An example is given from the Fairburn section (Fig. 19), which shows two superimposed facies successions. At many localities, only the upper three lithofacies are present (Fig. 20). In addition, lithofacies D and C can alternate or be repetitively bedded (representing variations in morphology between algal mat and stromatolite). This succession of lithofacies is interpreted as a shallowing-upward carbonate sequence that represents lake basin filling and emergence. Similar examples have been described as low-energy, bench-type lacustrine carbonate systems (Murphy and Wilkinson, 1980, Platt and Wright, 1991).

The depth of these lakes was presumably deeper than indicated by the thickness of the carbonate basin fill. The evidence that the lakes were stratified includes the following: that burrowing organisms were restricted to the littoral zone; that deeper water facies were dominated by undisturbed cryptalgal lamination, algal laminites, and thinly bedded marls; and that articulated fish skeletons were found only in the profundal sediments. Stratification is generally a function of water temperature profiles and wind stress on the lake surface, typically requiring water depths greater than 10 m. There are several possible mechanisms to foster stratification in shallower water, one mechanism being dense vegetation surrounding small ponds (to mitigate wind stress), another possible mechanism being salinity (e.g., Ruttner, 1963). For example, meromictic conditions in Fayetteville Green Lake of New York have been attributed to discharge of saline groundwater at lake bottom springs (Ludlam, 1969; Brunskill and Harriss, 1969). Either of these mechanisms could have been a factor in the development of stratification in the small lakes and ponds of the Chadron Formation, but at this time, the paleo-depth of these lakes is not well understood.

The life span of these lakes was probably relatively short. At some localities there are as many as 50 algal laminites, which are interpreted as rhythmites that may represent seasonal or annual productivity events (Wright, 1990; Gomez Fernandez and Melendez, 1991). This suggests that individual ponds may have persisted for at least several decades. However, the lake fill successions are relatively thin, with a maximum thickness of 2.1 m. Even if low sedimentation rates were assumed, none of these lakes could have been existed for more than 10^5 yr (see following section).

The most curious aspect of these lake deposits is the lack of siliciclastic detritus in the carbonate. Each lacustrine sequence is encased in siliciclastic floodplain sequences. The absence of incorporated siliciclastic detritus, coupled with the close spatial association of limestone with underlying spring deposits, suggests that the lakes were at least partially controlled by groundwater discharge. The groundwater chemistry was clearly alkaline (given the mineralogy of the deposits) and possibly saline (given the apparent restriction of faunas to the littoral zone or overlying water column, and apparent stratification of the water column). The possible role of saline groundwater discharge at facilitating water column stratification is discussed earlier.

The fossils represent a high abundance, low-diversity assemblage with evidence of environmental stress, including the "fish-kill" horizons, and life assemblages of *Cypris* ostracodes (A. Smith, written communication, 1991). In addition to the "normal" lake filling and emergence cycle, the taphonomic evidence suggests either relatively rapid lake level drawdown and emergence (possibly due to drought), or possible overturning of the stratified water column (by an unknown mechanism).

Paleohydrology

Several lines of evidence suggest that carbonate deposition in the Chadron Formation was controlled by paleogroundwater discharge. These lines of evidence include: (1) the presence of paleogroundwater deposits (tufas, travertines, and nonpedogenic calcrete pinnacles) in the Chadron Formation, (2) the spatial relationship of lacustrine limestones and paleogroundwater deposits in the Chadron Formation, (3) the absence of siliciclastic detritus in the lacustrine limestones and apparent lack of association of the limestones to any fluvial subenvironment, (4) the geochemical evidence for groundwater input, and (5) the association of carbonate with structural features (Fig. 21).

The first three points can be dealt with briefly, as they have been discussed previously. The Chadron Formation contains tufas and travertines, both organized in mounds consisting of

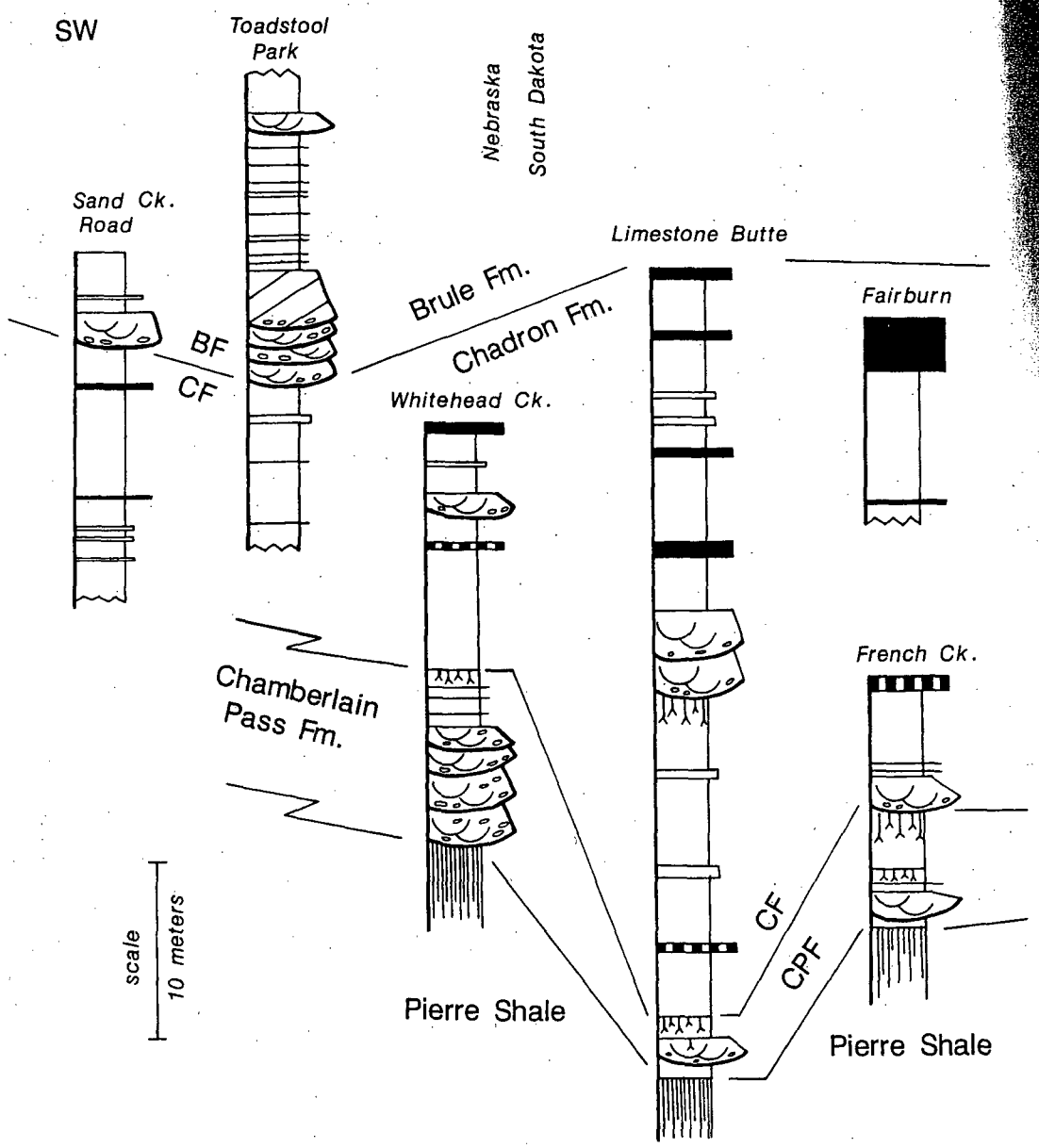
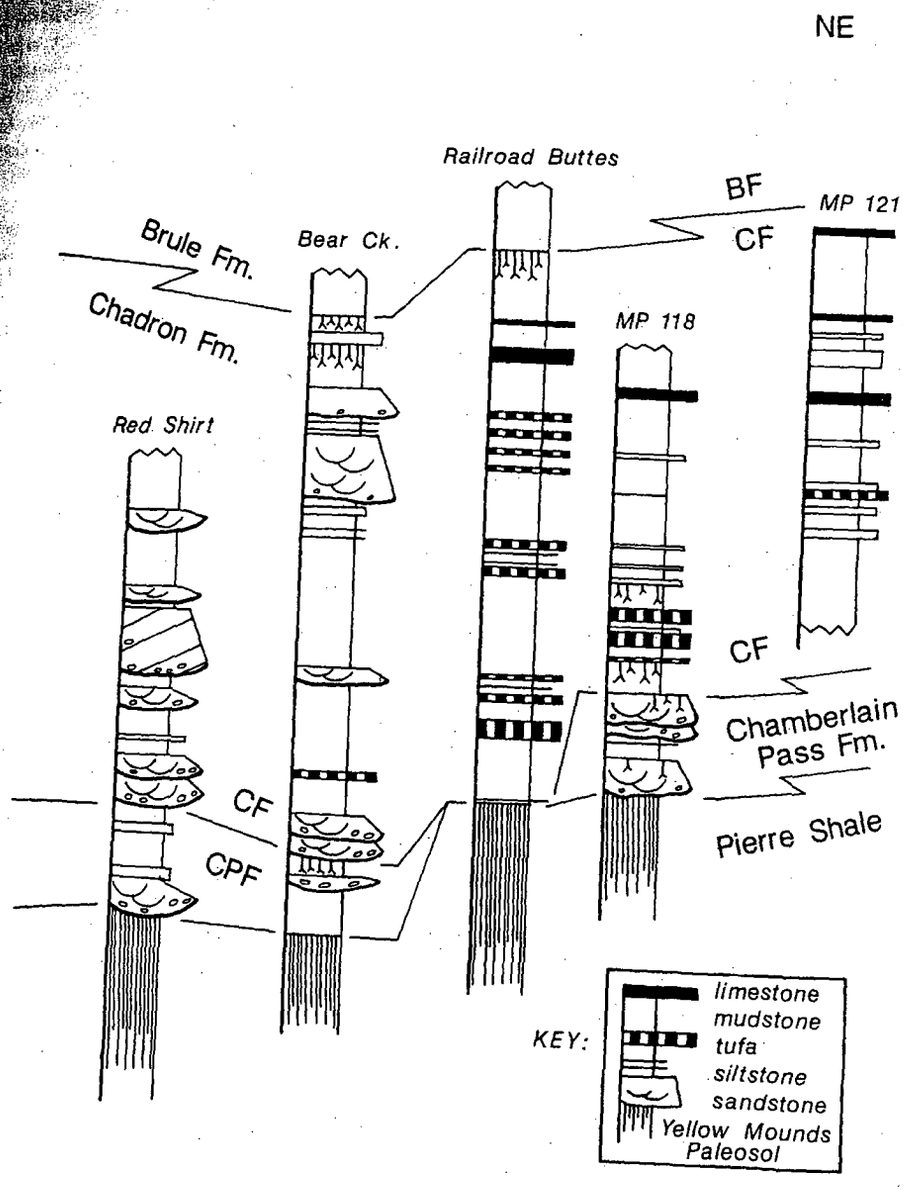


Figure 18. Stratigraphic correlation diagram for most of the sections discussed in this study. The Chamberlain Pass Formation is a new unit at the base of the White River Group (Evans and Terry, 1994). Note that while tufas, travertines, and other ground water associated deposits are found throughout the Chadron Formation, the occurrences of lacustrine limestones are concentrated toward the top of the unit.


NE
Railroad Buttes
BF
MP 121
CF
Brule Fm.
Bear Ck.
Chadron Fm.
MP 118
Red Shirt
CF
Chamberlain Pass Fm.
CF
CPF
Pierre Shale
KEY:
limestone
mudstone
tufa
siltstone
sandstone
Yellow Mounds Paleosol

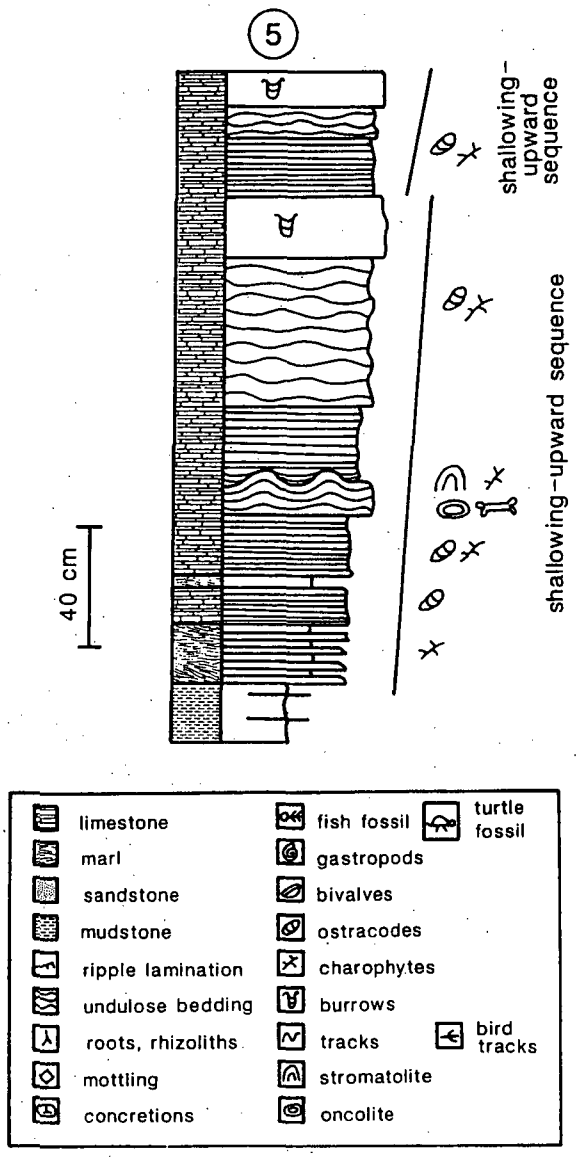


Figure 19. Detailed stratigraphic section of the lacustrine limestones found at Fairburn, South Dakota (location 5 in Fig. 1). At this location there are two shallowing-upward lacustrine sequences consisting of marl (lithofacies E), laminated limestone and stromatolite (lithofacies C and D), undulose-bedded limestone (lithofacies B), and charophyte-rich massive limestone (lithofacies A). The top sequence is lithofacies C, B, and A only.

offlapping sheets of carbonate that are partly to completely draped by siliciclastic mudstones, and incorporating rootcasts, plant debris, and vertebrate fossils, clearly indicating paleogroundwater discharge at subaerial springs into the Chadron depositional environment. These features are identical in all details to descriptions of modern groundwater features (Hardie et al., 1978). Underlying many of the tufa and travertine mounds are vertical pinnacles of nonpedogenic calcrete and calcrete that are interpreted as conduits for groundwater flow upward to the overlying springs. All of these features are spatially related to lacustrine limestones: They tend to be found in the same locations, and typically the paleogroundwater features underlie the lacustrine deposits. This, coupled with the lack of siliciclastic detritus or evidence for fluvial/deltaic features in the limestones, suggest that the lakes developed at locations of paleogroundwater discharge, were groundwater-fed, and were isolated from fluvial systems.

The geochemistry of the deposits also supports this hypothesis. Features in the limestones are consistent with an alkaline and saline water chemistry, as discussed earlier. In addition, a stable isotope study of the White River Group by Lander (1991) found evidence for mixing of meteoric and groundwater fluids in the geochemistry of "calcretes." Lander's results can only be applied in general terms. It was an excellent study in all respects except that Lander did not recognize the presence of tufas and travertines. He therefore did not distinguish among the stable isotope composition of the different forms of carbonate present in the White River Group: pedogenic calcrete, nonpedogenic calcrete, lacustrine limestone, tufa, and travertine. Lander's study also preceded several significant changes in stratigraphic nomenclature (Evans and Terry, 1994; Terry and LaGarry, this volume). A re-investigation of stable isotope geochemistry of the different forms of carbonate in the White River Group is ongoing (J. E. Evans, unpublished data, 1998).

Finally, strong evidence for paleogroundwater discharge as the control over carbonate deposition in the Chadron Formation is the spatial relationship of carbonate and structural features (Fig. 21). The lacustrine limestones, tufas, and travertines are found adjacent to faults, axial-plane fracture systems, and LANDSAT lineaments. An analysis of the lineaments by Shurr et al. (1996) concluded that they represented tectonic (and not modern geomorphic) features. Sedimentary evidence suggests that these faults were active syndepositionally with the White River Group. The Chamberlain Pass Formation was deposited in a basin between the Sage Creek and Pine Ridge fault zones (Evans and Terry, 1994). Stratigraphic offsets and reversals in sense of displacement have been documented within the Brule Formation near the Pine Ridge fault zone (Simpson, 1985). At Slim Buttes (northwest South Dakota) faults cut the White River Group but not the overlying Arikaree Group (Shurr et al., 1996). Clearly, some of these structures were in existence prior to or during the deposition of the Chadron Formation.

The association of carbonate deposits of probable groundwater origin with structural features can be explained if the structures served as discontinuities that facilitated groundwater flow. Seeps and springs along faults are common occurrences (e.g., Fetter, 1994). In addition, in many cases the drainage disruption associated with active faulting can result in the formation of ponds and lakes.

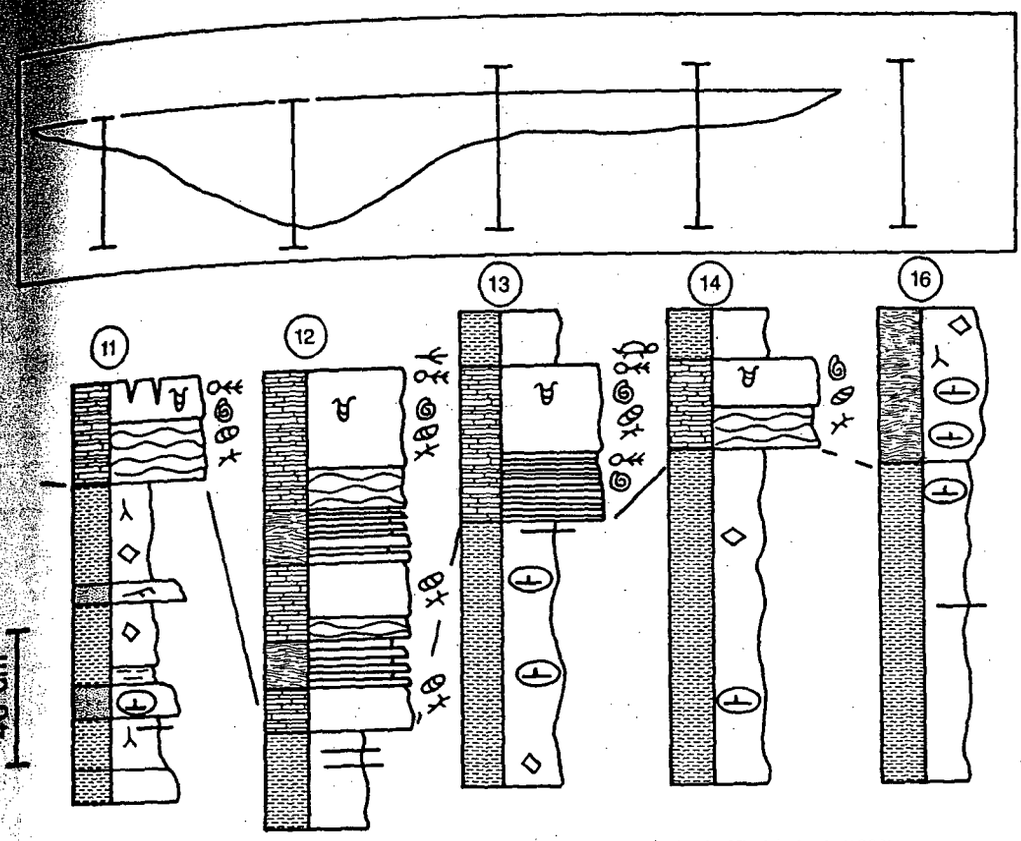


Figure 20. Detailed stratigraphic sections of the Bloom Basin Limestone Bed of Welzenbach (1992), from locations 11–14, 16 (see Fig. 1 for locations). This bed may represent one or more phases of a larger but shallower lake than shown in Figure 19. Most of the lacustrine limestones found in the Bloom Basin Limestone Bed are the upper parts of the shallowing-upward sequence (lithofacies A and B). Symbols are the same as in Figure 19.

Source of regional groundwater system

Lacustrine limestones in the White River Group are found only in the upper part of the Chadron Formation, while most of the related paleogroundwater deposits (tufas, travertines, and nonpedogenic calcretes) are also restricted to the Chadron Formation (Fig. 18). The evidence has been presented that the lacustrine limestones were deposited in groundwater-fed lakes. It is therefore proposed that a unique paleohydrologic event, the initiation of a regional groundwater discharge system that supplied water and dissolved constituents to these springs and freshwater lakes, occurred in this region during the late Eocene. It is further hypothesized that the cause of initiating this regional groundwater flow system was the unroofing of the Black Hills uplift, which created recharge and discharge areas and a regional hydraulic gradient, and also dissolved carbonate source rocks in the Black Hills. Finally, it is hypothesized that carbonate deposition became less important in the Brule Formation because of paleoclimatic change in this region during the early Oligocene.

Several lines of evidence support of these ideas. First, there is independent evidence from provenance data (sandstone petrofacies and paleocurrents) of the Chamberlain Pass Formation and Chadron Formation, suggesting that the main pulse of unroofing of the Black Hills uplift, including exposure of core rock lithologies, occurred during the late Eocene (Evans, 1996). Second, such unroofing would also expose and cause the dissolution of Paleozoic and Mesozoic carbonates in the Black Hills. Solution features in these older carbonate strata have been documented at the Tertiary erosion surface in parts of the Black Hills (Redden, 1996). Third, the modern groundwater system is centripetal from recharge zones in the Black Hills, with high flow rates associated with paleokarst features in the limestones (Rahn and Gries, 1973; Klemp, 1996; Hayes, 1996).

The proposed paleogroundwater flow system is shown in

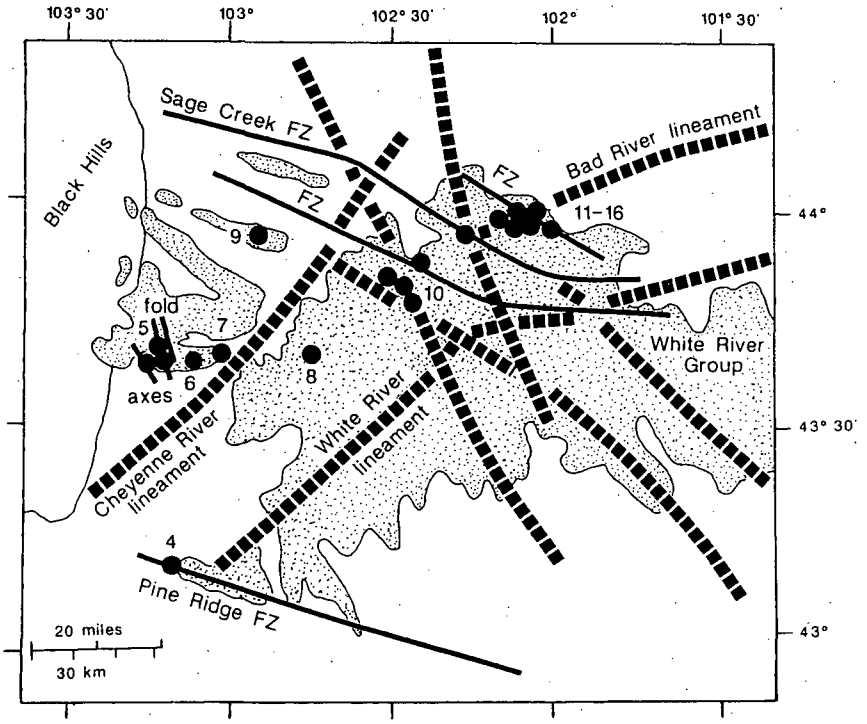


Figure 21. Map showing the spatial relationships between faults (solid lines), lineaments (dashed lines), and outcrops of tufas and limestones in the Chadron Formation (dots) for South Dakota (diagram modified from Shurr et al., 1996). Within the Chadron Formation, tufas, travertines, and lacustrine carbonates of possible groundwater origin appear to be localized adjacent to structural features. There is independent evidence that at least some of these structures were pre- or syn-depositional (see text). One unifying explanation would be that structural control of paleogroundwater discharge determined the location of carbonate within the Chadron Formation.

Figures 22 and 23. Figure 22 is a cross section showing schematically that uplift and unroofing of the Black Hills generated hydraulic gradients, recharge and discharge areas within the region. The chief difference between this representation and the modern groundwater flow system (under a significantly more arid climate) is that today recharge can occur locally at the distal end of the flow system by leakage across the aquicludes (mostly Cretaceous shales) through fractures (Rahn and Hayes, 1996).

Paleogroundwater discharge in the Chadron depositional environment may have been facilitated by two mechanisms. The first was discharge along fractures and faults, as discussed earlier. Most of the carbonate in the Chadron Formation is spatially related to such structural features (Figs. 21, 23). A second mechanism may explain those occurrences that are not located on structures. This mechanism is flow into and along the axis of paleovalleys (Fig. 23). Such paleovalleys were incised into bedrock and backfilled with permeable sandy fluvial sediments (Evans and Terry, 1994; Terry et al., 1995). Buried bedrock valleys can be significant conduits for groundwater flow (Fetter, 1994).

In summary, the evidence suggests that carbonate deposition in the Chadron Formation signaled the initiation of a regional paleogroundwater flow system, which in turn was a consequence of late Eocene unroofing of the Black Hills uplift. The question arises why lacustrine limestones are absent (and tufas and travertines much less abundant) in the overlying Brule Formation. One obvious explanation is the effect of Oligocene paleoclimatic change in this region on the paleoflow systems. Increasing aridity across the Eocene-Oligocene boundary is well documented, based on independent evidence such as paleosols (Retallack, 1983, 1992), land-snail faunas (Evanoff et al., 1992), and reptile and amphibian faunas (Hutchison, 1992).

Lakes are well understood as indicators of fluctuations in precipitation and evaporation regimes (e.g., Van Houten, 1964; Picard and High, 1972; Olsen, 1986). For the Chadron Formation, the absence of siliciclastic detritus in the lacustrine limestones indicates that flood-waters did not enter these lakes. Input

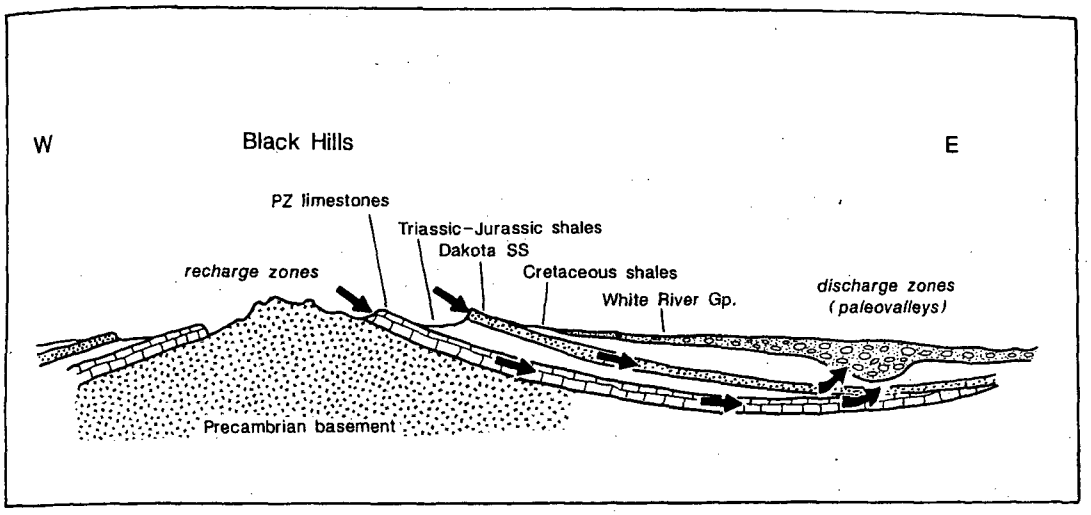


Figure 22. Schematic cross section of the regional groundwater system in the Black Hills and Badlands of South Dakota, showing the location of recharge zones (erosional unroofing of the Black Hills uplift) and discharge zones (upward transport via lithologic discontinuities such as paleovalleys, or structural discontinuities such as faults and fractures). The modern regional groundwater system, under a more arid climate, is structurally similar to what is shown here, but recharge is occurring, rather than discharge, through these distal discontinuities (diagram modified from Rahn and Hayes, 1996).

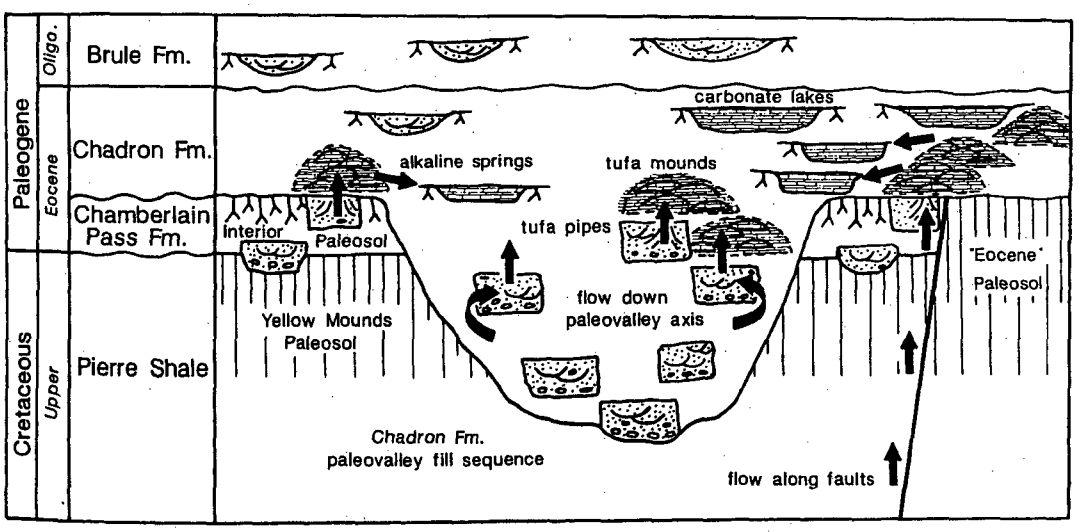


Figure 23. Interpretive lithostratigraphic diagram showing structural and lithologic discontinuities that may have served to localize paleogroundwater discharge during deposition of the Chadron Formation.

was therefore direct precipitation and groundwater discharge, and output was either evaporation or a combination of evaporation and discharge into outflowing streams. All of these variables are susceptible to paleoclimatic fluctuations. Given the structure of the regional paleogroundwater flow system described above, decreased precipitation and increased evaporation could reduce flow rates and possibly could reverse hydraulic gradients in portions of the paleoflow system (discussed earlier).

Isolation of lacustrine systems

The coexistence (within close proximity) of pure carbonate and pure siliciclastic depositional systems requires conditions of either essentially vertical accretion or the presence of a barrier to flood-waters. The evidence from the Chadron Formation is not conclusive about the relative importance of these isolation mechanisms.

Vertical accretion could be accomplished under conditions of matching sediment accumulation rates and subsidence rates. Cojan (1993) has described one such combined carbonate and siliciclastic fluvial depositional system from a setting characterized by equal but relatively low sediment accumulation and subsidence rates. The average sediment accumulation rate of the entire Chadron Formation can be shown to have been relatively low. The Chadron Formation is a diachronous unit and is not synonymous with the duration of the Chadronian North American Land-Mammal Age, but for the purposes of this simple calculation, such age control is used. The Chadronian-Duchesnean contact has an $^{40}Ar/^{39}Ar$ age of about 37.2 Ma (an average of the Bracks Rhyolite at 36.7 ± 0.1 Ma and the Buckshot Ignimbrite at 37.8 ± 0.2 Ma, according to Prothero and Swisher, 1992). Tuff 5, being 6.4 m below the Chadronian-Orellan contact, has an $^{40}Ar/^{39}Ar$ age of 33.59 ± 0.021 Ma (Obradovich et al., 1995). Thus the unit represents deposition over a time interval of about 3.6 m.y. The maximum thickness of the Chadron Formation is about 55 m in South Dakota (Harksen and Macdonald, 1969a) and about 100 m thick in northwest Nebraska, giving a sediment accumulation rate of between 1.53 and 2.78 cm/k.y. These relatively low sedimentation rates are in accord with the extensive pedogenesis within the Chadron Formation (Retallack, 1983). Sediment accumulation rates for lacustrine limestones are typically higher than this range of values, suggesting that vertical accretion of the adjacent carbonate and siliciclastic depositional environments could have occurred.

Barriers to flood waters could include the following scenarios: (1) fluvial channels were incised while carbonate ponds formed on alluvial terraces, topographically above the reach of flood-waters; (2) lateral channel migration, levee crevasses, and/or channel avulsions were asymmetric (i.e., flooding was typically on one side of the channel complex); (3) carbonate ponds were protected from flood-waters by the presence of constructional topography in the form of an alluvial ridge; or (4) some combination of the above. These barrier hypotheses are possible, but it should be noted that the carbonate ponds are interstratified with fluvial deposits, so any such barriers were temporary (conversely, the lakes were short-lived). Evidence for incision and terrace development is not strong in Chadron fluvial facies (J. E. Evans, unpublished data, 1994–95). Finally, it should also be noted that Chadron fluvial channels are east- or southeast-directed across the region where the carbonate ponds formed, so it is not clear how asymmetric channel migration would have been an effective mechanism for barrier formation.

In summary, it is likely that the coexistence of pure carbonate and pure siliciclastic depositional systems in close spatial proximity was the result of vertical accretion (relatively equivalent sediment accumulation rates and subsidence rates). For the Chadron Formation, sediment accumulation rates for the siliciclastic facies was relatively low (1–3 cm/k.y.), which could have been matched by carbonate facies. Within this context, local barriers to flooding may have permitted limestone to accumulate. The formation and removal of such sedimentologic barriers may have been a response to autocyclic processes of the fluvial system, or due to short-term paleoclimatic fluctuations.

SUMMARY AND CONCLUSIONS

In western South Dakota and northwestern Nebraska, the White River Group contains lacustrine limestones (restricted to the upper portion of the Chadron Formation) and paleogroundwater deposits including tufas, travertines, and nonpedogenic calcrete pinnacles (mostly restricted to the Chadron Formation). These deposits signal a unique paleohydrologic event in this region.

The lacustrine deposits resemble the deposits of modern low-energy, bench-type carbonate lakes. These deposits contain a low-diversity, high-abundance assemblage of fish, turtles, ostracodes, gastropods, bivalves, aquatic macrophytes, charophytes, and filamentous algae (stromatolites, oncolites, and algal mats). Evidence has been presented that the individual lakes were relatively small (i.e., carbonate ponds), and relatively short-lived. Many of the ponds were relatively shallow (above wave-base), but some show evidence of stratification of the water column and deposition of laminites in deeper water. Some of the lake deposits show evidence for terminal emergence and drying up. The lacustrine limestones contain <1% siliciclastic detritus and show no obvious relationship to fluvial channel or proximal overbank environments, indicating that they are not fluvial carbonates (infilling of fluvial channels). Lake waters were alkaline and probably saline. The spatial relationship of lacustrine limestones to paleogroundwater deposits, the low siliciclastic content of the limestones, the evidence that they were not fluvial carbonates, and the alkaline and saline water chemistry strongly suggest that the lakes were groundwater fed.

The paleogroundwater deposits resemble modern tufa and travertine mounds, containing pisolites, intraclasts, vertebrate bones, rootcasts, and other features. Underlying many of these are vertical "pinnacles" of calcrete and silcrete that show evidence of being conduits for paleogroundwater flow. In addition to their morphology and internal composition, evidence for the origin of these paleogroundwater features includes spatial relationship to structural features such as faults and fracture systems that would have facilitated groundwater flow, and the geochemical and stable isotope evidence for mixing of meteoric and groundwater fluids.

Within the White River Group, all of the lacustrine limestones are found in the upper Chadron Formation, and most of the paleogroundwater deposits are within the Chadron Formation. It is hypothesized that this can be explained as the initiation of a regional groundwater flow system during the late Eocene. Such system could in turn be explained as a consequence of unroofing of the Black Hills uplift, establishment of regional hydraulic gradients and flow systems, development of recharge and discharge areas, and dissolution of Paleozoic and Mesozoic carbonate units in the Black Hills. Paleogroundwater discharge into the Chadron depositional environment appears to have been facilitated by groundwater flow along faults and fractures (pre-

or syn-depositional features) and/or flow through buried bedrock valleys (paleovalleys). The reason why there are fewer carbonate features in the overlying Brule Formation is not conclusive, although strong evidence exists for paleoclimatic change and increasing aridity across the Eocene-Oligocene boundary; this would have affected the paleohydrologic balance in the region and may thus have affected paleogroundwater flow rates and possibly even flow directions.

Finally, the presence within the Chadron Formation of two mutually exclusive depositional systems (a pure carbonate lacustrine depositional system, and an overbank-dominated, siliciclastic fluvial depositional system) requires explanation. Either there were barriers to the extent of flood-waters, which isolated the carbonate depositional systems, or vertical accretion of both depositional systems occurred (sediment accumulation rates matched subsidence rates). There is evidence in support of both ideas, but further research is warranted.

ACKNOWLEDGMENTS

This study represents work during the interval of 1989–1996. We are grateful to Dennis Terry (former student of J. E. E.), who introduced us to the area and first called our attention to these deposits. L. C. W. did the stratigraphy of a portion of this study area ("Bloom Basin Limestone Bed") as well as much of the paleontologic descriptions for a master's thesis at Bowling Green State University. J. E. E. identified the paleogroundwater deposits, facies assemblages, and extended the study area.

We have benefited from advice and paleontologic confirmations from Allison Smith, Richard M. Forester, Richard Hoare, Gerald R. Smith, Scott Foss, and John Howe. We acknowledge a special debt of thanks to Charles F. Kahle for work with carbonate microfacies. The comments of four reviewers—Andrew Cohen, Carl Drummond, Hannan LaGarry, and Dennis O. Terry, Jr.—have greatly improved this manuscript. We also wish to thank the staffs of Badlands National Park and Buffalo Gap National Grasslands for logistical support and landowner contacts. This research was supported by several grants from the Faculty Research Committee of Bowling Green State University.

APPENDIX: SITE LOCATIONS AND STRATIGRAPHIC DESCRIPTIONS

Sand Creek Road section. The section at Sand Creek Road, Sioux County, Nebraska (SW1/4 sec. 24, T. 33 N., R. 53 W.) contains two limestones in the upper part of the Chadron Formation. The base of the section is >5.5 m of Chadron Formation gray mudstone and siltstone. The overlying limestone is 8-cm-thick, gray to reddish-brown on weathered surfaces, and siliceous (both fabric- and nonfabric-selective chalcedony as replacement). The limestone consists of 4 cm of stromatolitic limestone (lithofacies D), overlain by 1 cm of laminated limestone (lithofacies C), and 3 cm of massive limestone, with *Taenidium* and *Skolithos* burrows (lithofacies A).

The upper limestone overlies 6.2 m of gray mudstone. The limestone is a siliceous bench-forming unit 20 cm thick. It consists of 6 cm of laminated limestone (lithofacies C), overlain by 12 cm of undulose bedded limestone with oriented ostracode fragments (lithofacies B), and 2 cm of charophyte-rich massive limestone with *Taenidium* burrows (lithofacies A).

The uppermost Chadron Formation at this locality is 2.45 m of gray mudstone, incised into by a fluvial channel sequence of the Orella Member of the Brule Formation (2.43 m of broadly lenticular, trough and planar cross-bedded, pebbly, medium- to very coarse grained, tan, argillaceous feldspathic arenite with *Skolithos* burrows, overlain by 1.15 m of laminated and ripple-laminated, fine-grained, tan feldspathic arenite).

Whitehead Creek section. The section at Whitehead Creek, Sioux County, Nebraska (SW1/4 sec. 36, T. 34 N., R. 4 E.) contains a tufa mound and overlying limestone. The base of the section is in the Cretaceous Pierre Shale and superimposed Yellow Mounds Paleosol. These are overlain by a multistory fluvial channel sequence (9.85 m of trough cross-bedded and massive, pebbly, coarse- to very coarse grained, ferruginous, argillaceous quartz arenite with pebble stringers and mud intraclasts near the base) and a mud-filled channel deposit (2.8 m of mottled red and green mudstone, with roots) of the Chamberlain Pass Formation.

The overlying Chadron Formation consists of 7.2 m of gray mudstone overlain by a tufa mound sequence. The tufa mound consists of several offlapping sheets of arenaceous pisolite-intraclast wackestone with clay-lined root molds, and pore-filling chalcedony cement, each about 5 cm thick (lithofacies F).

The limestone overlies a flood-plain and channel-margin sequence (2.2 m of gray mudstone; 54 cm of cross-bedded and rippled, fine-grained, feldspathic arenite; 1.75 m of gray mudstone; 20 cm of trough cross-bedded, pebbly, very coarse grained siliceous feldspathic arenite with intraclasts; and 1.5 m of gray mudstone). The limestone is a gray, indurated, caprock unit that weathers white. It consists of 10 cm of ostracode-charophyte wackestone and clotted micrite with *Skolithos* burrows (lithofacies A).

Rattlesnake Butte section. One limestone was found at Rattlesnake Butte, Dawes County, Nebraska (NE1/2 sec. 9, T. 34 N., R. 49 W.). The base of the section is in the Cretaceous Pierre Shale and superimposed Yellow Mounds Paleosol. The overlying Chamberlain Pass Formation consists of a multistory channel sequence (2.5 m of trough cross-bedded and massive, pebbly, coarse- to very coarse grained, ferruginous, argillaceous quartz arenite with pebble stringers) overlain by a flood-plain sequence (1.5 m of red and green mudstone with glaebules and roots).

The overlying Chadron Formation consists of 2.0 m of gray mudstone with a limestone caprock. The limestone consists of 8 cm of ostracode-charophyte wackestone and clotted micrite (lithofacies A).

Limestone Butte section. The section at Limestone Butte, near Oelrichs, Fall River County, South Dakota (SW1/4 sec. 16, T. 10 S., R. 8 E.) contains one tufa mound and four limestones. The base of the section is in the Cretaceous Pierre Shale and superimposed Yellow Mounds Paleosol. The overlying Chamberlain Pass Formation consists of a channel sequence (1.2 m of trough cross-bedded and massive, pebbly, coarse- to very coarse grained, ferruginous, argillaceous quartz arenite with pebble stringers), and overlying floodplain sequence (1.25 m of red and green mudstone with glaebules and roots).

The rest of the section is in the Chadron Formation. The tufa mound sequence overlies a flood-plain sequence (7.8 m of gray mudstones, 0.75 m of prominent white siltstone, and 5.25 m of gray mudstone). The tufas consist of offlapping sheets of arenaceous, pisolitic–intraclast wackestone about 20 cm thick, with "mudball" intraclasts (lithofacies F).

The lowest limestone overlies a flood-plain sequence (5.25 m of gray mudstone and 1.3 m of dark gray mudstone with roots) that has been incised up to 2 m by a channel sequence (4.2 m of trough cross-bedded and massive, fine-grained, feldspathic arenite with mud intra-

clasts), and overlain by 3.2 m of gray mudstone. The limestone bed is a siliceous bench-forming unit 57 cm thick. It consists of 36 cm of algal bindstone at the base (lithofacies C), overlain by 12 cm of undulose-bedded limestone with shell accumulations and oncolites (lithofacies B), overlain by 9 cm of ostracode-charophyte wackestone (lithofacies A).

The second limestone overlies 5.55 m of gray mudstone. The limestone is a siliceous bench-forming unit 27 cm thick. It consists of 5 cm of stromatolitic limestone with small, laterally linked hemispheroids (lithofacies D), overlain by 5 cm of laminated limestone (lithofacies C), overlain by 11 cm of undulose bedded limestone with oncolites (lithofacies B), overlain by 6 cm of charophyte-rich massive limestone with charophyte floatstone laminae at the top (lithofacies A).

The third limestone overlies a flood-plain sequence (1.1 m of gray mudstone, 18 cm of siltstone with pedogenic nodules, 0.5 m of mottled red and green mudstone, 20 cm of siltstone with pedogenic nodules, and 5.15 m of mottled red and green mudstone). The limestone is a bench-forming unit 40 cm thick. It consists of massive, rooted, ostracode-charophyte wackestone and clotted micrite (lithofacies A).

The upper limestone overlies 2.2 m of mottled red and green mudstone. The limestone is siliceous and forms the 50-cm-thick mesa caprock unit. It consists of massive, rooted, ostracode-charophyte wackestone and clotted micrite with *Taenidium* burrows (lithofacies A).

Fairburn section. Two limestones are found at an exposure near the town of Fairburn, Custer County, South Dakota (NE1/4 sec. 18, T. 4 S., R. 8 E.). The section is entirely within the Chadron Formation. The base is exposed in a roadside ditch as 0.1 m of massive, fine-grained feldspathic arenite, overlain by a limestone 20 cm thick. This limestone is gray but weathers reddish, is laminated at the base (lithofacies C), and is a massive peloidal packstone with *Paleophycus* burrows near the top (lithofacies B). This limestone is overlain by 0.9 m of gray mudstone that weathers reddish, and 6.8 m covered section (with mudstone and limestone float).

The upper limestone forms the mesa caprock, and is 2.1 m thick. It consists of 25 cm of intercalated red, black, and gray marl and clotted micrite with *Planolites* tracks (lithofacies E), overlain by 12 cm of finely laminated algal bindstone, 3 cm of marl, and 20 cm of finely laminated algal bindstone (transitional between lithofacies D and E). This is overlain by an algal stromatolite consisting of laterally linked hemispheroids up to 12 cm tall (lithofacies D). Overlying the stromatolite is 25 cm of flat algal bindstone that pinches and swells in thickness to fill existing microtopography (lithofacies C). Overlying the algal mat is undulose-bedded limestone 50 cm thick, with bioclast accumulations, and nonfabric selective, replacement chalcedony (lithofacies B). The shallowing-upward sequence is capped by charophyte-rich massive limestone 20 cm thick (lithofacies A). Immediately above this is a second shallowing upward sequence, consisting of 20 cm of laminated limestone (lithofacies C), 10 cm of undulose bedded limestone (lithofacies B), and 15 cm of charophyte-rich massive limestone (lithofacies A) that forms the caprock.

French Creek section. The section at French Creek, Custer County, South Dakota (NE1/4 sec. 34, T. 4 S, R. 9 E.) contains one tufa mound. The base of the section is in the Pierre Shale and superimposed Yellow Mounds Paleosol. The overlying Chamberlain Pass Formation consists of a channel sequence (1.95 m of trough cross-bedded, pebbly, medium- to very coarse grained, argillaceous quartz arenite), and overlying flood-plain sequence (0.9 m of mottled red and green mudstone with roots and glaebules).

The overlying Chadron Formation consists of 3.6 m of gray mudstone incised by a multistory sandstone channel complex (47 cm of trough cross-bedded, pebbly, coarse- to very coarse grained feldspathic arenite, 12 cm of gray siltstone; and 1.3 m of planar and trough cross-bedded, pebbly, medium- to very coarse grained feldspathic arenite containing turtle plastron plates, with a vertebrate trackway on the upper bedding surface). This channel complex is overlain by 4.6 m of gray mudstone. The tufa mound complex forms a 30-cm-thick caprock unit. It consists of pisolite-intraclast wackestones (lithofacies F).

Bear Creek section. The section at Bear Creek, Pennington County, South Dakota (SE1/4 sec. 11, T. 3 S., R. 13 E.) contains one tufa mound (Fig. 17). The base of the section is the Cretaceous Pierre Shale and superimposed Yellow Mounds Paleosol. The overlying Chamberlain Pass Formation consists of a floodplain sequence (1.6 m of mottled red and green mudstone with glaebules and clay-filled rootcasts) overlain by a channel sequence (1.35 m of lenticular, massive, pebbly, medium- to coarse-grained argillaceous quartz arenite with mosaic crack fills and clay-filled rootcasts in the upper part), overlain by a second flood-plain sequence (1.5 m of mottled red and green mudstone with rootcasts), and a second multistory channel sequence (2.53 m of trough cross-bedded and massive, pebbly, coarse- to very coarse grained argillaceous quartz arenite, with lenses of imbricated gravel and mosaic crack fills).

The overlying Chadron Formation consists of 2.1 m of gray mudstone capped by a tufa mound sequence. The tufas are offlapping sheets, up to 25 cm thick, of arenaceous pisolite-intraclast wackestone, with rootcasts (lithofacies F).

Railroad Buttes section. Railroad Buttes is located southeast of Farmingdale, South Dakota (NE1/4 sec. 7, T. 1 S., R. 11 E.). The section consists of two limestones and four tufa complexes in the Chadron Formation.

The base of the section is in the Cretaceous Pierre Shale. The pedogenically altered upper portion of the Pierre Shale (Yellow Mounds Paleosol) is approximately 9.4 m thick at this location, noted by blocky ped structures and argillans. The Chamberlain Pass Formation is not present at this location. The remainder of the section is the Chadron Formation.

Overlying the Yellow Mounds Paleosol is an overbank sequence of gray mudstone 384 cm thick that is a slope former. An overlying tufa sequence 93 cm thick consists of numerous, thin (3-8 cm thick), pisolite–mud intraclast wackestones and porous micrite (lithofacies F) that are partly to completely draped by siliciclastic mudstones several centimeters thick. The tufa sequence is overlain by gray siliciclastic mudstones 146 cm thick interstratified with a few thin (1–3 cm thick), laterally discontinuous tufa beds. A 40-cm-thick pisolite-intraclast wackestone overlies this, and forms a notable step in the topography. This tufa is overlain by 520 cm of gray mudstone with a few thin (<1 cm thick), laterally discontinuous tufa beds.

Overlying this is the third tufa complex, being 203 cm thick in total. This tufa complex consists of numerous pisolite-intraclast wackestones and alternating massive and porous micrites that are partly to completely draped by thin, gray, siliciclastic mudstones. This tufa complex forms a prominent bench in the topography, and the individual tufa layers thicken upward to a capping bed 43 cm thick, which is massive in the lower portion and porous in the top. Overlying this tufa complex is 360 cm of gray siliciclastic mudstone.

The fourth tufa complex is 215 cm thick, again consisting of pisolite-intraclast wackestone and porous micrite with interbedded siliciclastic mudstones. The lowest tufa contains rhizoliths, mud intraclasts, and vertebrate bone debris. It weathers spheroidally, with tan-red hues. The siliciclastic mudstones fill paleotopography over the tufas, thickening to 80 cm in places. Overlying this tufa complex is a gray mudstone 360 cm thick with one thin (5 cm), laterally discontinuous, tufa 110 cm up from the base.

The overlying limestone is 75 cm thick, consisting of laminated limestone about 5 cm thick at the base (lithofacies C), the rest being a massive ostracode wackestone (lithofacies A) with mudcracks at the top. Bioturbation is evident in the upper portion of the limestone (*Taenidium* and *Skolithos* burrows). Approximately 180 cm of gray mudstone overlies the limestone. The second limestone was not found in place, but consists of float—as much as 5 cm thick—of an ostra-

code-charophyte wackestone (lithofacies A) that overlies the mudstone and forms the top of Railroad Buttes.

Mile post 114 section. This section is located east of Wall, Pennington County, South Dakota, near mile post 114 of Interstate 90 (SE1/4 sec. 10, T. 1 S., R. 16 E.). The section is considered part of the Bloom Basin Limestone Bed of Welzenbach (1992).

The base of this section is in the Cretaceous Pierre Shale and superimposed Yellow Mounds Paleosol. The overlying Chamberlain Pass Formation consists of a flood-plain sequence (1.1 m of mottled red and green mudstone), an overlying channel sequence (1.6 m of trough cross-bedded and ripple-laminated, coarse-grained, argillaceous quartz arenite with glaebules and pedogenic nodules), and a flood-plain and channel-margin sequence (1.1 m of mottled red and green mudstone; 30 cm of massive, fine-grained, argillaceous quartz arenite; and 1.85 m of mottled, red and green mudstone with rootcasts).

The overlying Chadron Formation consists of a flood-plain and channel-margin sequence (5.1 m of gray mudstone; 30 cm of massive, fine-grained feldspathic arenite; 40 cm of gray mudstone; 25 cm of massive, fine-grained feldspathic arenite; and 7.2 m of gray mudstone) capped by a tufa sequence 12 cm thick. The tufa is an arenaceous, pisolite-intraclast wackestone with rootcasts (lithofacies F) that forms a local caprock to small mesas or mounds.

The limestone overlies a channel-margin sequence (70 cm of gray mudstone; 30 cm of very fine grained, rippled, feldspathic arenite; 50 cm of gray mudstone; 80 cm of trough cross-bedded and ripple laminated medium-grained feldspathic arenite), and flood-plain sequence (1.8 m of gray mudstone). The limestone is a bench-forming unit 25 cm thick. It consists of 17 cm of stromatolitic and finely laminated algal bindstone with *Planolites* at the base (lithofacies D and C), overlain by 4 cm of undulose bedded limestone (lithofacies B), overlain by 4 cm of charophyte-rich massive limestone (lithofacies A) with a charophyte floatstone laminae at the top.

Bloom Basin section. The section at Bloom Basin, east of Wall, Pennington County, South Dakota (NE1/4 sec. 35, T. 1 S., R. 16 E.), includes a composite limestone section about 1.1 m thick and is the type section of Welzenbach's (1992) Bloom Basin Limestone Bed. This section is entirely in the Chadron Formation. The base of the section is in gray mudstone, overlain by a tufa mound sequence 1.3 m thick. The tufa consists of offlapping sheets 15–30 cm thick, with partial to complete mudstone drapes. Each sheet consists of pisolite-intraclast wackestone, with mudball intraclasts, and root molds (lithofacies F).

The limestone overlies the following succession: flood-plain deposits (3.1 m of gray mudstone), a channel sequence (1.3 m of trough cross-bedded, medium-grained, feldspathic arenite), flood-plain deposits (3.2 m of gray mudstone), a channel-margin sequence (10 cm of massive, very fine grained feldspathic arenite; 50 cm of mottled green mudstone with rootcasts; 20 cm of massive, very fine grained feldspathic arenite; 75 cm of gray mudstone; 20 cm of massive, very fine grained feldspathic arenite; 80 cm of green mudstone with rootcasts; 25 cm of massive, very fine grained feldspathic arenite with pedogenic nodules), and flood-plain deposits (2.4 m of gray mudstone).

The limestone is a bench-forming unit 1.1 m thick that consists of three separate shallowing-upward sequences. The lowest sequence consists of thin (<1 cm) algal bindstone (lithofacies C), overlain by thin (<1 cm) undulose bedded limestone (lithofacies B), overlain by 15 cm of charophyte-rich massive limestone (lithofacies A). The middle limestone consists of 15 cm of marl (lithofacies E), overlain by 5 cm of laminated limestone (lithofacies C), 5 cm of undulose bedded limestone (lithofacies B), and 10 cm of charophyte-rich massive limestone with a fish bone and turtle plastron plate floatstone laminae at the top (lithofacies A). The upper limestone consists of 15 cm of marl (lithofacies E), overlain by 10 cm of laminated limestone with *Planolites* (lithofacies C), 10 cm of undulose bedded limestone (lithofacies B), and 15 cm of charophyte-rich massive limestone at the top, with *Taenidium* burrows (lithofacies A). This unit forms the mesa caprock.

Mile post 118 section. This section is located south of mile post 118 on Interstate 90, Pennington County, South Dakota (NE1/4 sec. 31, T. 1 S., R. 17 E.). This section exposes a series of tufa mounds and tufa tubes, with an overlying limestone.

The base of the section is in the Cretaceous Pierre Shale and superimposed Yellow Mounds Paleosol. The overlying Chamberlain Pass Formation consists of a multistory sandstone channel complex (2.1 m of massive, mottled, fine-grained, argillaceous quartz arenite with glaebules and clay-filled rootcasts) overlain by a channel-margin sequence (65 cm of mottled red and green mudstone; 20 cm of green mudstone, and 7 cm of rippled, very fine grained calcareous quartz arenite), and by a channel sequence (2.6 m of trough cross-bedded, fine- to medium-grained, argillaceous quartz arenite with mud intraclasts and clay-filled rootcasts).

The overlying Chadron Formation consists of 1.7 m of gray mudstone with mottling and glaebules near the top, followed by a tufa mound sequence (30 cm of tufa, 75 cm of gray mudstone, 60 cm of tufa with an interstratified partial mudstone drape 2 cm thick, 20 cm of marl, 50 cm of gray mudstone, and 80 cm of tufa). Each of these tufas are pisolite-intraclast wackestone (lithofacies F), containing rootcasts and forming local caprocks for mounds or small mesas. Laterally correlated with this sequence of tufas are pinnacles of sandstone standing in erosional relief at about 1.5 m tall. The pinnacles represent local places of calcite and silica cementation (i.e., nonpedogenic calcrete) in the sandstone. The features exhibit internal zoning and fluid escape features that partially disrupt or destroy primary sedimentary structures. These pinnacles are interpreted as vertical or subvertical conduits associated with groundwater discharge at springs (J. E. Evans, unpublished data, 1994).

The limestone overlies the following succession: flood-plain deposits (1.3 m of gray mudstone), a channel-margin sequence (25 cm of ripple-laminated, very fine grained feldspathic arenite; 60 cm of gray mudstone; 10 cm of ripple-laminated, very fine grained, feldspathic arenite with pedogenic nodules; 85 cm of gray mudstone; 10 cm of massive, very fine grained calcareous feldspathic arenite); and flood-plain deposits (4.8 m of gray mudstone, 20 cm of massive, very fine-grained feldspathic arenite, and 3.1 m of gray mudstone).

The limestone is 25 cm thick, and consists of 12 cm of laminated limestone with *Planolites* (lithofacies C) overlain by 2 cm of undulose bedded limestone (lithofacies B), overlain by 11 cm of charophyte-rich massive limestone (lithofacies A) with a gastropod floatstone and mudcracks at the top. There are possible synaeresis cracks in the laminated limestone.

Mile post 121 section. This section, located south of mile post 121 on Interstate 90, Pennington County, South Dakota (SW1/4 sec. 34, T. 1 S., R. 17 E.) is entirely within the Chadron Formation and includes three limestones and one tufa sequence. The base of the section is a floodplain sequence (6.5 m of gray mudstone overlain by 50 cm of mottled red and gray mudstone), overlain by a tufa ~5 cm thick. The tufa is an arenaceous pisolite-intraclast wackestone with root molds (lithofacies F).

The lower limestone overlies a flood-plain sequence (60 cm of mottled red and gray mudstone and 4.5 m of gray mudstone). The limestone is 25 cm thick and consists of 3 cm of laminated limestone at the base (lithofacies C), overlain by 10 cm of undulose bedded limestone (lithofacies B), capped by 12 cm of charophyte-rich massive limestone with a charophyte floatstone laminae at the top (lithofacies A).

The second limestone overlies a flood-plain sequence (1.8 m of gray mudstone, 75 cm of mottled red and gray mudstone, 70 cm of gray mudstone, 35 cm of mottled red and gray mudstone, and 93 cm of gray mudstone). The limestone is a bench-forming unit 10 cm thick. It consists of 4 cm of laminated limestone (lithofacies C) at the base, overlain by 6 cm

of undulose bedded limestone with oncolites (lithofacies B), with a thin (<1 cm) charophyte floatstone laminae (lithofacies A) at the top.

The upper limestone overlies 4.3 m of poorly exposed gray mudstone with limestone float. The limestone is a siliceous caprock unit 5 cm thick. It consists of 2 cm of laminated limestone at the base (lithofacies C), overlain by 3 cm of charophyte-rich massive limestone at the top with *Taenidium* burrows (lithofacies A).

Street Hill section. This section, located at Street Hill (NW1/4 sec. 26, T. 1 S., R. 17 E.) is a small exposure of a single lacustrine limestone within the Chadron Formation. The limestone is 24 cm thick, and consists of 6 cm of undulose-bedded limestone (lithofacies B) that overlies green shale. This is in turn overlain by 3 cm of charophyte-rich massive limestone (lithofacies A), 8 cm of undulose-bedded limestone (lithofacies B), 3 cm of laminated limestone (lithofacies C), and 4 cm of charophyte-rich massive limestone (lithofacies A). Charophyte stems are particularly well preserved at this outcrop.

Walker Hill section. This section is located about 3.1 km (2 mi) south of Cottonwood, South Dakota (SE1/4 sec. 19, T. 1 S., R. 19 E.), at Walker Hill. This outlier of the Chadron Formation consists of calcareous mudstones and nodular calcrete and probably represents lateral facies equivalents to the limestone of the Bloom Basin Limestone Bed.

REFERENCES CITED

Beaumont, E. A., 1979, Depositional environments of Fort Union sediments (Tertiary, northwest Colorado) and their relation to coal: American Association of Petroleum Geologists Bulletin, v. 63, p. 194–217.

Bradley, W. H., 1964, Geology of the Green River Formation and associated Eocene rocks in southwestern Wyoming and adjacent parts of Colorado and Utah: U.S. Geological Survey Professional Paper, 496-A, 86 p.

Bromley, R. G., and Asgaard, U., 1979, Triassic freshwater ichnocoenses from Carlberg Fjord, East Greenland: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 28, p. 39–80.

Brunskill, G. J., and Harriss, R. C., 1969, Fayetteville Green Lake, New York. IV. Interstitial water chemistry of the sediments: Limnology and Oceanography, v. 14, p. 858–861.

Bump, J. D., 1956, Geographic names for the members of the Brule Formation in the Big Badlands of South Dakota: American Journal of Science, v. 254, p. 429–432.

Cabrera, Ll., and Saez, A., 1987, Coal deposition in carbonate-rich shallow lacustrine systems: the Calaf and Mequinenza sequences (Oligocene, eastern Ebro Basin, NE Spain): Journal of the Geological Society of London, v. 144, p. 451–461.

Chamberlain, C. K., 1975, Recent lebensspuren in nonmarine aquatic environments, *in* Frey, R. E., ed., The study of trace fossils: New York, Springer-Verlag, p. 431–458.

Clark, J., 1954, Geographic designation of the members of the Chadron Formation in South Dakota: Carnegie Museum Annals, v. 33, p. 197–198.

Clark, J., Beerbower, J. P., and Kietzke, K. K., 1967, Oligocene sedimentation, stratigraphy, paleoecology, and paleoclimatology in the Big Badlands of South Dakota: Fieldiana Geology Memoir 5, 158 p.

Cohen, A. S., and Thouin, C., 1987, Nearshore carbonate deposits in Lake Tanganyika: Geology, v. 15, p. 414–418.

Cojan, I., 1993, Alternating fluvial and lacustrine sedimentation: tectonic and climatic controls (Provence Basin, S. France, Upper Cretaceous/Palaeocene), *in* Marzo, M., and Puigdefabregas, C., eds., Alluvial sedimentation: International Association of Sedimentologists Special Publication 17, p. 425–438.

Cook, H. J., and Mansfield, W. C., 1933, A new mollusk from the Chadron Formation (Oligocene) of Nebraska: Washington Academy of Sciences Journal, v. 23, p. 263.

D'Alessandro, A., and Bromley, R. G., 1987, Meniscate trace fossils and the *Muensteria-Taenidium* problem: Paleontology, v. 30, p. 743–763.

Daley, B., 1972, Macroinvertebrate assemblages from the Bembridge M[illegible] (Oligocene) of the Isle of Wight, England, and their environmen[illegible] significance: Palaeogeography, Palaeoclimatology, Palaeoecolog[illegible] v. 11, p. 11–32.

Dean, W. E., and Fouch, T. D., 1983, Lacustrine environment, *in* Scholle, P. A., Bebo[illegible] D. G., and Moore, C. H., eds., Carbonate depositional environments: Americ[illegible] Association of Petroleum Geologists Memoir 33, p. 96–130.

Demicco, R. V., Bridge, J. S., and Cloyd, K. C., 1987, A unique freshwater ca[illegible] bonate from the Upper Devonian Catskill magnafacies of New Yor[illegible] State: Journal of Sedimentary Petrology, v. 57, p. 327–334.

Ekdale, A. A., Bromley, R. G., and Pemberton, S. G., 1984, The use of trac[illegible] fossils in sedimentology and stratigraphy: Society of Economic Paleon[illegible] tologists and Mineralogists, Short Course 15, 317 p.

Evanoff, E., Prothero, D. R., and Lander, R. H., 1992, Eocene-Oligocene climatic changes in North America: the White River Formation near Douglas, east-central Wyoming, *in* Prothero, D. R. and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 116–130.

Evans, J. E., 1996, Evidence for late Eocene unroofing of the Black Hills uplift: Geological Society of America Abstracts with Programs, v. 28, no. 7, p. 373.

Evans, J. E., and Terry, D. O., 1994, The significance of incision and fluvial sedimentation in the basal White River Group (Eocene-Oligocene), Badlands of South Dakota: Sedimentary Geology, v. 90, p. 137–152.

Fetter, C. W., 1994, Applied hydrogeology, third edition: New York, Macmillan, 691 p.

Flores, R. M., 1981, Coal deposition in fluvial paleoenvironments of the Paleocene Tongue River Member of the Fort Union Formation, Powder River Basin, Wyoming and Montana, *in* Ethridge, F. G., and Flores, R. M., eds., Recent and ancient non-marine depositional environments: models for exploration: Tulsa, Society of Economic Mineralogists and Paleontologists Special Publication 13, p. 169–190.

Flugel, E., 1982, Microfacies Analysis of limestones: Berlin, Springer-Verlag, 633 p.

Forester, R. M., 1991, Pliocene climate history of the western United States derived from lacustrine ostracodes: Quaternary Science Reviews, v. 10, p. 33–146.

Frey, R. W., and Pemberton, S. G., 1984, Trace fossil facies models, *in* Walker, R. G., ed., Facies models, second edition: Geoscience Canada Reprint Series 1, p. 189–208.

Freytet, P., and Plaziat, J. C., 1982, Continental carbonate sedimentation and pedogenesis—Late Cretaceous and Early Tertiary of southern France, *in* Purser, B. H., ed., Contributions to Sedimentology 12, 213 p.

Gierlowski-Kordesch, E., Gomez Fernandez, J. C., and Melendez, N., 1991, Carbonate and coal deposition in an alluvial-lacustrine setting: Lower Cretaceous (Weald) in the Ilerian Range (east central Spain, *in* Anadon, P., Cabrera, Ll., and Kelt, K., eds., Lacustrine facies analysis: International Association of Sedimentologists Special Publication 13, p. 109–125.

Gomez Fernandez, J. C., and Melendez, N., 1991, Rhythmically laminated lacustrine carbonates in the Lower Cretaceous of La Serrania de Cuenca Basin (Iberian Ranges, Spain), *in* Anadon, P., Cabrera, Ll., and Kelts, K., eds., Lacustrine facies analysis: International Association of Sedimentologists Special Publication 13, p. 245–256.

Gries, J. P., and Bishop, G. A., 1966, Fossil invertebrates from the Big Badlands of South Dakota: South Dakota Academy of Sciences Proceedings, v. 45, p. 57–61.

Hardie, L. A., Smoot, J. P., and Eugster, H. P., Saline lakes and their deposits: a sedimentological approach, *in* Matter, A., and Tucker, M. E., eds., Modern and ancient lake sediments: International Association of Sedimentologists Special Publication 2, p. 7–41.

Harksen, J. C., and Macdonald, J. R., 1969a, Guidebook to the major Cenozoic deposits of southwestern South Dakota: South Dakota Geological Survey, Guidebook 2, 103 p.

Harksen, J. C., and Macdonald, J. R., 1969b, Type section for the Chadron and

Brule Formations of the White River Oligocene in the Big Badlands, South Dakota: South Dakota Geological Survey Report of Investigations 99, 23 p.

Hayes, T. S., 1996, Suspended-sediment "reddening" in Cascade Springs, southern Black Hills, South Dakota: Geological Society of America Abstracts with Programs, v. 28, no. 4, p. A11.

Hutchison, J. H., 1992, Western North America reptile and amphibian record across the Eocene/Oligocene boundary and its climatic implications, *in* Prothero, D. R. and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 149–164.

Jones, B., and Renaut, R. W., 1995, Noncrystallographic calcite dendrites from hot-spring deposits at Lake Bogoria, Kenya: Journal of Sedimentary Research, v. A65, p. 154–169.

Kelts, K., and Hsu, K. J., 1978, Freshwater carbonate sedimentation, *in* Lerman, A., ed., Lakes: chemistry, geology, and physics: New York, Springer- Verlag, p. 295–323.

Klemp, J. A., 1996, Source aquifers for large springs in northwest Lawrence County, South Dakota: Geological Society of America Abstracts with Programs, v. 28, no. 4, p. 13.

Lander, R. H., 1991, White River Group diagenesis [Ph.D. thesis]: Urbana, University of Illinois at Urbana-Champaign, 143 p.Ludlam, S. D., 1969, Fayetteville Green Lake, New York. III. The laminated sediments: Limnology and Oceanography, v. 14, p. 848–857.

Murphy, D. H., and Wilkinson, B. H., 1980, Carbonate deposition and facies distribution in a central Michigan marl lake: Sedimentology, v. 27, p. 123–135.

Obradovich, J. D., Evanoff, E., and Larson, E. E., 1995, Revised single-crystal laser-fusion $^{40}Ar/^{39}Ar$ ages of Chadronian tuffs in the White River Formation of Wyoming: Geological Society of America Abstracts with Programs, v. 27, p. A77–78.

Olsen, P. E., 1986, A 40 million-year lake record of early Mesozoic orbital climatic forcing: Science, v. 234, p. 842–848.

Ordonez, S., and Garcia del Cura, M., 1983, Recent and Tertiary fluvial carbonates in central Spain, *in* Collinson, J. D., and Lewin, J., eds., Modern and ancient fluvial systems: International Association of Sedimentologists Special Publication 6, p. 485–497.

Pemberton, S. G., and Frey, R. W., 1982, Trace fossil nomenclature and the *Planolites-Paleophycus* dilemma: Journal of Paleontology, v. 56, p. 843–881.

Picard, M. D., and High, L. R., Jr., 1972, Criteria for recognizing lacustrine rocks, *in* Rigby, J. K., and Hamblin, W. K., eds., Recognition of ancient sedimentary environments: Tulsa, Society of Economic Mineralogists and Paleontologists Special Publication 16, p. 108–145.

Platt, N. H., 1989, Lacustrine carbonates and pedogenesis: sedimentology and origin of palustrine deposits from the Early Cretaceous Rupelo Formation, W. Cameros Basin, N. Spain: Sedimentology, v. 36, p. 665–684.

Platt, N. H., and Wright, V. P., 1991, Lacustrine carbonates: facies models, facies distributions, and hydrocarbon aspects, *in* Anadon, P., Carbrere, Ll., and Kelts, K., eds., Lacustrine facies analysis: Oxford, U. K., International Association of Sedimentologists Special Publication 13, p. 57–74.

Prothero, D. R., and Swisher, C. C., III, 1992, Magnetostratigraphy and geochronology of the terrestrial Eocene-Oligocene transition in North America, *in* Prothero, D. R. and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 46–73.

Rahn, P. H., and Gries, J. P., 1973, Large springs in the Black Hills, South Dakota and Wyoming: South Dakota Geological Survey Report of Investigations 107, 46 p.

Rahn, P. H., and Hayes, T. S., 1996, Hydrogeology of the central Black Hills: Road Log, Field Trip 3, *in* Paterson, C. J., and Kirchner, J. G., eds., Guidebook to the geology of the Black Hills, South Dakota: South Dakota School of Mines and Technology Bulletin 19, p. 19–29.

Redden, J. A., 1996, Phanerozoic stratigraphic and structural features of the 100,000 scale geologic map of the Black Hills Geological Society of America Abstracts with Programs, v. 28, no. 4, p. A35–36.

Retallack, G. J., 1983, Late Eocene and Oligocene fossil soils from Badlands National Park, South Dakota: Geological Society of America Special Paper 193, 82 p.

Retallack, G. J., 1992, Paleosols and changes in climate and vegetation across the Eocene/Oligocene boundary, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 382–398.

Ruttner, F., 1963, Fundamentals of limnology: Toronto, University of Toronto Press, 307 p.

Ryder, R. T., Fouch, T. D., and Elison, J. H., 1976, Early Tertiary sedimentation in the western Uinta Basin, Utah: Geological Society of America Bulletin, v. 87, p. 496–512.

Sanz, M. E., Zarza, A. M. A., and Calvo, J. P., 1995, Carbonate pond deposits related to semi-arid alluvial systems: examples from the Tertiary Madrid Basin, Spain: Sedimentology, v. 42, p. 437–452.

Scholl, D. W., 1960, Pleistocene algal pinnacles at Searles Lake, California: Journal of Sedimentary Petrology, v. 30, p. 414–431.

Scholl, D. W., and Taft, W. H., 1964, Algae, contributors to the formation of calcareous tufa, Mono Lake, California: Journal of Sedimentary Petrology, v. 34, p. 309–319.

Shurr, G. W., Ashworth, A. C., Benton, R., Murphy, E. C., and Biek, R. F., 1996, Regional framework for Tertiary tectonism in the northern Great Plains, *in* Paterson, C. J., and Kirchner, J. G., eds., Guidebook to the geology of the Black Hills, South Dakota: South Dakota School of Mines and Technology Bulletin 19, p. 129– 134.

Simpson, S., 1975, Classification of trace fossils, *in* Frey, R. W., ed., The study of trace fossils: New York, Springer-Verlag, p. 39– 54.

Simpson, W. F., 1985, Geology and paleontology of the Oligocene Harris Ranch badlands, southwestern South Dakota, *in* Martin, J. E., ed., Fossiliferous Cenozoic deposits of western South Dakota and northwest Nebraska: Dakoterra, v. 2, pt. 2, p. 303–333.

Smith, G. R., and Miller, R. R., 1985, Taxonomy of fishes from Miocene Clarkia lake beds, Idaho, *in* Smiley, C. J., ed., Late Cenozoic history of the Pacific Northwest: San Francisco, American Association for the Advancement of Science, Pacific Division, p. 75–83.

Squires, R. L., and Advocate, D. M., 1984, Meniscate burrows from Miocene lacustrine-fluvial deposits, Diligencia Formation, Orocopia Mountains, southern California: Journal of Paleontology, v. 58, p. 593–597.

Stanley, K. O., 1976, Sandstone petrofacies in the Cenozoic High Plains sequence: eastern Wyoming and Nebraska: Geological Society of America Bulletin, v. 87, p. 279–309.

Stanley, K. O., and Collinson, J. W., 1979, Depositional history of Paleocene–Lower Eocene Flagstaff Limestone and coeval rocks, central Utah: American Association of Petroleum Geologists Bulletin, v. 63, p. 311–323.

Swisher, C. C., III and Prothero, D. R., 1990, Laser-fusion $^{40}Ar/^{39}Ar$ dating of the Eocene-Oligocene transition in North America: Science, v. 249, p. 760–762.

Taylor, J., 1978, The Paleogene freshwater community, *in* McKerrow, W. S., ed., The ecology of fossils: Cambridge, Massachusetts Institute of Technology Press, p. 343–345.

Terry, D. O., 1991, The study and implications of comparative pedogenesis of sediments from the White River Group, South Dakota [M.S. thesis]: Bowling Green, Ohio, Bowling Green State University, 183 p.

Terry, D. O., and Evans, J. E., 1994, Pedogenesis and paleoclimatic implications of the Chamberlain Pass Formation, basal White River Group, Badlands of South Dakota: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 10, p. 197–215.

Terry, D. O., LaGarry, H. E., and Wells, W. B., 1995, The White River Group revisited: vertebrate trackways, ecosystems, and lithostratigraphic revision, redefinition, and redescription, *in* Diffendal, R. F., and Flowerday,

C. A., eds., Geologic field trips in Nebraska and adjacent parts of Kansas and South Dakota: Nebraska Conservation and Survey Division Guidebook 10, p. 43–57.

Van Houten, F. B., 1964, Cyclic lacustrine sedimentation, Upper Triassic Lockatong Formation, central New Jersey and adjacent Pennsylvania: Kansas Geological Survey Bulletin, v. 169, p. 497–531.

Wanless, H. R., 1922, Lithology of the White River sediments: American Philosophical Society Proceedings, v. 61, p. 184–203.

Wanless, H. R., 1923, The stratigraphy of the White River Beds of South Dakota: American Philosophical Society Proceedings, v. 62, p. 190–269.

Welzenbach, L. C., 1992, Limestones in the Lower White River Group (Eocene-Oligocene), Badlands of South Dakota: Depositional environment and paleoclimatic implications [M.S. thesis]: Bowling Green, Ohio, Bowling Green State University, 131 p.

Whittlesey, K. E., and Prothero, D. R., 1995, Magnetic stratigraphy and stratigraphy of the Chadronian/Orellan boundary in the White R Group: Geological Society of America Abstracts with Programs, v p. A95.

Wray, J.L., 1977, Calcareous algae: *in* Developments in Paleontology Stratigraphy, New York, Elsevier, v. 4, p. 107–137.

Wright, V. P., 1990, Lacustrine carbonates, *in* Tucker, M. E., and Wright, V eds. Carbonate sedimentology: London, Blackwell, p. 164–190.

Yuretich, K., 1989, Paleocene lakes of the central Rocky Mountains, west United States: Palaeogeography, Palaeoclimatology, Palaeoecolo v. 70, p. 53–63.

MANUSCRIPT ACCEPTED BY THE SOCIETY DECEMBER 5, 1997

C

Printed in U.S.A.

Geological Society of America
Special Paper 325
1998

The Big Cottonwood Creek Member: A new member of the Chadron Formation in northwestern Nebraska

Dennis O. Terry, Jr.
Department of Geology, University of Nebraska—Lincoln, Lincoln, Nebraska 68588-0340
Hannan E. LaGarry
Conservation and Survey Division (IANR) and State Museum, University of Nebraska—Lincoln, Lincoln, Nebraska 68588-0514

ABSTRACT

Based on detailed lithologic correlations of the White River Group in Nebraska and South Dakota, we revise and redescribe the Chadron Formation. We redescribe the upper two-thirds of the Chadron Formation ("Chadron B" and "Chadron C"), and the lowermost Brule Formation ("Orella A") in northwestern Nebraska as a new lithostratigraphic member of the Chadron Formation. We name this new unit the Big Cottonwood Creek Member for exposures along Big Cottonwood Creek at Toadstool Park, 30 km northwest of Crawford, Nebraska. The terms Chadron B, Chadron C, and Orella A are abandoned. The Big Cottonwood Creek Member is composed primarily of volcaniclastic overbank silty claystones interbedded with tabular and lenticular channel sandstones, lacustrine limestones, pedogenic calcretes, marls, volcanic ashes, and gypsum. The Big Cottonwood Creek Member differs from underlying bluish-green, gray, and olive claystones of the Peanut Peak Member of the Chadron Formation in that it is siltier, variegated in color, has a higher unaltered vitric volcaniclastic component, contains more lacustrine limestone, pedogenic calcrete, and marl interbeds, and tends to erode into cliffs rather than low hummocks. The lower boundary of the Big Cottonwood Creek Member is an intertonguing contact with the underlying Peanut Peak Member of the Chadron Formation, or it is a local unconformity where the Big Cottonwood Creek Member fills valleys and depressions. The upper boundary is recognized at the lithologic change from pedogenically modified green, red, and pink volcaniclastic silty claystones of the Big Cottonwood Creek Member to thinly interbedded and less pedogenically modified brown, orange, and tan volcaniclastic clayey siltstones and sheet sandstones of the Orella Member of the Brule Formation. This revised upper boundary occurs 9 to 10 m above the "upper purplish-white layer," the previously recognized boundary between the Chadron and Brule Formations, and is intertonguing except where channel sandstones of the overlying Orella Member of the Brule Formation incise the Big Cottonwood Creek Member.

INTRODUCTION

The late Eocene-Oligocene White River Group of the northern Great Plains is composed of volcaniclastic and siliciclastic claystones and siltstones, limestones, and sandstones deposited within fluvial, lacustrine, and eolian environments. These deposits are well known for their vertebrate fossils, the collection of which formed the basis for much of the early biostratigraphic and lithostratigraphic research in this region (Meek and Hayden, 1857, 1861; Hatcher, 1893; Darton, 1899;

Terry, D. O., Jr., and LaGarry, H. E., 1998, The Big Cottonwood Creek Member: A new member of the Chadron Formation in northwestern Nebraska, *in* Terry, D. O., Jr., LaGarry, H. E., and Hunt, R. M., Jr., eds., Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America): Boulder, Colorado, Geological Society of America Special Paper 325.

Wanless, 1922, 1923; Ward, 1922; Osborn, 1929; Clark, 1937, 1954; Schultz and Stout, 1955; Bump, 1956; Clark et al., 1967; Schultz and Falkenbach, 1968). During this early period, type areas for competing stratigraphic classifications for the White River Group, composed of the Chadron and Brule Formations, were established in Nebraska and South Dakota (Fig. 1). This competition between type areas, and the ensuing controversies, have been summarized by Clark et al. (1967), Harksen and Macdonald (1969), and Singler and Picard (1980). Since this early lithostratigraphic work, stratigraphic methods and procedures of codification have evolved (see NACSN, 1983). However, subsequent studies of the White River Group have focused on subjects other than lithostratigraphy and biostratigraphy, such as magnetostratigraphy (Prothero and Swisher, 1992), paleopedology (Retallack, 1983; Terry and Evans, 1994), subsurface geology (Swinehart et al., 1985), and geochronology (Swisher and Prothero, 1990).

Recently, however, there has been renewed interest in the lithostratigraphy and biostratigraphy of the White River Group. Reexamination of the earlier work and application of the North American Stratigraphic Code (NACSN, 1983) to the classic areas in Nebraska and South Dakota has led to the establishment of the Chamberlain Pass Formation in South Dakota (Evans and Terry, 1994), revision and redescription of the lower portion of the Chadron Formation and "Interior Zone" in northwestern Nebraska (Terry, this volume), revision of the lithostratigraphic boundaries of the Brule Formation in northwestern Nebraska (LaGarry, this volume), and a modified biostratigraphic zonation for selected geographic areas with exposures of the White River Group (Prothero and Whittlesey, this volume). This chapter reexamines the lithostratigraphy of the upper portion of the Chadron Formation in northwest Nebraska within the context of the North American Stratigraphic Code (NACSN, 1983). Based on our findings, we introduce a newly recognized lithostratigraphic unit, the *Big Cottonwood Creek Member* of the Chadron Formation.

CHADRON FORMATION IN NORTHWESTERN NEBRASKA

Regional uplift during the Laramide Orogeny forced the retreat of the Cretaceous Interior Seaway, resulting in the subaerial exposure and weathering of rock units from Early Cretaceous to Eocene age across the northern Great Plains. In northwestern Nebraska, the late Eocene-Oligocene White River Group rests unconformably on the pedogenically modified

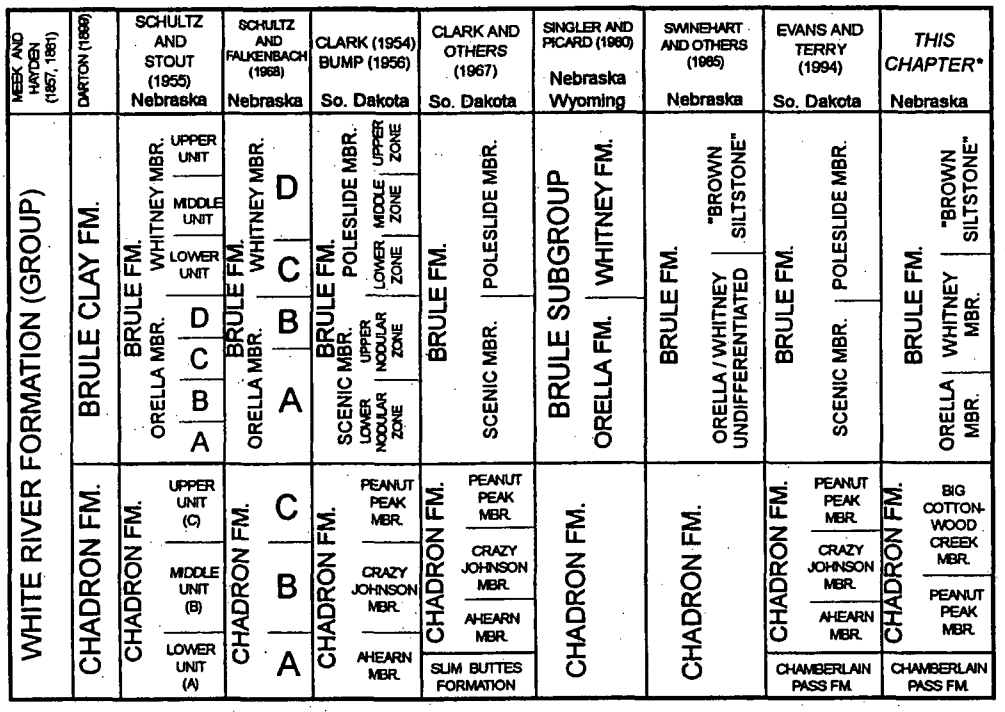


Figure 1. Stratigraphic nomenclature of the White River Group in northwestern Nebraska and southwestern South Dakota. The stratigraphy of Schultz and Stout (1955) is most commonly used for the White River Group in Nebraska. Asterisk symbol includes Terry (this volume), Terry and LaGarry (this chapter), and LaGarry (this volume). Modified from Singler and Picard (1980).

Cretaceous Pierre Shale (Pettyjohn, 1966; Retallack, 1983; Terry, this volume). The Hartville, Laramie, and Black Hills uplifts supplied sediment for rivers that flowed east-southeast across the study area (Clark, 1975; Stanley and Benson, 1979; Swinehart et al., 1985). Uplift of the Black Hills is believed to have ended in the Eocene prior to deposition of the White River Group (Lisenbee and DeWitt, 1993). The bulk of the White River Group is composed of airfall and reworked volcaniclastics derived from sources in Nevada and Utah (Larson and Evanoff, 1995; this volume). In northwestern Nebraska the White River Group is overlain by the Arikaree Group.

The stratigraphic classification of Schultz and Stout (1955) subdivided the Chadron Formation into units termed A, B, and C, and Lower, Middle, and Upper. The Orella and Whitney Members of the Brule Formation were also divided into Lower, Middle, and Upper units (Fig. 1). This classification has formed the basis for the majority of research on the White River Group in northwestern Nebraska, along with proposed correlations to other areas (Schultz et al., 1955; Leonard, 1957; Vondra, 1958, 1960; Swenson, 1959; Harvey, 1960; Schultz and Falkenbach, 1968; Stone, 1972; Singler and Picard, 1979, 1980, 1981; Schultz and Stout, 1980; Retallack, 1983; Martin, 1985; Swinehart et al., 1985; Prothero and Swisher, 1992; Murphy et al., 1993). Schultz and Stout (1955) defined boundaries between various subdivisions of the Chadron Formation at prominent marker beds that consisted of paleosols, volcanic ashes, gypsums, and limestones. They named these marker beds "purplish-white layers," and suggested that they marked regional unconformities. Based on the inferred presence of these unconformities, Schultz and Stout (1955) subdivided the Chadron Formation into the Chadron A, Chadron B, and Chadron C, which they, respectively, correlated to the Ahearn, Crazy Johnson, and Peanut Peak Members of Clark's (1954) Chadron Formation in the Big Badlands of South Dakota (Fig. 2).

The Chadron A, which Schultz and Stout (1955) also called the Yoder and correlated to the Yoder Beds of Schlaikjer

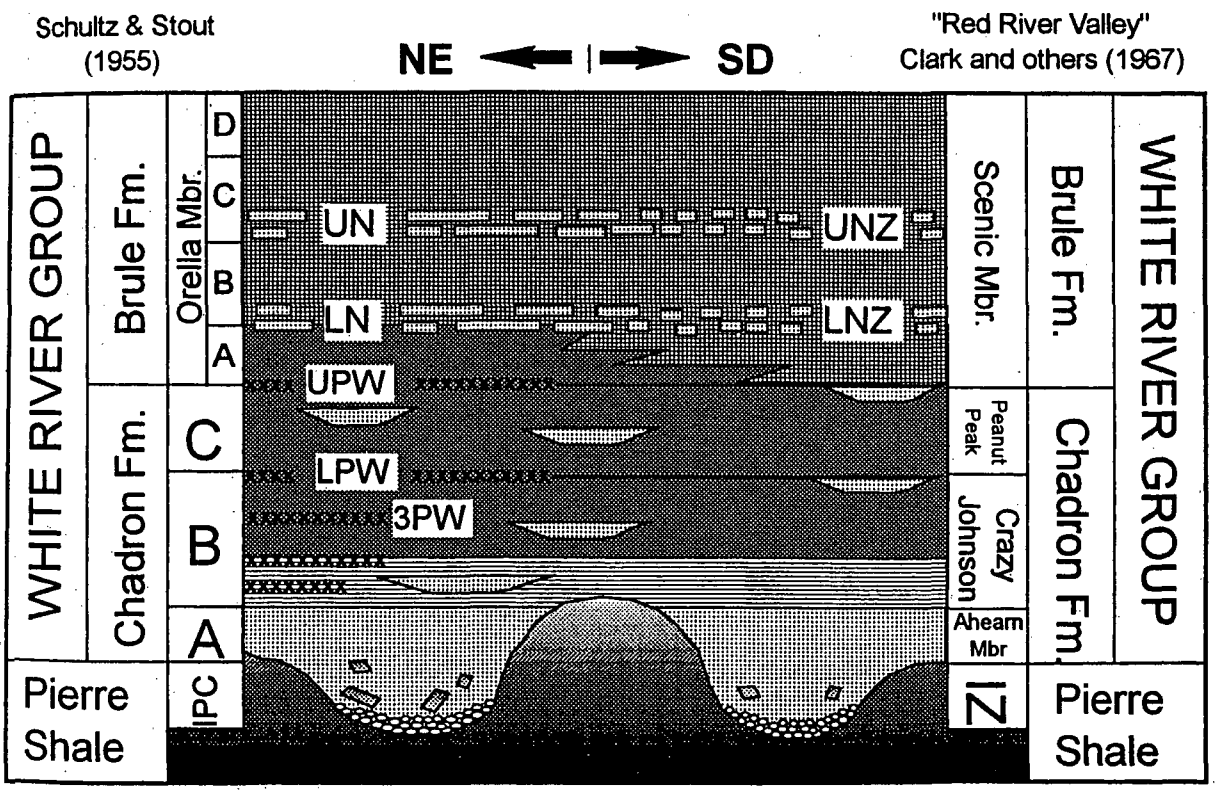


Figure 2. Correlation of the White River Group between Nebraska and South Dakota as suggested by Schultz and Stout (1955). See Terry (this volume) for a revised correlation model. UN = upper nodules; LN = lower nodules; UPW = upper purplish-white layer; LPW = lower purplish-white layer; 3PW = third purplish-white layer; IPC = Interior Paleosol Complex of Schultz and Stout (1955). UNZ = upper nodular zone; LNZ = lower nodular zone; IZ = Interior Zone of Clark et al. (1967). See Figure 4 for lithologic symbols used in this figure and throughout this chapter.

(1935), is a thick succession of yellowish-olive and white multistory channel sandstone bodies with a basal pebble and cobble lag. The Chadron A is incised into what Schultz and Stout (1955) called the Interior Paleosol Complex, a weathered zone developed on various units ranging from early Cretaceous to Eocene age across the Great Plains (Pettyjohn, 1966). Their Chadron B is composed of bluish-green and gray mudstones that weather into convex outward, hummocky mounds at its base and become variegated and siltier up section, ending at Schultz and Stout's (1955) "lower purplish-white layer" (Fig. 2). Schultz and Stout's (1955) Chadron C is defined as the variegated siltstones between their lower purplish-white layer and their upper purplish-white layer. Schultz and Stout (1955) described three additional purplish-white layers within the Chadron B—the third, fourth, and fifth purplish-white layers—which were also used for regional correlations (Fig. 2).

Schultz and Stout's (1955) contact between the Chadron and Brule Formations in northwestern Nebraska is marked by the upper purplish-white layer, a volcanic ash of regional extent (Evanoff et al., 1992). They divided the Brule Formation of northwestern Nebraska into the Orella (lower) and Whitney (upper) Members. They subdivided the Orella Member into the Orella A, Orella B, Orella C, and Orella D, based on the presence of paleosol complexes, nodular beds, and other distinctive marker horizons (Fig. 1). The Orella A rests directly on the upper purplish-white layer and overlies their Chadron Formation (Fig. 2). The Orella A at Toadstool Park is marked at its top by the lower nodules (Schultz and Stout, 1955), a zone of multiple sheet sandstones that fracture and weather into nodular shapes (Wells et al., 1994, 1995). See LaGarry (this volume) for a detailed discussion of proposed lithostratigraphic revisions for the Brule Formation in northwestern Nebraska.

BIG COTTONWOOD CREEK MEMBER OF THE CHADRON FORMATION

Based on preliminary geologic mapping, LaGarry-Guyon and Hunt (1992, 1993) concluded that the marker beds used by Schultz and Stout (1955) to define lithostratigraphic boundaries within the White River Group in northwestern Nebraska were areally restricted and not mappable outside the type area. Subsequent geologic mapping (LaGarry and LaGarry, 1997) supported these conclusions and contributed to a revised lithologic definition of the Chadron Formation suggested by Terry and LaGarry (1994, 1995), Terry (1995), LaGarry (1995), and Terry et al. (1995). However, none of these prior treatments satisfied the publication requirements of the North American Stratigraphic Code (NACSN, 1983) for the revision, redescription, and redefinition of lithostratigraphic units. The following description, definition, and discussion of the Big Cottonwood Creek Member is intended to satisfy these requirements and is based on guidelines contained within Articles 23 through 26 of the North American Stratigraphic Code (NACSN, 1983).

Intent and utility

Earlier workers in different regions decided how best to describe lithostratigraphic features, conceptualize stratigraphic models, and synthesize data in a published format. In the case of the White River Group of northwestern Nebraska, particular concepts and preferences became entrenched in the literature. Schultz and Stout's (1955) classification of the White River Group was based on a cycle-driven model of erosion and deposition developed for Quaternary sediments (Schultz and Stout, 1945, 1948, 1955; 1977; Stout, 1971, 1973, 1977, 1978). Within this model, synchronous time, rock, and faunal units were bracketed by unconformities. Within the White River Group, these unconformities were manifested as paleosols and associated marker beds (the purplish-white layers of Schultz and Stout, 1955, and Schultz et al., 1955). In the current terminology of the North American Stratigraphic Code (NACSN, 1983), an unconformity-bounded classification is an allostratigraphic classification (e.g., allogroups, alloformations, and allomembers), which is not directly equivalent to, and must exist separately from, a standard lithostratigraphic classification (groups, formations, and members).

Schultz and Stout's (1955) stratigraphic classification of the White River Group was explicitly intended as a guide for the stratigraphic collection of fossil vertebrates. Although these unconformity defined units (Fig. 2), along with their associated underlying concepts, have been directly compared to lithologically defined units elsewhere (e.g., Clark et al., 1967), Schultz and Stout's (1955) recognition, usage, and correlation of lithostratigraphic units within the White River Group of northwestern Nebraska is based on criteria other than lithology. This practice is not recommended by the North American Stratigraphic Code (NACSN, 1983). The code requires that lithostratigraphic units be recognized and defined according to the following criteria:

1. Solely on observable rock (lithic) characteristics (defined by the code as lithostratigraphy);
2. Independent from inferred geologic history (defined by the code as allostratigraphy);
3. Independent from faunal content (defined by the code as biostratigraphy);
4. Independent from time concepts (defined variously by the code as chronostratigraphy, magnetostratigraphy, or tephrostratigraphy).

The North American Stratigraphic Code (NACSN, 1983) also requires that (1) unit boundaries be placed only at observable lithologic changes; (2) once established, type sections can never be changed; and (3) lithostratigraphic units should correspond to genetic units whenever feasible. Placement of lithostratigraphic unit boundaries at observable changes in lithology precludes boundary placement at *marker beds* (e.g., prominent gypsums, limestones, ashes), *event horizons* (e.g., paleosols, volcanic ashes), or at local or regional *disconformities*, unless a demonstrable change in lithology is present at the marker bed, event horizon, or disconformity. We suggest that the marker

beds, event horizons, and disconformities within the Chadron Formation of northwestern Nebraska, while useful in biostratigraphy, should not be employed in the definition of lithostratigraphic units. Therefore, such use of event bed boundaries within the Chadron Formation of northwestern Nebraska should be discontinued, and a lithologically based classification applied instead. We propose, based on the tenets of the North American Stratigraphic Code (NACSN, 1983), that the upper two-thirds of Schultz and Stout's (1955) Chadron B, Chadron C, and the Orella A be combined into a new lithostratigraphic unit that we designate as the Big Cottonwood Creek Member of the Chadron Formation.

Category, rank, and name

The Big Cottonwood Creek Member is a lithologically defined (lithostratigraphic) member of the Chadron Formation. We name this unit the Big Cottonwood Creek Member for stratotype exposures along Big Cottonwood Creek at Toadstool Park, Sioux County, Nebraska (Fig. 3).

Priority and preservation of established names

The history of stratigraphic nomenclature for the White River Group of Nebraska and South Dakota has had various interpretations depending on the opinions of previous authors. In-depth discussions, somewhat flavored by parochialism, of the historical accounts of stratigraphic nomenclature and type sections for the White River Group are given by Harksen and Macdonald (1969) and Singler and Picard (1980). Although the type area of the White River Group is in the Big Badlands of South Dakota (White River Formation of Meek and Hayden, 1857, 1861), Darton

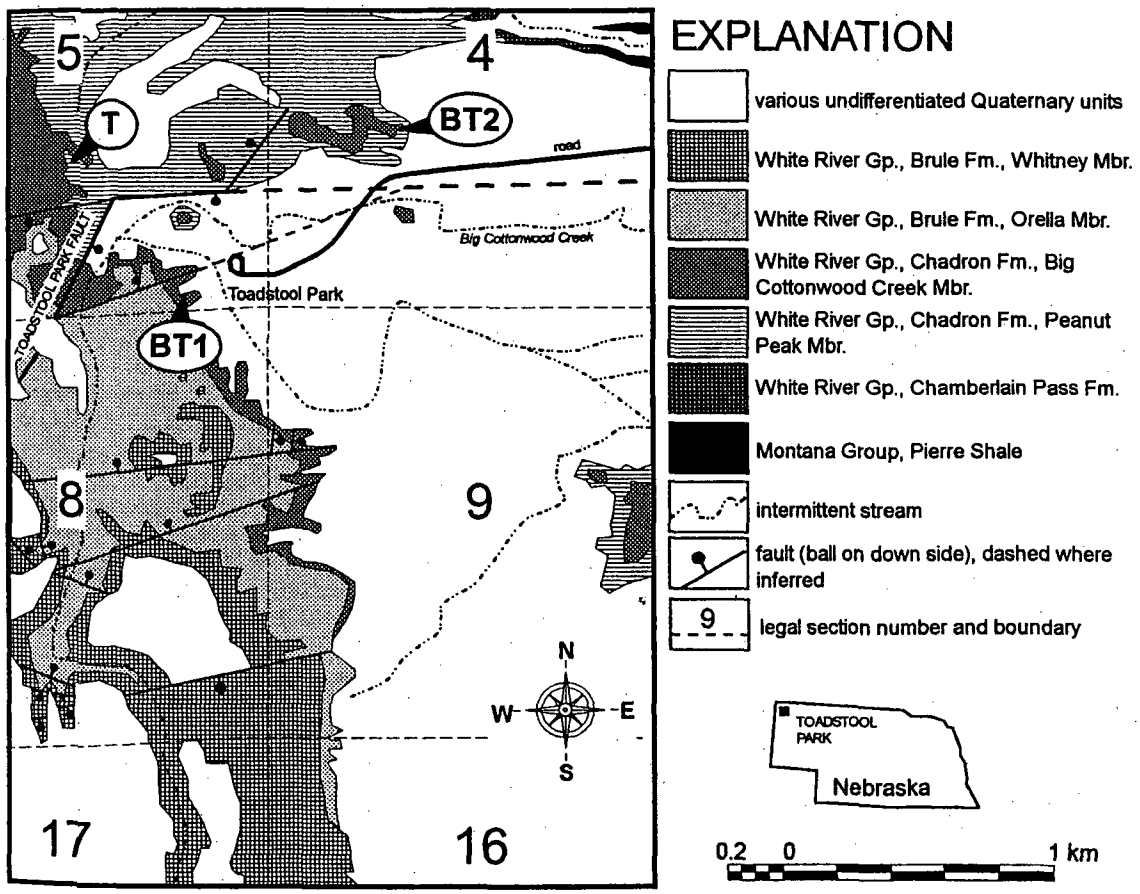


Figure 3. The type area of the Big Cottonwood Creek Member of the Chadron Formation at Toadstool Park, Sioux County, Nebraska. See Figure 4 for measured sections. T = stratotype, BT1 = boundary stratotype 1, BT2 = boundary stratotype 2. No faulting of Quaternary units was observed in this area. Geologic map data after LaGarry and LaGarry (1997).

(1899) was the first to formally subdivide the White River Group into the Chadron and Brule (=Brule Clay) Formations in his "Roundtop to Adelia" section in northwestern Nebraska. Clark (1937, 1954) subdivided the Chadron Formation in South Dakota into the Ahearn, Crazy Johnson, and Peanut Peak Members in his "Red River Valley." Bump (1956) subdivided the Brule Formation in South Dakota into the Scenic and Poleslide Members. Schultz and Stout (1938, 1955) were the first to subdivide the Chadron and Brule Formations in Nebraska into members (see discussion in Terry, this volume; and LaGarry, this volume).

The members of the Chadron and Brule Formations in South Dakota have pronounced and observable differences in lithology with those of northwestern Nebraska, and therefore the descriptions of Clark (1937, 1954), Bump (1956), and Schultz and Stout (1955) are equally justified and valid based on respective lithologic differences. Such lithologic variability is not unique to the White River Group, nor is it a problem under the current North American Stratigraphic Code (NACSN, 1983). To the best of our knowledge, units assigned to the White River Group are mainly claystones, siltstones, and minor components that vary within limits acceptable for group-rank units as specified by the code. In our view, detailed correlations between the two regions can and should reveal the distribution of member-rank lithostratigraphic units. In any case, unless demonstrated to be lithologically continuous with previously defined lithostratigraphic units and subsequently synonymized, the White River Group (=Formation) of Meek and Hayden (1857); the Chadron and Brule Formations of Darton (1899); the Orella and Whitney Members of Schultz and Stout (1938, 1955); the Ahearn, Crazy Johnson, and Peanut Peak Members of Clark et al. (1967); and the Scenic and Poleslide Members of Bump (1956) are all valid and must be retained.

Type locality, stratotypes, boundary stratotypes, and stratotype descriptions

Lithologies that we classify as the Big Cottonwood Creek Member are well exposed across northwestern Nebraska. However, we have selected the continuous exposures near Toadstool Park (Figs. 3, 4) as the type area for the Big Cottonwood Creek Member. Exposures near Toadstool Park are also the type localities of the Orella and Whitney Members of the Brule Formation (Schultz and Stout, 1955; Fig 3, sections 4–9) and the location of several reference sections for the Chadron A, Chadron B, and Chadron C (Schultz and Stout, 1955; Fig. 3, sections 1–2). We have interpreted the provisions of the North American Stratigraphic Code (NACSN, 1983) regarding type sections to indicate that redescription of Schultz and Stout's (1955) "type" sections of the Chadron Formation would be preferable to erecting others elsewhere. Therefore, we designate part of Schultz and Stout's (1955; Fig. 3) section 2 along Big Cottonwood Creek as our stratotype (Fig. 4). The upper and lower contacts of the Big Cottonwood Creek Member vary, so we also designate two boundary stratotypes along the exposures at Toadstool Park (Fig. 4). Boundary stratotype 1 was previously described by Schultz and Stout (1955). However, our reinterpretation of the lithostratigraphy of the Chadron Formation requires us to erect a new boundary stratotype 2. For detailed descriptions of lithostratigraphic units above and below the Big Cottonwood Creek Member, see LaGarry (this volume) and Terry (this volume), respectively.

Stratotype. The stratotype of the Big Cottonwood Creek Member is located on the east-facing escarpment about 0.3 km northwest of Toadstool Park, Sioux County, Nebraska. Schultz and Stout (1955; Fig. 3, section 2) used this section as part of their composite type section (actually, 10 sections) for the Chadron Formation. The section begins in the bottom of the small gully below the channel exposure where Schultz and Stout (1955; Fig. 4) photographed their "third purplish-white layer," and continues up the cliff face to the west (Fig. 5A).

Observations. At the base of the section is 5 m of hummocky olive, bluish-gray, red, and yellow claystone (Figs. 4, 5A). The red and yellow color phases appear to be the product of diagenesis or pedogenesis, as there are no depositional contacts between them and the surrounding olive and bluish-gray claystones, nor is there a change in lithology. Within these claystones are infrequent, thin (<0.03 m) limestone interbeds of two distinct morphologies. The first type consists of either discontinuous layers of welded nodules or resistant, ledge-forming beds that pinch and swell laterally and commonly contain the fossilized egg chambers of insects and dung beetle balls. The second type consists of laterally continuous beds of uniform thickness containing algal laminae, gastropods, pelecypods, and invertebrate trace fossils. The bluish-green and olive claystones form a wedge-shaped intertongue within overlying silty claystones. Within the intertongue are several thin, lenticular, nodular-appearing, horizontally laminated, and cross-laminated grayish-brown sheet sandstones (Fig. 6C). These sandstone beds are 0.10 to 0.15 m thick and are usually separated by a similar thickness of claystone. The nodular appearance is likely the result of compaction and either lenticular or discontinuous bedding, rather than the presence of actual nodules or concretions. At the contact between the claystone and the overlying lithology is a 0.05- to 0.07-m-thick continuous limestone (3PW of Schultz and Stout, 1955), likely of lacustrine origin, containing low- and high-spired gastropods (Fig. 6C).

Overlying the intertongue of bluish-green and olive claystone is 28 to 30 m of variegated red, green, buff, orange, and olive laminated volcaniclastic silty claystones (Figs. 4, 5A). Within these silty claystones are limestone and volcanic ash interbeds. The limestones are similar to the two types described previously. The volcanic ashes are laterally continuous and occur at the middle and top of the sequence. The lower of the two ashes (lower [=second] purplish-white layer of Schultz and Stout, 1955) is 1.5 m thick and intermixed with the silty claystone. It has a distinctive gray or grayish-green color and is laterally continuous along the escarpment north of Toadstool Park (Fig. 5A). The upper of the two ashes (upper purplish-white layer of Schultz

and Stout, 1955) is thinner (about 1 m thick), whiter in color, and less intermixed with epiclastic sediment (Fig. 5A). At the top of this sequence are 3 to 4 m of cross-bedded, medium- to coarse-grained, multistoried brown sandstones.

Interpretations. Terry et al. (1995) referred to the bluish-green, gray, and olive claystone at the base of this section as the Peanut Peak Member of the Chadron Formation (see also Terry, this volume). They suggested that the hummocky, bluish-green, gray, and olive claystones in the Toadstool Park area, the Peanut Peak Member within the Red River Valley of Clark (1937, 1954), and the "undifferentiated Chadron Formation" of Clark et al. (1967) are lithostratigraphically equivalent.

The limestone at the contact between the intertongue of bluish-green, gray, and olive claystone and the overlying variegated silty claystone is Schultz and Stout's (1955) third purplish-white layer (Figs. 5A, 6C). Schultz and Stout (1955) classified the hummocky, bluish-green, gray, and olive claystones along with the overlying variegated silty claystones below the lower volcanic ash, the lower (=second) purplish-white layer, as their Chadron B. The overlying sequence of silty claystones up to the second volcanic ash, their upper purplish-white layer, was classified as their Chadron C. Schultz and Stout (1955) originally described these volcanic ashes as gypsiferous paleosols. Prior to our reinterpretation, Schultz and Stout's (1955) upper purplish-white layer was designated as the contact between the Chadron and Brule Formations (Fig. 5A). The variegated silty claystone above their upper purplish-white layer was their Orella A subdivision of the Orella Member of the Brule Formation.

The third purplish-white layer at the top of the hummocky, bluish-green, gray, and olive claystone intertongue does not mark a lithologic change from claystone to silty claystone. Instead, this lithologic change occurs at the base of the claystone intertongue (Fig. 5A), which is our contact between the Peanut Peak and Big Cottonwood Creek Member (see also Terry et al., 1995; Terry, this volume). Close examination reveals no lithologic difference between the variegated volcaniclastic silty claystones immediately above and below upper and lower purplish-whites (in agreement with Emry et al., 1987), and therefore we recognize the variegated silty claystones as a single lithostratigraphic unit, our Big Cottonwood Creek Member of the Chadron Formation. The 3- to 4-m-thick sequence of multistoried sandstones at the top of our stratotype is an erosional remnant of Schultz and Stout's (1955) "Toadstool Park channel complex," which is a channel fill within the Brule Formation. We measured 31 m of section at this outcrop, whereas Schultz and Stout (1955) reported 35.5 m of exposed bedrock.

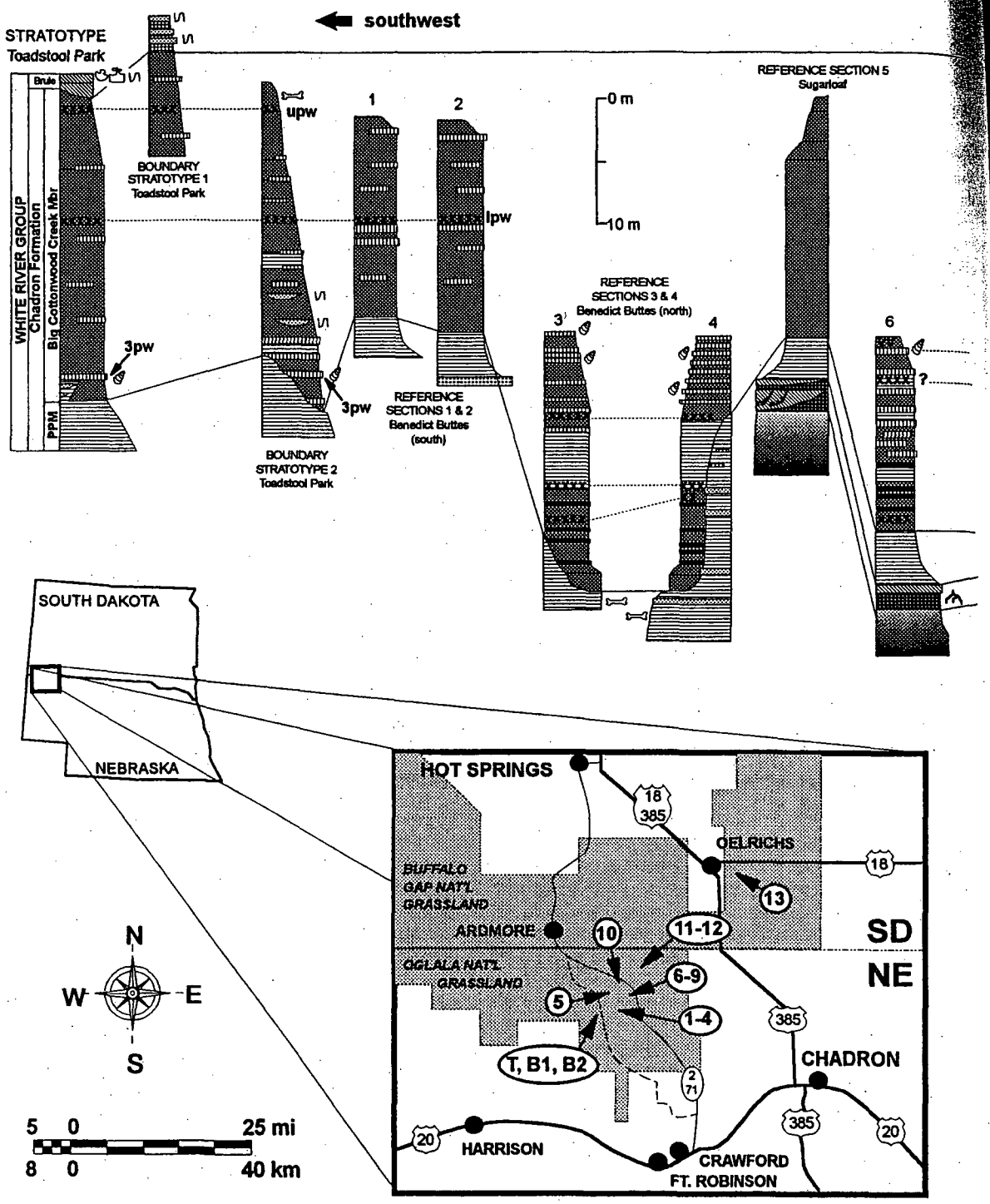
The intertonguing contact between the Peanut Peak and Big Cottonwood Creek Members is well demonstrated in the stratotype in the vertical cross section through the intertongue of interbedded bluish-green, gray, and olive claystone and sandstone. These sandstones are probably of fluvial origin (Fig. 6C). This large intertongue likely represents a remnant of the depositional system that produced the underlying bluish-green, gray, and olive claystones. The limestone interbeds within our stratotype for the Big Cottonwood Creek Member are both pedogenic and lacustrine varieties. The pedogenic limestones represent the downward translocation and precipitation of calcium carbonate to form distinctive paleosol horizons (Terry, 1995). Evans and Welzenbach (this volume) suggest that the lacustrine limestones of the Chadron Formation formed in springfed lakes. In addition to the gastropods observed here, lacustrine limestones within the Big Cottonwood Creek Member also contain algal laminations, ostracodes, fish bones, and charophytes (Evans and Welzenbach, this volume).

Boundary stratotype 1. This section is located along the east-facing escarpment about 20 m south of the Toadstool Park visitor trail, and about 0.4 km southeast of the stratotype (Figs. 3, 4). Schultz and Stout (1955; Fig. 3, section 6) also used this section as part of their composite type section for the Chadron Formation. The section begins at the base of the escarpment and continues upward (Fig. 5B). Although not figured, part of this sequence continues up the escarpment for another 5 to 7 m (see LaGarry, this volume).

Observations. This 15-m-thick section consists of a sequence of red, green, and buff laminated volcaniclastic silty claystones. Within these silty claystones are interbeds of limestone and volcanic ash. The limestones are thin (<0.05 m), laterally discontinuous, and pinch and swell as if composed of welded nodules. Nine meters above the base of the section is a prominent volcanic ash (UPW = upper purplish-white layer) intermixed with various amounts of epiclastic sediment (Fig. 5B). The ash zone is white or gray, about 0.75 m thick, and laterally continuous (but offset by faulting) with the upper of the two ashes described in the stratotype. About 6 m above the prominent volcanic ash the silty claystones intertongue with an overlying sequence of thinly interbedded brown sandstones and tan, light brown, brown, and brownish-orange laminated volcaniclastic clayey siltstones. This lithologic contact is clearly visible in outcrop (Fig. 5B). The thin sandstones occur about 10 m above the ash (C in Fig. 5B). At this level there is a 4-m-thick zone of interbedded sandstones and clayey siltstones.

Interpretations. Schultz and Stout (1955) placed the contact between the Chadron and Brule Formations at the upper purplish-white layer (Fig. 5B), and referred to the lowermost of the interbedded sandstones as their lower nodules. The strata between the upper purplish-white layer (UPW) and the lower nodules (C) shown in boundary stratotype 1 was previously classified as the Orella A (Fig. 5B). However, based on a strictly lithologic interpretation, this interval cannot be placed within the Brule Formation. We follow the suggestion of Terry et al. (1995) and formally place the contact of the Chadron and Brule Formations at the lithologic change from brown interbedded sheet sandstones and tan, light brown, brown, and brownish-orange volcaniclastic clayey siltstones to greenish and reddish volcaniclastic silty claystones.

Based on our lithostratigraphic reinterpretation, this section shows an intertonguing contact between the volcaniclastic silty claystones of the Chadron Formation and the interbedded sand-


southwest
STRATOTYPE
Toadstool Park
Brule
WHITE RIVER GROUP
Chadron Formation
Big Cottonwood Creek Mbr
PPM
3pw
BOUNDARY
STRATOTYPE 1
Toadstool Park
upw
3pw
BOUNDARY
STRATOTYPE 2
Toadstool Park
1
2
lpw
REFERENCE
SECTIONS 1 & 2
Benedict Buttes
(south)
0 m
10 m
REFERENCE
SECTIONS 3 & 4
Benedict Buttes (north)
3'
4
REFERENCE SECTION 5
Sugarloaf
6
?
SOUTH DAKOTA
NEBRASKA
N
W
E
S
5 0 25 mi
8 0 40 km
HOT SPRINGS
18
385
OELRICHS
18
13
BUFFALO
GAP NAT'L
GRASSLAND
ARDMORE
10
11-12
SD
NE
OGLALA NAT'L
GRASSLAND
6-9
5
1-4
385
T, B1, B2
CHADRON
2
71
385
20
20
HARRISON
CRAWFORD
FT. ROBINSON

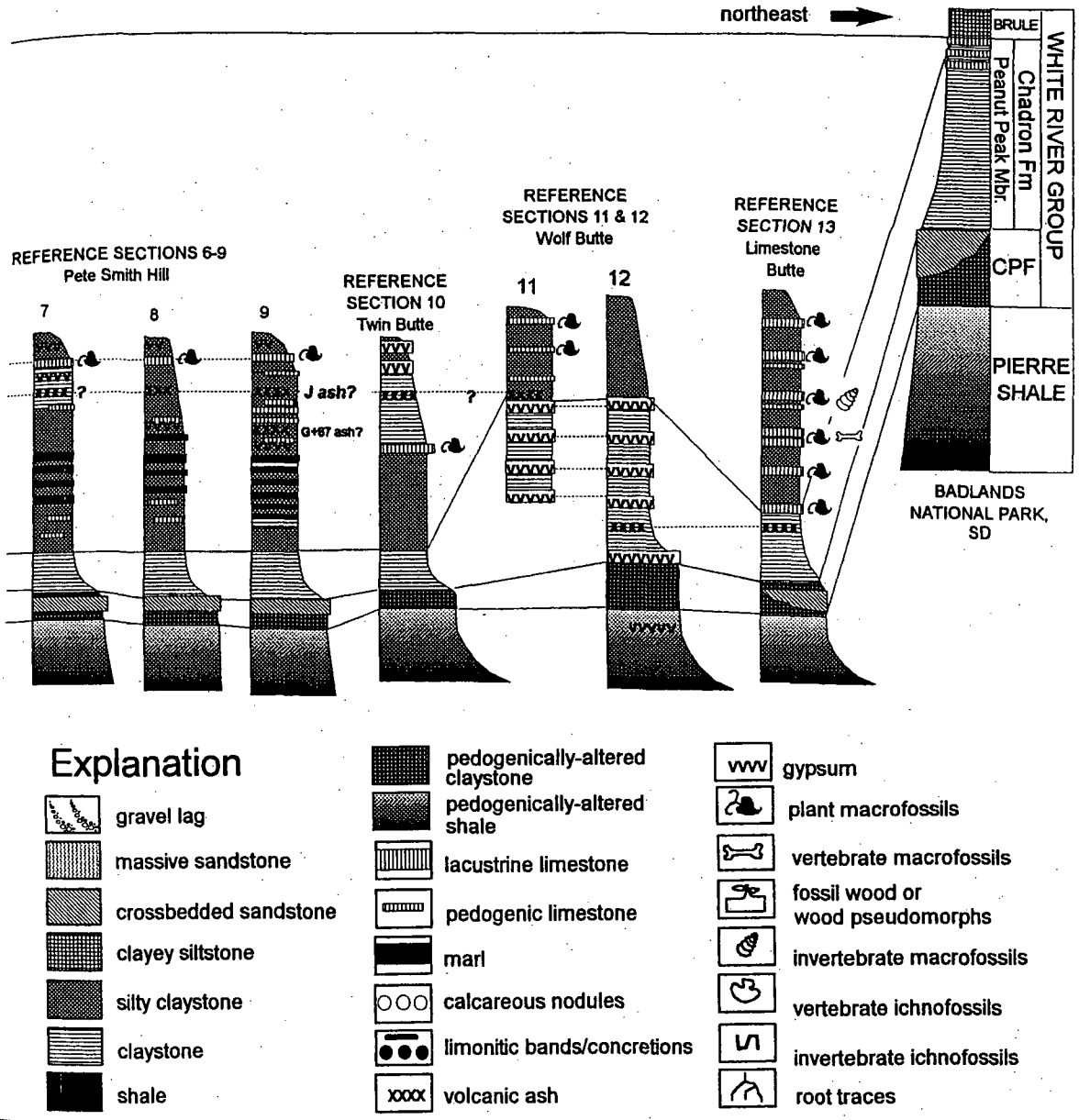


Figure 4. The Big Cottonwood Creek Member of the Chadron Formation in northwestern Nebraska and southwestern South Dakota. Section locations are as follows: Stratotype: NE1/4SW1/4SW1/4SE1/4 sec. 5, T. 33 N., R. 53 W., Sioux County, Nebraska. Boundary stratotype 1: NE1/4SE1/4NW1/4SW1/4 sec. 4, T. 33 N., R. 53 W., Sioux County, Nebraska. Boundary stratotype 2: SE1/4NW1/4NE1/4NE1/4 sec. 8, T. 33 N., R. 53 W., Sioux County, Nebraska. Reference section 1: SE1/4SW1/4NE1/4NE1/4 sec. 29, T. 33 N., R. 52 W., Dawes County, Nebraska. Reference section 2: NE1/4SW1/4NE1/4NE1/4 sec. 29, T. 33 N., R. 52 W., Dawes County, Nebraska. Reference sections 3 and 4: NW1/4NE1/4 sec. 20, T. 33 N., R. 52 W., Dawes County, Nebraska. Reference section 5: NW1/4SE1/4NW1/4 sec. 27, T. 34 N., R. 53 W., Dawes County, Nebraska. Reference section 6: W1/2NW1/4 sec. 20, T. 34 N., R. 52 W., Dawes County, Nebraska. Reference section 7: NE1/4NE1/4 sec. 20, T. 34 N., R. 52 W., Dawes County, Nebraska. Reference section 8: SW1/4SE1/4 sec. 17, T. 34 N., R. 52 W., Dawes County, Nebraska. Reference section 9: SE1/4SE1/4 sec. 17, T. 34 N., R. 52 W., Dawes County, Nebraska. Reference section 10: SW1/4NW1/4 sec. 7, T. 34 N., R. 52 W., Dawes County, Nebraska. Reference section 11: E1/2NW1/4SW1/4 sec. 34, T. 35 N., R. 52 W., Dawes County, Nebraska. Reference section 12: W1/2NE1/4SE1/4 sec. 33, T. 35 N., R. 52 W., Dawes County, Nebraska. Reference section 13: S1/2SW1/4 sec. 16, T. 10 S., R. 8 E., Fall River County, South Dakota. See text for descriptions of measured sections. PPM = Peanut Peak Member of the Chadron Formation; CPF = Chamberlain Pass Formation of Terry (this volume). Other abbreviations as in Figure 2. Lithologic symbols in explanation are used throughout this chapter.

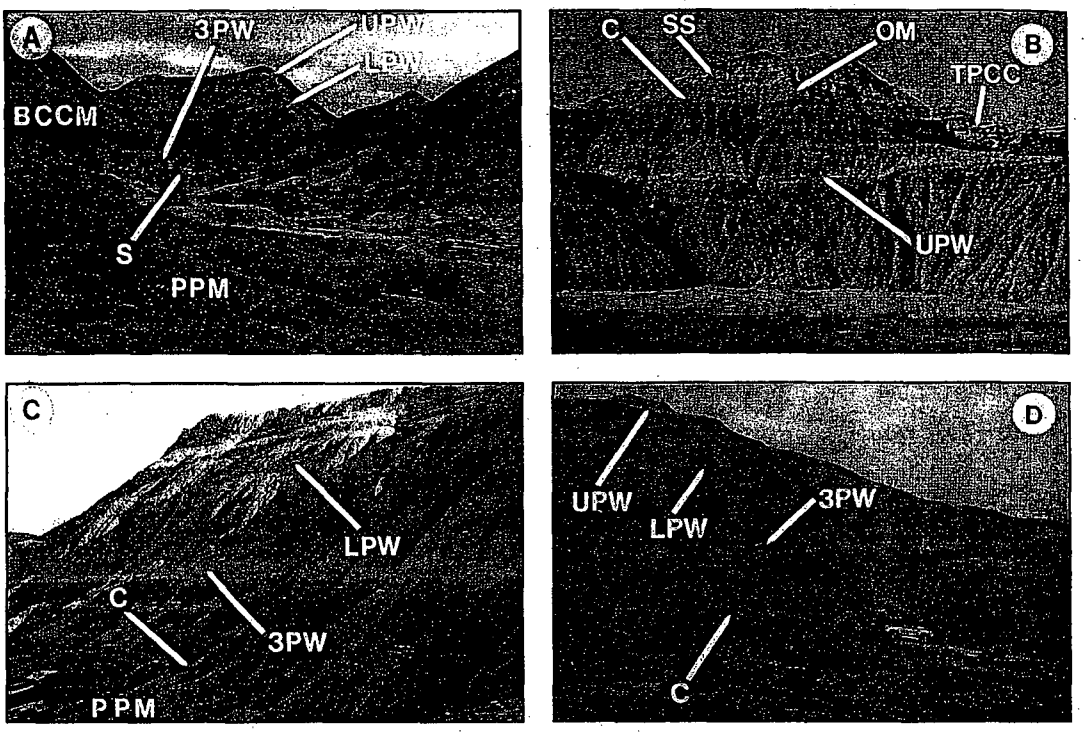


Figure 5. Stratotype and boundary stratotypes of the Big Cottonwood Creek Member (see Fig. 3) at Toadstool Park. A, Stratotype section, part of Schultz and Stout's (1955; Fig. 3, section 2) composite "type" section of the Chadron Formation, showing intertonguing contact (dashed line) of Big Cottonwood Creek (BCCM) and Peanut Peak (PPM) Members. B, Boundary stratotype 1 showing intertonguing contact (C) of Big Cottonwood Creek Member and Orella Member (OM) of the Brule Formation (sensu LaGarry, this volume). C and D, Big Cottonwood Creek Member in boundary stratotype 2. C = Lithologic contact between the Peanut Peak and Big Cottonwood Creek Members in C and D, and the Chadron and Brule Formations in B. S = Lenticular and nodular sandstone bodies at the contact of the Peanut Peak and Big Cottonwood Creek Members; SS = sheet sands of Schultz and Stout's (1955) lower nodules. TPCC = Toadstool Park channel complex of Schultz and Stout (1955). Other abbreviations as in previous figures.

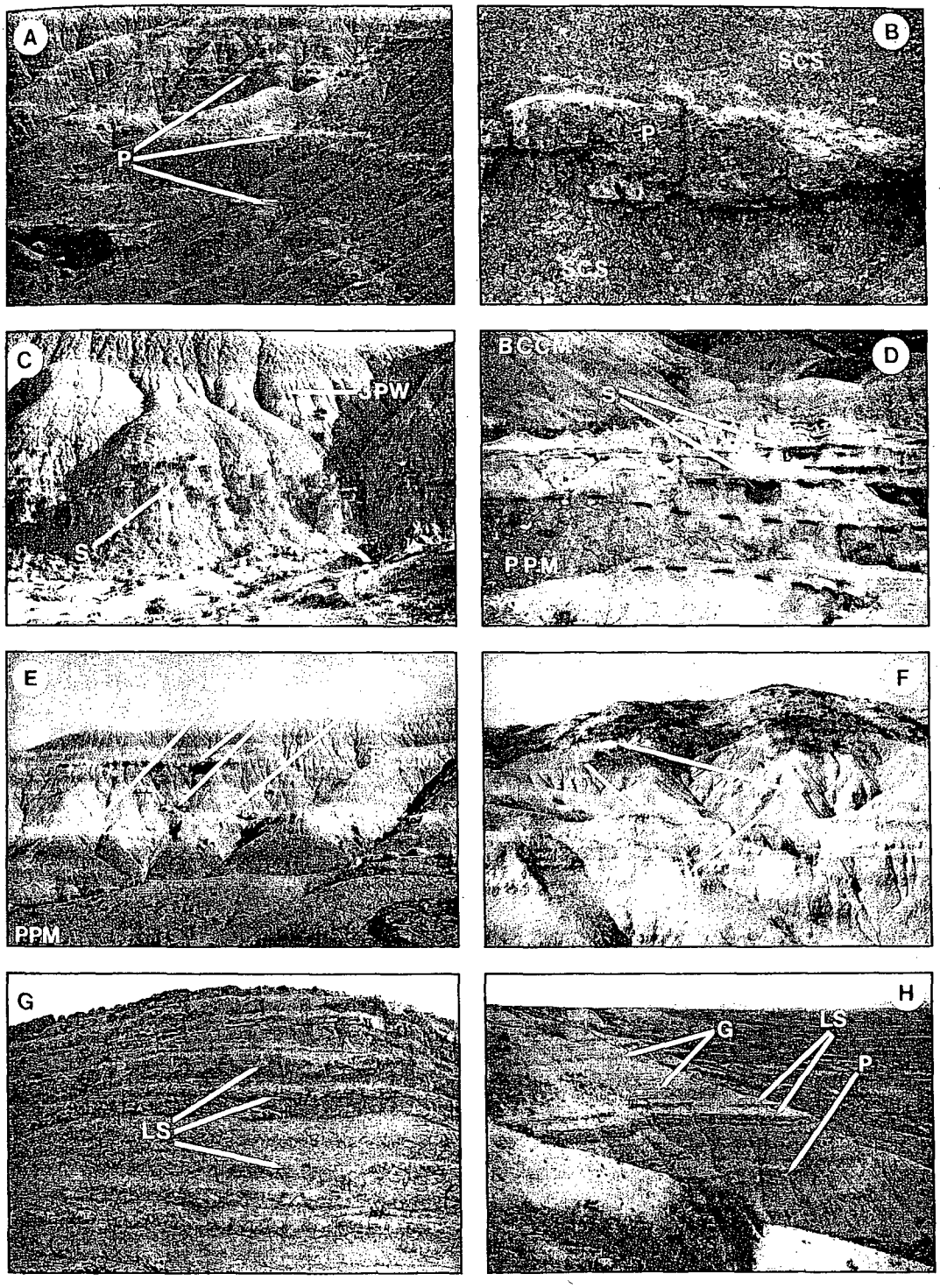
stones and volcaniclastic clayey siltstones of the Orella Member of the Brule Formation. In the Toadstool Park area, the "Toadstool Park channel complex" of Schultz and Stout (1955) cuts through this intertonguing contact into the underlying Big Cottonwood Creek Member of the Chadron Formation (see Terry et al., 1995, and the discussion in the following section).

Boundary stratotype 2. This section was measured at the east end of the outlying butte north of the Toadstool Park access road (Figs. 3, 4, 5C,D). This section is about 0.8 km east of the stratotype, and was previously described by Terry et al. (1995; Fig. 3E).

Observations. At the base of the section are 3 to 6 m of hummocky, bluish-green, gray, and olive claystones (Fig. 5D). These claystones appear to have been eroded to form a shallow paleovalley into which the strata of the overlying Big Cottonwood Creek Member were deposited. On the flanks of the paleovalley wall, the claystones are red and yellow. As in the stratotype, there are no depositional contacts between the red and yellow zones and the surrounding olive and bluish-gray claystones. Aside from color, the yellow, red, olive, and bluish-gray phases are lithologically identical. Within these claystones are lacustrine limestone and pedogenic calcrete interbeds identical to those in the stratotype (Fig. 6A,B,E).

The shallow paleovalley incised into the underlying hummocky claystones was apparently filled and overtopped by 25 m of interbedded volcaniclastic silty claystones, volcaniclastic claystones, lenticular sandstones, thin lacustrine limestones, pedo-

Figure 6. Selected lithologic features and interbeds within the Big Cottonwood Creek Member. A and B, Boundary stratotype 2; C, 75 m east of stratotype; D, along Whitehead Creek, 3 km northwest of stratotype; E, boundary stratotype 2; F, 20 m south of reference section 3; G, reference section 4; H, reference section 7. G = gypsum; P = pedogenic calcrete; LS = lacustrine limestone; S = sandstone; SCS = silty claystone; V = volcanic ash. Other abbreviations as in previous figures.


A
P
B
SGS
P
SGS
C
PW
S
D
S
PPM
E
PPM
F
G
LS
H
G
LS
P

genic calcretes, and volcanic ashes (Figs. 5C,D; 6A,E). The volcaniclastic silty claystones are laminated and red, green, buff, orange, and olive in color. The volcaniclastic claystones are also laminated and are green, brown, and orange. The volcaniclastic silty claystones are identical to those described in the stratotype. The lenticular sandstones are brown or grayish-brown, and appear massive. This lack of internal structure may be the result of bioturbation, as many of these sandstones contain abundant invertebrate burrows. To the west of Toadstool Park, cross-bedded, green tabular sandstones are present at the base of this sequence (Fig. 6D). The limestone interbeds within the variegated claystones and silty claystones are tabular or lenticular, and identical to those in the stratotype. The volcanic ashes are laterally continuous along the butte and occur at the same stratigraphic position as in the stratotype (Fig. 5D). The lower of the two ashes is 1.5 m thick and intermixed with silty claystone. It has a distinctive gray or grayish-green color. The upper of the two ashes is thinner (about 1 m thick) and is less intermixed with epiclastic sediment. Although the entire sequence contains vertebrate fossils, there is a local concentration above the upper volcanic ash (LaGarry-Guyon and Hunt, 1993). These volcanic ashes are identical in all other respects to those described in the stratotype (Figs. 4, 5A).

Interpretations. As in the stratotype, the olive, bluish-gray, red, and yellow claystones, previously assigned by Schultz and Stout (1955) to their Chadron B, have been classified by Terry et al. (1995) and Terry (this volume) as the Peanut Peak Member of the Chadron Formation. Unlike the stratotype, however, this section shows an erosional contact between the claystones of the Peanut Peak Member and the overlying lithology. The 25 m of variegated claystone and silty claystone correspond to Schultz and Stout's (1955) remaining Chadron B, Chadron C, and Orella A. As in the stratotype, the prominent limestone and the lower and upper ashes correspond to Schultz and Stout's (1955) third purplish-white layer, lower purplish-white layer, and upper purplish-white layer (Figs. 4, 5A–D). As in the stratotype, there is no demonstrable lithologic difference between any of these subdivisions, and combined they constitute our Big Cottonwood Creek Member of the Chadron Formation.

Reference sections and boundary descriptions

Outside the type area, interbeds within the clayey siltstones of the Big Cottonwood Creek Member are quite variable. Therefore, we recognize and describe 13 reference sections illustrating the geographic and stratigraphic variations within this unit. At the beginning of each discussion we review previous workers' classifications (if any) of strata within the referenced sections, and follow with a discussion of the relevant lithologic differences between the reference sections and our stratotypes. For bed-by-bed descriptions of these reference sections, see LaGarry and LaGarry (1997).

Reference sections 1 and 2. Reference sections 1 and 2 were measured at the north and south ends, respectively, of Benedict Buttes, 11 km southeast of Toadstool Park in Dawes County, Nebraska (Fig. 4).

Remarks. The significant differences between these exposures and those in the type area are that the silty claystones in these sections are more red and less green in color and lacustrine limestones are absent. Pedogenic calcretes as described within the stratotype are present, however. The increased red coloration in these sections is likely the result of better drained and more oxidizing paleoenvironments in which the paleosols formed. The ash horizon within these sections is about 1.5 m thick, is highly intermixed with silt and clay, and likely corresponds to the lower of the two ashes within the stratotype (Figs. 4, 5A, 7A).

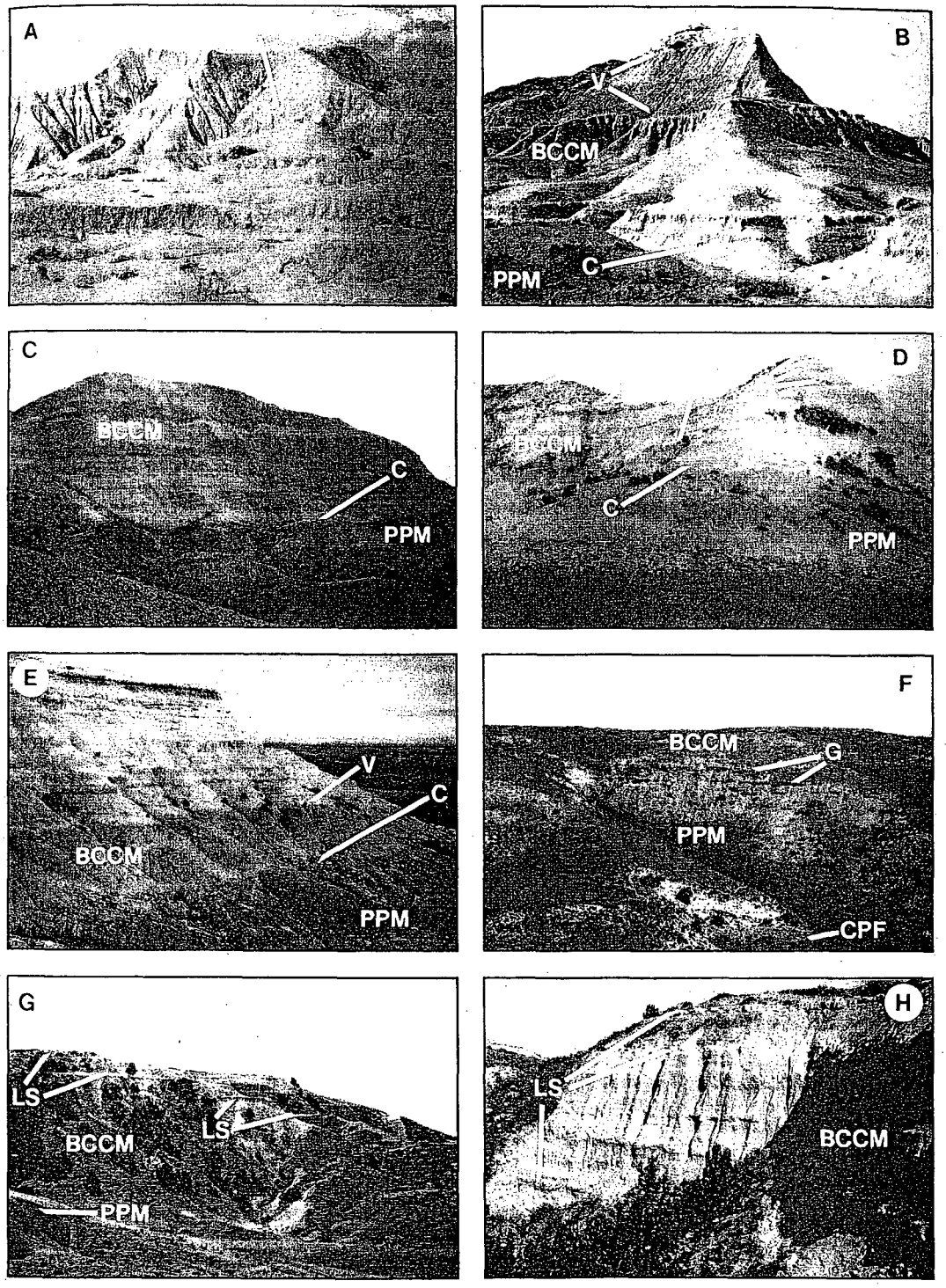
Reference sections 3 and 4. These sections are located 1.6 km north of reference sections 1 and 2 (Figs. 4, 8). Schultz and Stout (1955; Fig. 3, section 10) previously measured reference section 3 as part of their composite type section for the Chadron Formation. They assigned these beds to the upper Chadron B and Chadron C, and recognized the upper and lower ashes (which they identified as gypsiferous paleosols) as their upper and lower purplish-white layers, respectively. However, because of lithologic differences between these sections and those in the type area, we argue that these correlations are unjustified.

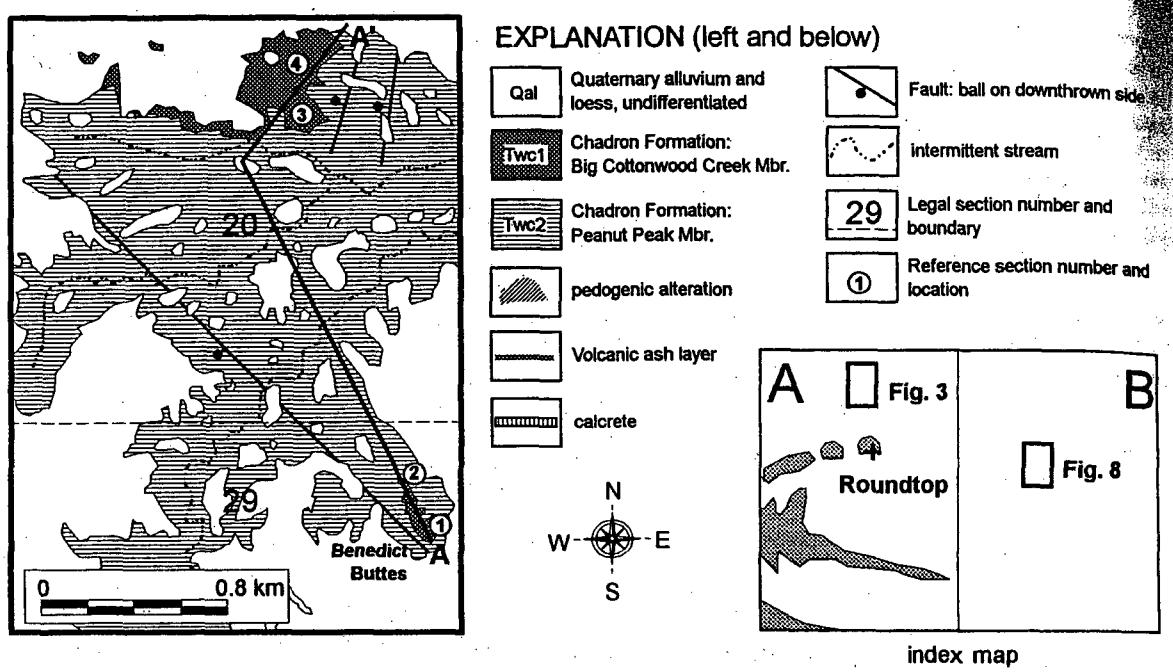
Remarks. The pedogenic modification that produced the variegated coloration in reference sections 1 and 2 is absent in reference sections 3 and 4 (Figs. 4, 7B). In addition, lacustrine limestones are far more abundant and pedogenic calcretes are rare (Fig. 6G). The differences among these sections and those previously described may be the result of synchronous deposition in markedly different paleoenvironments (conditions here were wetter and less oxidizing). However, several lines of evidence suggest that the differences are stratigraphic rather than geographic.

We believe that reference sections 3 and 4 occur stratigraphically below reference sections 1 and 2, and those in the type area, based on the current topography, interpretation of the paleotopography, and recent geologic mapping. It should be noted that topographic and geologic maps were not available to Schultz and Stout (1955). Strata of the White River Group in the Toadstool Park area are essentially flat lying and exposed in a semi-circular outcrop belt with Roundtop, a prominent hilltop within the study area, at the center. The farther away from Roundtop the rocks are exposed, the greater their dip, up to about 7° (LaGarry and LaGarry, 1997). Therefore, unless offset by faulting, strata can be considered flat lying over distances of less than 2 or 3 km.

Geologic mapping of the Horn 7.5′ Quadrangle (LaGarry and LaGarry, 1997) showed that the bases of reference sections 1 and 2 occur at or near the 3,700-ft (1,128 m) contour, whereas the bases of reference sections 3 and 4 occur at or near the

Figure 7. Reference sections for the Big Cottonwood Creek Member described in this report. A, Reference section 2, Benedict Buttes; B, reference section 3, Benedict Buttes; C, reference section 5, Sugarloaf; D and E, reference section 6, Pete Smith Hill; F, reference section 11, Wolf Butte; G and H, reference section 13, Limestone Butte, near Oelrichs, South Dakota. G = gypsum. CPF = Chamberlain Pass Formation of Terry (this volume), M = marl. Other abbreviations as in previous figures.


A
B
V
BCCM
C
PPM
C
BCCM
C
PPM
D
BCCM
C
PPM
E
V
C
BCCM
PPM
F
BCCM
G
PPM
CPF
G
LS
LS
BCCM
PPM
H
LS
BCCM



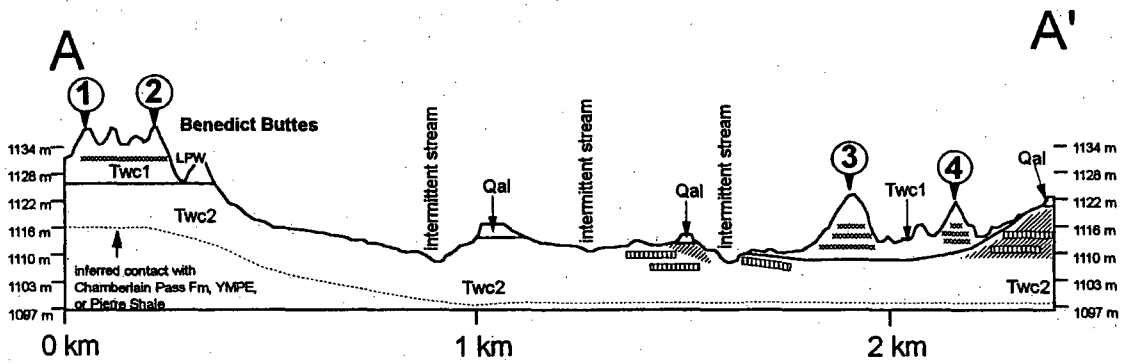


Figure 8. Hypothesized stratigraphic relationships between upland (reference sections 1 and 2) and lowland (reference sections 3 and 4) phases of the Big Cottonwood Creek Member exposed in the vicinity of Benedict Buttes, Dawes County, Nebraska. Index map showing locations of geologic maps used in this report (shaded area indicates Pine Ridge escarpment): A, Roundtop 7.5′ Quadrangle, Sioux and Dawes Counties, Nebraska; B, Horn 7.5′ Quadrangle, Dawes County, Nebraska. YPME = Yellow Mounds Paleosol Equivalent of Terry (this volume). Geologic map data after LaGarry and LaGarry (1997).

3,640-ft (1,110 m) contour. Assuming a maximum dip of 7°, the bases of reference sections 3 and 4 should occur at or near the 3,700-ft (1,128 m) contour (Fig. 8). Either the beds in reference sections 3 and 4 are offset downward 60 ft (18 m) by faulting with respect to reference sections 1 and 2, or the beds in reference sections 3 and 4 occur stratigraphically below those in reference sections 1 and 2. LaGarry and LaGarry (1997) recognized several faults in the area (Fig. 8). However, these faults occur parallel to the head of the canyon where reference sections 3 and 4 were measured and do not have sufficient displacement to account for the topographic offset between reference sections 1 and 2 and reference sections 3 and 4.

The paleotopographic evidence is based on the geometry of the upper contact of the Peanut Peak Member of the Chadron Formation in reference sections 3 and 4. In reference section 3, this contact slopes upward to the southwest and eventually occurs

5 m higher in the canyon wall 30 m away. In reference section 4, this contact slopes upward to the north and is 20 m higher in the canyon wall 20 m to the north. On the upper walls of this paleovalley, the Peanut Peak Member is predominately yellow, red, lavender, and orange, rather than the typical olive and bluish-gray, on the upper paleovalley walls. This change in color is likely due to better drained and more oxidizing soil conditions in the upper portions of the paleovalley. In addition to changes in color, parts of the paleovalley wall display pedogenic calcretes, some of which are roughly parallel to the paleovalley slope (Fig. 8). The topographic and paleopedologic evidence, along with the geometry of this contact, suggests that the Big Cottonwood Creek Member in reference sections 3 and 4 was deposited in a 20-m-deep paleovalley within the underlying Peanut Peak Member. Since these sections occur stratigraphically lower than previously thought (cf. Schultz and Stout, 1955), they may represent a previously unrecognized depositional interval of the Chadron Formation. However, this interval is of the same lithology as the stratotype and is included within the Big Cottonwood Creek Member. Swinehart et al. (1985) recognized the presence of paleovalley systems within the Chadron Formation using subsurface data. The paleovalley exposures at Benedict Buttes, and others in following discussions, may represent a part of these paleovalley systems.

These conclusions are supported by our observations in the type area of the Big Cottonwood Creek Member. In the stratotype (Figs. 4, 5A), the contact between the Big Cottonwood Creek Member and the underlying Peanut Peak Member is intertonguing. However, in boundary stratotype 2 (Figs. 4, 5C–D), the contact is erosional and the Big Cottonwood Creek Member occupies a shallow paleovalley within the Peanut Peak Member. The north wall of the paleovalley in boundary stratotype.2 shows red and yellow color phases of the Peanut Peak Member as opposed to the typical olive. The red and yellow color phases probably indicate oxidation during pedogenic alteration of well-drained, upland soils. Based on our observations, we interpret the base of the stratotype as representing a continuous transition, without evident hiatus, from paleoenvironments of the Peanut Peak to Big Cottonwood Creek Members. Boundary stratotype 2 represents a shallow paleovalley in which pedogenesis along the upper paleovalley slopes proceeded while sedimentation was interrupted prior to the deposition of the Big Cottonwood Creek Member. Reference sections 3 and 4 were probably deposited in an older, deeper paleovalley that was filled prior to deposition of the Big Cottonwood Creek Member on the surrounding uplands (sections in the type area and reference sections 1 and 2).

Reference section 5. This section was measured at Sugarloaf, a prominent butte 5 km northeast of Toadstool Park, in Dawes County, Nebraska (Figs. 4, 7C). Schultz and Stout (1955; Fig. 3, section 12) assigned the sandstones at the base of the section to the Chadron A (=Chamberlain Pass Formation of Terry, this volume), the next 31 m to the Chadron B, and the next 10.6 m to the Chadron C. They measured 10.5 m from the base of the Chadron B to the "4th purplish-white layer" (a gypsum), 12.8 m between the 4th and 3rd purplish-white layers (a bedding plane), and 7.9 m to the lower purplish-white layer (a disconformity). Vondra (1958; Plate IX, section 29) did not recognize the sandstones at the base of the section, but described them as limestones within the base of the Chadron B. Vondra (1958) recognized the next 33.2 m as Chadron B, followed by 8.5 m of Chadron C. He measured 17.9 m from the base of the Chadron B to the 4th purplish-white layer (a pink or white siltstone), 6.4 m between the 4th and 3rd purplish-white layers (a bedding plane), and 8.8 m to the lower purplish-white layer (a disconformity).

Remarks. Terry (this volume) assigns the bluish-green, gray, and olive hummocky claystones at the base of Schultz and Stout's (1955) and Vondra's (1958) Chadron B to the Peanut Peak Member of the Chadron Formation. We assign the remainder to the Big Cottonwood Creek Member of the Chadron Formation. In addition to the disparities between Schultz and Stout's (1955) and Vondra's (1958) measured thicknesses between the various purplish-white layers, they described these beds as different lithologies. This suggests to us that recognition of these marker beds at this locality is either prohibitively difficult or totally arbitrary. We do not recognize any marker beds that can be unambiguously correlated to any purplish-white layer or to any other bed described in the Toadstool Park vicinity by Schultz and Stout (1955). Because of the difficulties in recognizing purplish-white layers or their equivalents, we suggest that their use in correlation, and as marker beds, be abandoned.

Reference sections 6, 7, 8, and 9. These sections were measured along Pete Smith Hill, 11 km northeast of Toadstool Park, in Dawes County, Nebraska (Fig. 4). Schultz and Stout (1955; Fig. 3, sections 13–16) were the first to describe sections at this locality. Schultz and Stout (1955; Fig. 3, section 16) recognized the lower ash in reference section 6 as their third purplish-white layer, and the white layer near the top of reference section 6 as their lower (=second) purplish-white layer (Fig. 7D,E). Based on their recognition of these marker beds, they assigned the hummocky, bluish-green, gray, and olive claystones to their "Chadron B^3," and the green silty claystones to their "Chadron B^4" and Chadron C. Based on these unit designations, they correlated these beds directly to the Chadron Formation exposed at Toadstool Park (Fig. 5A–D). Subsequently, Vondra (1958; Plate VII, section 19) assigned the hummocky, bluish-green, gray, and olive claystones to the lower Chadron, the green silty claystones below the algal limestone (his lower purplish-white) to the middle Chadron, and the remainder to the upper Chadron. Vondra (1958) recognized two ashes in his measured section (probably <0.2 km east of reference section 9), but considered the overlying algal limestone to represent the lower purplish-white layer, and the ashes to represent the third purplish-white layer. Vondra's (1958) ashes were figured as occurring only centimeters apart, and are probably the same ashes in the upper part of reference section 9 (Fig. 4).

Remarks. These beds are predominantly laminated volcaniclastic silty claystones and interbedded lacustrine limestone, pedogenic calcrete, marl, gypsum, and volcanic ash (Fig. 7D,E), which is consistent with the lithology of the Big Cottonwood

Creek Member in the type area. Within these sections is the first occurrence of repeating sequences of green silty claystone, red silty claystone, and limestone that LaGarry and LaGarry (1997) refer to as "triplets." These sections may represent part of the same paleovalley interval of the Big Cottonwood Creek Member described in reference sections 3 and 4. This interpretation is based on the lithologic similarity of these beds to those north of Benedict Buttes (Fig. 7B), the geometry of the contact between the Peanut Peak and Big Cottonwood Creek Members, and unpublished analyses of volcanic ashes in reference sections 6 through 9. A discussion of these observations follows.

We interpret the presence of thin interbeds of laminated olive volcaniclastic claystone to represent small, relict depositional environments in which Peanut Peak–like deposition continued. These beds become less frequent higher in the Big Cottonwood Creek Member at this locality as they are replaced by beds of silty claystone. The silty claystone beds at this locality show subdued paleosol coloration as in reference sections 3 and 4, as opposed to the prominently colored paleosols in the stratotype, boundary stratotypes, and reference sections 1 and 2. Thus, the strata in reference sections 6 through 9 likely represent deposition in lowland environments that were wetter than the better drained, more upland environments represented by sections in the type area (cf. Figs. 7D,E with 5A–D).

When viewed from a distance, the prominent limestones and the lowest volcanic ash of the Big Cottonwood Creek Member appear lenticular, with their thickest point at the base of reference section 6 (Figs. 4, 7D,E). The contact between the Peanut Peak and Big Cottonwood Creek Members slopes upward 5° to 7° to the east and west of reference section 6. The lenticular bed geometry, along with the slope of the contact between the lower and upper members of the Chadron Formation, suggests that beds that we classify as the Big Cottonwood Creek Member were deposited in a local basin.

Mineralogic analyses of the upper ash in reference section 6 through 9 (Fig. 3) by Larson (unpublished data) suggest that it correlates to the "J ash" in the Flagstaff Rim area of Wyoming as described by Emry (1973) and the "3c tuff" in the Douglas, Wyoming area (Fig. 9) described by Evanoff et al. (1992). According to Evanoff et al. (1992), the upper purplish-white layer ("persistent white layer," or PWL) correlates to the "5 tuff" at Douglas. If correct, these correlations make both ashes in reference sections 6 through 9 older than the upper and lower purplish-white layers at Toadstool Park, and thus within an older paleovalley depositional sequence of the Big Cottonwood Creek Member. According to Larson and Evanoff (this volume and personal communication to Terry, 1996), however, new analyses suggest that the upper purplish-white layer (persistent white layer, or PWL in Wyoming) is not the same as the 5 tuff at Douglas, but is instead better correlated to the 3c tuff at Douglas and J ash at Flagstaff Rim. The correlation of the upper purplish-white layer at Toadstool Park to tuffs in Douglas and Flagstaff Rim is problematic due to the small grain size of the upper purplish-white layer. If these analyses are correct, the upper ash in reference sections 6 through 9 correlates to the upper purplish-white layer at Toadstool Park, and suggests that those reference sections may not belong to an older paleovalley-fill sequence of the Big Cottonwood Creek Member. See Larson and Evanoff (this volume) for detailed descriptions of the methods used for analyses of ashes within the White River Group.

Reference section 10. This is a composite section measured at Twin Buttes (Fig. 4), 11 km northeast of Toadstool Park, in Dawes County, Nebraska.

Remarks. These beds are lithologically similar to those described in reference sections 3 and 4 and 6 through 9 (Figs. 4, 7A,B,D,E), and we likewise interpret these beds to represent the lowland depositional interval of the Big Cottonwood Creek Member, with olive claystones and silty claystones within the Big Cottonwood Creek Member representing relict paleoenvironments of the Peanut Peak Member. As there is no direct evidence of deposition in a paleovalley (see LaGarry and LaGarry, 1997), these olive claystones and silty claystones may simply represent the intertonguing contact between the Big Cottonwood Creek and Peanut Peak Members of the Chadron Formation.

An exception to the previously described similarities is the presence of the prominent gypsum beds in reference section 10. The gypsum in these beds consists of coarse-grained, fibrous, matrix-supported granules. Schultz and Stout (1955) and Vondra (1958) reported gypsum interbeds within this stratigraphic interval of the Chadron Formation in exposures to the south and east. Much of the gypsum Schultz and Stout (1955) described was probably diagenetic. We did not recognize the gypsum to the southwest, in contrast to Schultz and Stout's (1955) measured sections. The gypsum present in this section is unmistakable, however, as the beds are thick (0.2–0.4 m), and are prominent ledge formers.

Reference sections 11 and 12. These sections were measured on the northeastern flank of Wolf Butte, 17.5 km northeast of Toadstool Park, in Dawes County, Nebraska (Fig. 4). Schultz and Stout (1955) and Vondra (1958) described sections on the southern flank of Wolf Butte 5 to 10 km south of reference sections 11 and 12. Because of the distance between our sections and theirs, we do not make detailed comparisons. However, Schultz and Stout (1955) and Vondra (1958) consistently correlated their measured sections to the interval exposed at Toadstool Park as they did at Benedict Buttes and Pete Smith Hill.

Remarks. The thick gypsum beds seen in reference section 10 are more abundant in reference sections 11 and 12 in the Peanut Peak Member (Fig. 7F). Along with the limestones within the Big Cottonwood Creek Member, these gypsum beds may indicate the presence of a local depocenter or depression more extensive than at previous reference sections. Also more prevalent within these sections are green silty claystone, red silty claystone, and limestone triplets (LaGarry and LaGarry, 1997), suggesting that the depositional setting first seen in reference sections 6 through 9 may be more prevalent to the northeast (see the following discussion of reference section 13). Based on the lithologies present in reference sections 11 and 12, we interpret

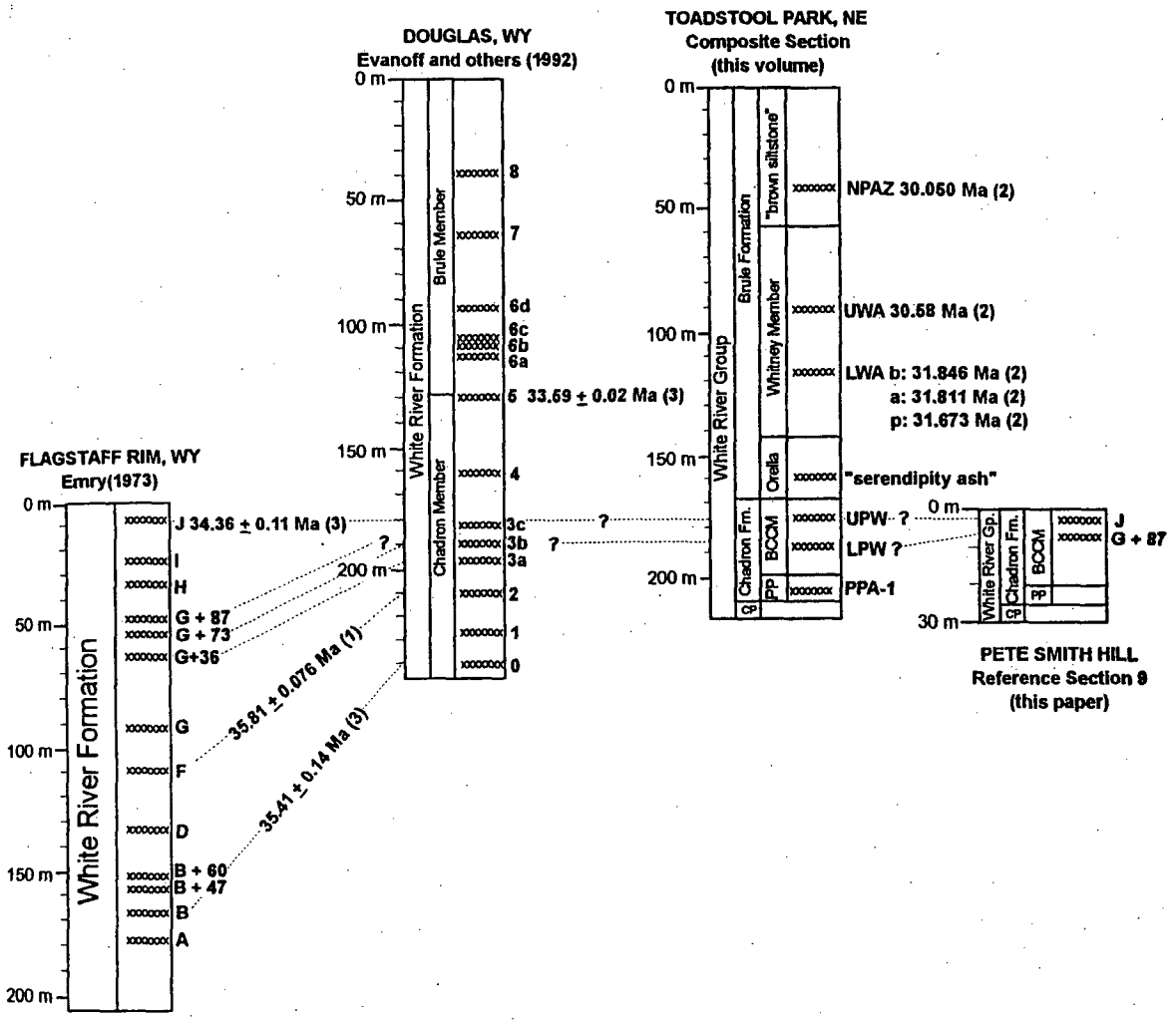


Figure 9. Correlation of volcanic ashes (tuffs) within the White River Group between Flagstaff Rim and Douglas, Wyoming, and Toadstool Park and reference section 9 at Pete Smith Hill in northwestern Nebraska (Fig. 4). The J ash (Flagstaff Rim) and 5 tuff (Douglas) are datums. Ash dates from: (1) Evanoff (1994, written communication to D. Terry); (2) Prothero and Swisher (1992): a = anorthoclase, b = biotite, p = plagioclase; (3) Obradovich et al. (1995). Pete Smith Hill ash correlations by E. E. Larson (unpublished data). Scale of Pete Smith Hill section increased for clarity. NPAZ = Nonpareil ash zone of Swinehart et al. (1985) and Tedford et al. (1996). UWA = upper Whitney ash; LWA = lower Whitney ash; UPW = upper purplish-white layer; LPW = lower (=second) purplish-white layer of Schultz and Stout (1955). PPA-1 = Peanut Peak ash #1 of D. O. Terry (unpublished data). See Emry (1973) and Evanoff et al. (1992) for discussions of Flagstaff Rim and Douglas, Wyoming, lithologies, respectively. CP = Chamberlain Pass Formation, and PP = Peanut Peak Member of Terry (this volume).

them to represent the same depositional setting for the Big Cottonwood Creek Member that is prevalent in reference sections 3, 4, and 6 through 10 (Figs. 4; 7A,B,D,E).

Reference section 13. This section occurs about 1.5 km east of Oelrichs, Fall River County, South Dakota (Fig. 4). This section was described by Terry et al. (1995; Fig. 3B), who noted distinctive and repeating sequences of green silty claystone, red silty claystone, and lacustrine limestone (Fig. 7G,H). According to Terry et al. (1995) and Evans and Welzenbach (this volume), the limestones contain ostracodes, charophytes, fish bones, and plant macrofossils.

Remarks. This section represents the northeasternmost documented extension of the Big Cottonwood Creek Member of the Chadron Formation. Lacustrine limestone beds, referred to as the "Bloom Basin limestone bed" by Welzenbach (1992) for outcrops near Wall, South Dakota, or a prominent regional unconformity, occupies this stratigraphic position in the Big Badlands to the northeast (see following discussion). According to Evans and Welzenbach (this volume), these limestones within the Big Cottonwood Creek Member probably represent localized spring-fed lakes that may have been present throughout the region.

Composite section description

Based on the preceding reference sections, a composite description of the Big Cottonwood Creek Member can be generated. The Big Cottonwood Creek Member consists of volcaniclastic silty claystones and thin interbeds of single and multistoried sandstone, pedogenic calcrete, lacustrine limestone, marl, gypsum, and volcanic ash. Prominent paleosol horizons in the upper 30 m of this unit give a characteristic banded appearance; common colors are yellow, buff, pink, tan, green, and red. The lower 15 to 20 m has more subdued colors, which include green, yellow, and dark red. Thin, discontinuously bedded gypsum, sandstone, pedogenic calcrete, lacustrine limestone, marl, volcanic ash, and calcite-cemented silty claystone beds 0.01 to 0.4 m thick are common throughout the unit, but are most frequent within the lower 15 to 20 m. These interbeds become fewer in number, but thicker and more prominent to the northeast. Pedogenic calcretes are most prevalent to the southwest, whereas lacustrine carbonates and gypsum beds are most common to the northeast. Lenticular, massive, medium- to coarse-grained sandstones (1–2 m thick) are infrequent near the base of the unit in the Toadstool Park vicinity. At least 5 prominent volcanic ashes are present. The lowest three are lenticular and restricted to paleovalleys in which the lower 20 m of the unit occurs.

This unit is locally calcareous in zones, having a nodular (5–10 cm diameter) or tabular-bedded appearance. This unit is highly fossiliferous, containing abundant vertebrate and infrequent invertebrate fossils. Preserved thicknesses outside paleovalleys average 22 m. Within paleovalleys, basins, and local depressions in underlying strata, preserved thicknesses average 20 m. When combined, the upland and paleovalley sections yield an overall thickness ranging from 42 to 52 m. The contact with the overlying Brule Formation is either intertonguing or a local erosional unconformity incised by fluvial deposits of the Brule Formation. The contact with the underlying Peanut Peak Member is also intertonguing or a local unconformity where the Big Cottonwood Creek Member fills paleovalleys. This unit weathers into smooth, near-vertical slopes and badlands.

Historical background

Darton (1899) was the first to recognize beds of the Big Cottonwood Creek Member as a discrete lithostratigraphic unit, but left them unnamed, including them within his "Brule Clays." According to his "Roundtop to Adelia" section (Darton, 1899, p. 757), this unit consisted of "greenish sands and sandy clays." These beds likely correspond to Schultz and Stout's (1955) Chadron B, Chadron C, and Orella A, and what we designate as the Big Cottonwood Creek Member. The greenish sands and sandy clays of Darton's (1899) description were overlain by sandstones that likely correspond to Schultz and Stout's (1955) Toadstool Park channel complex of the Orella Member of the Brule Formation, and were underlain by "greenish sands" that likely correspond to the Chadron A of Schultz and Stout (1955) that Terry (this volume) correlates to the Chamberlain Pass Formation of Evans and Terry (1994).

M. F. Skinner (unpublished field notes, 1951) recognized similar beds as a discrete lithologic unit in a section exposed at Trunk Butte, 9.6 km west of Chadron, Nebraska (Fig. 10C,D). Based on his observations, Skinner recognized that the Chadron Formation of northwestern Nebraska could be divided into two zones. Skinner's lower zone (Fig. 10C,D) was composed of up to 12 m of greenish-gray, smectite-rich mudstones that weather into convex-outward hills and slopes (the Peanut Peak Member of Terry, this volume). Skinner's upper zone (Fig. 10C,D), which he informally named the "Trunk Butte Member," was composed of up to 14 m of pink, greenish-pink, and brown to greenish silty clay with occasional nodular zones. This upper zone corresponds to part of our Big Cottonwood Creek Member, but whether this section represents an upland or lowland depositional setting requires additional work. Skinner placed the contact of the "Chadron and Orella" at a volcanic ash (Fig. 10C,D), labeled in his photographs as the "P.W. Ash" (for persistent white or purplish-white). Skinner classified the remaining 65 ft (19.8 m) of Trunk Butte as "Orella," stating that he could not see much evidence of an unconformity with the underlying Trunk Butte Member of the Chadron Formation, but that the true type of Orella fauna starts at this contact based on the relative increase in fossil material. R. J. Emry (written communication, 1996) stated that Skinner included all strata above the "P.W. Ash" and below the lowermost channels of the Brule Formation within his Trunk Butte Member. Skinner stated that this member could be recognized at various localities from the Nebraska-Wyoming border to near Chadron, Nebraska.

Correlations

According to Schultz and Stout (1955), the Chadron A, B, and C could be correlated to the three-fold division of the

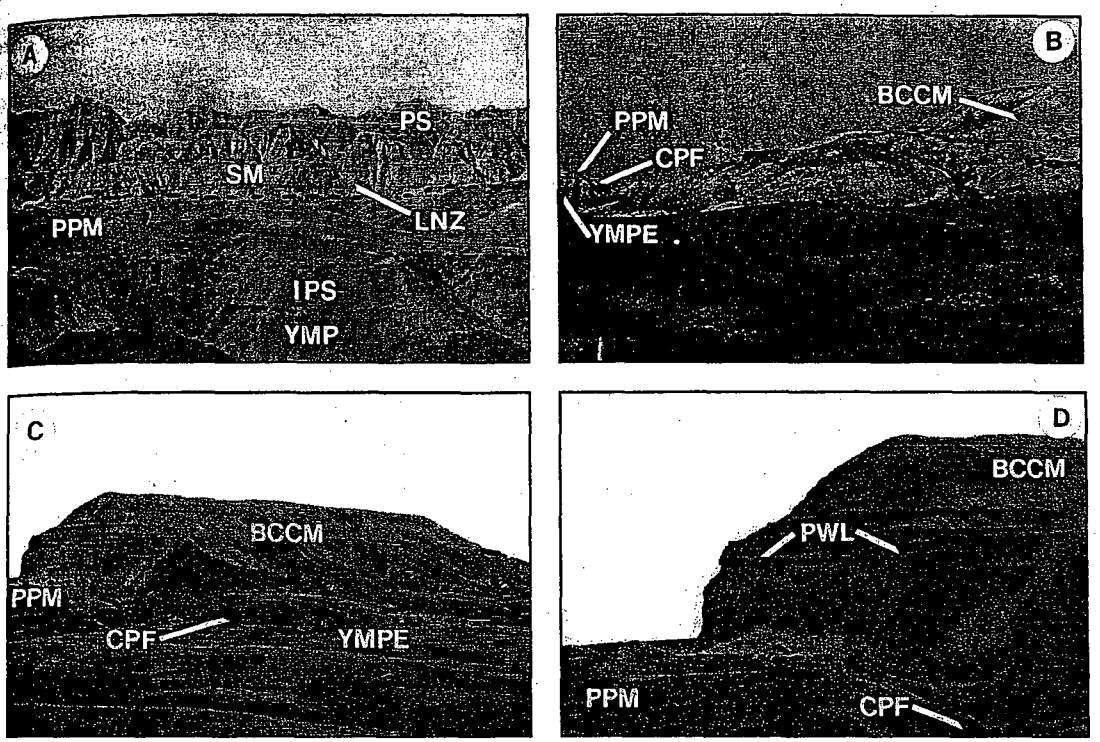


Figure 10. The Chadron Formation outside the Nebraska type area. A, Contact of the Chadron and Brule Formations at Dillon Pass, Badlands National Park, South Dakota (see Terry, this volume, Fig. 3a, location D); B, equivalent strata of the Big Cottonwood Creek Member at Lance Creek, Wyoming. The outcrop is located at the intersection of Highways 270 and 272, approximately 4.8 km (3 mi) northeast of Lance Creek (see Terry, this volume, Fig. 1, location C). C and D, Skinner's measured section at Trunk Butte, near Chadron, Nebraska. PS = Poleslide Member, and SM = Scenic Member of Bump (1956). PWL = Skinner's persistent white layer. IPS = Interior Paleosol Series, and YMP = Yellow Mounds Paleosol Series of Retallack (1983). Other abbreviations as in previous figures.

Chadron Formation in South Dakota established by Clark (1937, 1954). Schultz and Stout (1955, p. 31, 34, 38) stated that the Chadron A is "essentially the Yoder of Schlaikjer (1935), and the Ahearn or Lower Member of Clark (1954, 1937)," the Chadron B is . . . "probably only in part equivalent to the Crazy Johnson or Middle Member of Clark (1954, 1937)," and the Chadron C . . . "corresponds rather closely to the Upper Member or Peanut Peak of Clark (1954, 1937)." We question Schultz and Stout's (1955) correlation of the bluish-green, gray, and olive hummocky lower portion of the Chadron B to the Crazy Johnson Member of the Chadron Formation in South Dakota (Fig. 2) for the following reasons:

1. The lithology of the lower portion of the Chadron B in Nebraska is identical to Clark's (1937) "massive clays with intermittent limestone bands" of the undifferentiated Chadron Formation of Clark et al. (1967) in the Big Badlands (Fig. 10A), and not the "Pond limestones, bands of algal ball, massive clays, micaceous sandstones, calcareous sandstones, green, clayey sandstones, and conglomerates pink with feldspars . . . and . . . numerous channel fills" that characterize the Crazy Johnson Member in South Dakota (Clark, 1937, p. 281).

2. Based on the superposition of individual lithologic units, on the cross-cutting relationship of the Red River Valley and Chamberlain Pass Formation, and on the lateral tracing of the Chamberlain Pass Formation and the overlying bluish-green, gray, and olive hummocky claystone of the Chadron B of Schultz and Stout (1955) between Nebraska and South Dakota (see Fig. 9 of Terry, this volume), it is physically impossible to lithologically correlate or laterally connect the Chadron B with the Crazy Johnson Member within the Red River Valley of Clark (1937). We recommend the abandonment of the term Chadron B of Schultz and Stout (1955) for the bluish-green, gray, and olive hummocky mudstones that overlie the Chamberlain Pass Formation. Terry (this volume) proposes that these mudstones are lithologically equivalent to the undifferentiated mudstones of the

Chadron Formation deposited outside the Red River Valley of Clark (1937) and Clark et al. (1967) instead of the Crazy Johnson Member inside the Red River Valley, as suggested by Schultz and Stout (1955).

Schultz and Stout (1955) stated that the Chadron C corresponds rather closely to the Upper Member or Peanut Peak Member of Clark (1937, 1954). This correlation is not supported for several reasons:

1. We argue that the Chadron C is within the Big Cottonwood Creek Member and therefore cannot correlate to the Peanut Peak Member because the Chadron C overlies the Peanut Peak Member (Fig. 2). Terry (this volume) has demonstrated that the bluish-green, gray, and olive hummocky mudstones that form the base of the Chadron B in northwestern Nebraska are correlative with the "undifferentiated Chadron Formation" outside Clark's (1937, 1954) Red River Valley and with Terry's (this volume) Peanut Peak Member (Figs. 5A–D, 10A).

2. Based on guidelines within the North American Stratigraphic Code (NACSN, 1983), *lithic* correlation of the Chadron C to the Peanut Peak Member is rejected, in that the two units are of contrasting lithologies (Figs. 5A–D, 10A).

3. The designation of the Chadron C as a lithostratigraphic unit is not in agreement with the North American Stratigraphic Code (NACSN, 1983), in that the remaining strata of the underlying Chadron B and the overlying Orella A are lithologically identical to strata of the Chadron C (Figs. 4, 5A–D), and thus should also be incorporated as one lithic unit. In addition, the North American Stratigraphic Code (NACSN, 1983) suggests that event beds (purplish-white layers) not be used to define lithostratigraphic boundaries.

Dimensions and regional boundaries

We have observed outcrops identical in stratigraphic position and general lithic characteristics to the Big Cottonwood Creek Member west of the present study area. The Big Cottonwood Creek Member is not exposed south of the Pine Ridge Escarpment in the Panhandle of Nebraska.

Swinehart et al. (1985) identified three subsurface depositional sequences within the White River Group based on a detailed analysis of approximately 11,600 electric logs of oil and gas wells, and samples and logs from 500 other wells. The lowest of these three sequences encompassed the Chadron Formation to an unconformity within the Orella Member of the Brule Formation. This "Chadron Sequence" rests on an oxidized zone that is likely the result of the same period of pedogenesis that produced Schultz and Stout's (1955) Interior Paleosol Complex (=Yellow Mounds Paleosol and Interior Paleosol equivalents of Terry, this volume) in northwestern Nebraska and across the Great Plains. The Peanut Peak Member, as defined by Terry (this volume), is not discernible in the subsurface using the data of Swinehart et al. (1985), nor is our Big Cottonwood Creek Member.

Buttes near Lance Creek, Wyoming, display strata lithologically similar to the Big Cottonwood Creek Member (Fig. 10B). The strata are variegated, cliff forming, and rest on a sequence of strata lithologically similar to the Peanut Peak Member of the Chadron Formation, the Chamberlain Pass Formation, and pedogenically modified Pierre Shale (Yellow Mounds Paleosol equivalent). Luebke (1964) measured numerous sections through the White River Group in the Lance Creek area using the stratigraphic hierarchy of Schultz and Stout (1955). He reported many of the same units and prominent marker beds that Schultz and Stout (1955) described at Toadstool Park, including the Interior Paleosol Complex, the Chadron A, B, and C subdivisions and associated lower and upper purplish-white layers, the Orella B, C, and D subdivisions and associated lower and upper nodules, and the Whitney A, B, and C subdivisions.

The Big Cottonwood Creek Member may be present in the Douglas, Wyoming, area, but how the Big Cottonwood Creek Member correlates to the lithostratigraphy of Evanoff (1990) is uncertain. Evanoff (1990) classified strata in the Douglas area as the White River Formation, consisting of the Chadron and Brule Members. The difference in ranking was the result of difficulty in defining the boundary between the two members and because of their less distinct appearance in outcrop as compared to northwestern Nebraska (Evanoff et al., 1992). The names Chadron and Brule were retained due to the lithologic similarity to their respective counterparts in the Toadstool Park area of northwestern Nebraska.

Leonard (1957) measured numerous sections through the White River Group in the Douglas, Wyoming, area using the stratigraphic hierarchy of Schultz and Stout (1955). According to Leonard (1957), many of the stratigraphic units and marker beds described by Schultz and Stout (1955) at Toadstool Park are also present in the Douglas area. Within the Chadron Formation these include the Chadron B and C subdivisions and associated lower and upper purplish-white layers. Within the Brule Formation these include the Orella A, B, C, and D subdivisions and associated lower nodules and white bed, and the Whitney A, B, and C subdivisions and associated lower and upper Whitney ashes.

The Big Cottonwood Creek Member is not present in the Big Badlands/Badlands National Park area of South Dakota (Figs. 4, 10A, 11; Terry, Fig. 9, this volume). In the Big Badlands, the contact between the Chadron and Brule Formations is marked by a lithologic change from hummocky, bluish-green, gray, and olive claystones of Clark et al.'s (1967) "undifferentiated Chadron Formation" (=Peanut Peak Member of Terry, this volume) to the tan, beige, and buff clayey siltstones and sandstones of Bump's (1956) Scenic Member of the Brule Formation (Figs. 10A, 11). Lacustrine limestones occasionally occur within the Chadron Formation at or near this lithologic contact. According to Evans and Welzenbach (this volume), these limestones contain charophytes, oogonia, ostracodes, gastropods, pelecypods, fish fossils, and a very low amount of siliciclastics. Probable lithologic equivalents of these limestones are interbedded with clayey siltstones of the Big Cottonwood Creek Member at Limestone Butte, near Oelrichs, South Dakota (Figs. 4, 7G,H).

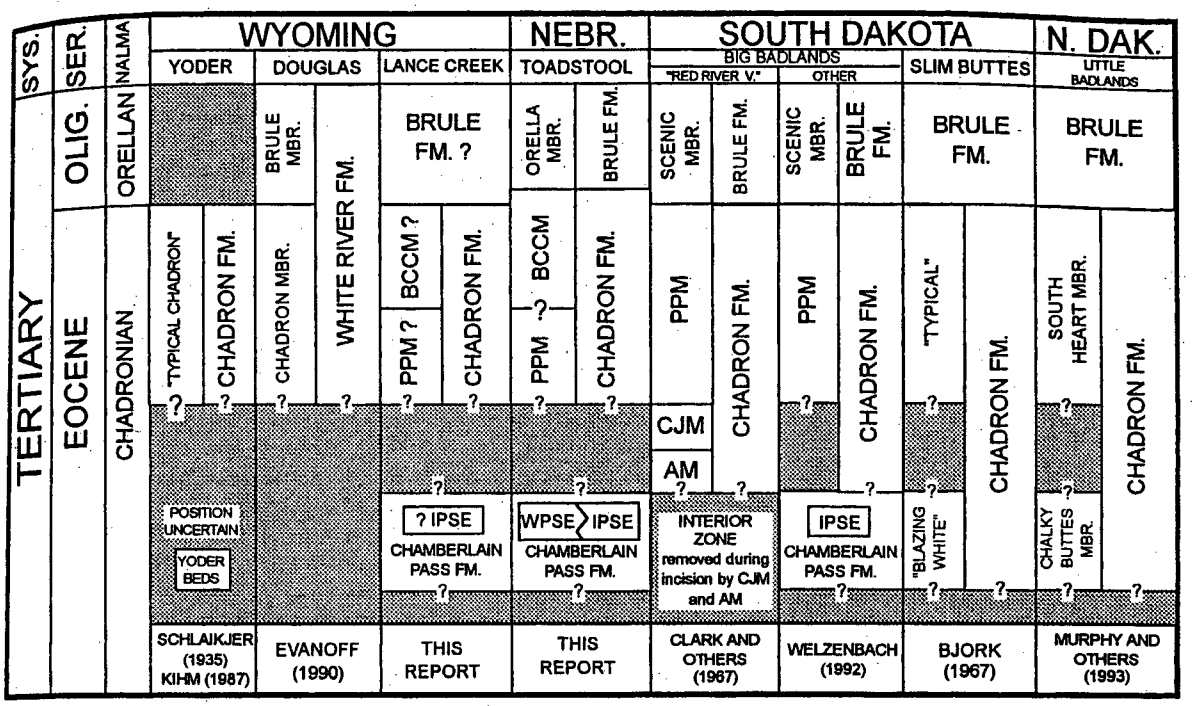


Figure 11. Lithologic equivalency chart for the White River Group at various locations across the Great Plains. See also Terry (this volume, Fig. 1). IPSE and WPSE = the interior and Weta paleosol series equivalents of Terry (this volume). Other abbreviations as in previous figures.

Age, dating, and calibration

Vertebrate fossils within both the underlying Peanut Peak Member and the Big Cottonwood Creek Member indicate a Chadronian North American land-mammal "age," or NALMA (Prothero and Whittlesey, this volume). Prothero and Swisher (1992) have placed the Chadronian NALMA in the late Eocene. The boundary between the Chadronian and Orellan land-mammal "ages" in the Toadstool Park area is unknown, but possibly corresponds to our lithologic boundary between the Big Cottonwood Creek Member (Chadron Formation) and Orella Member (Brule Formation). Based on recent $^{40}Ar/^{39}Ar$ single-crystal laser fusion dates of tuffs in the Douglas and Flagstaff Rim areas of Wyoming, the upper purplish-white layer at Toadstool Park, now thought to be correlative with the J tuff at Flagstaff Rim and 3c tuff at Douglas, has an age of 34.36 ± 0.11 Ma (Obradovich et al., 1995; Larson and Evanoff, this volume).

The majority of magnetostratigraphic studies within the Toadstool Park area (Prothero, 1985; Prothero and Swisher, 1992) sampled the upper part of the Chadron Formation and younger strata. In the Toadstool Park area the Chadron Formation occurs within Chron C13, with the transition from the Chadron to Brule Formations (*sensu* Schultz and Stout, 1955) occurring within Chron C13R (Prothero and Swisher, 1992; Prothero, 1996). Using our lithostratigraphic classification, the magnetostratigraphic transition from Chron 13r to 13n would occur within the upper 10 m of the Big Cottonwood Creek Member above the upper purplish-white layer. The base of Prothero and Swisher's (1992) Toadstool Park paleomagnetic section stops within what may be Chron C13R at the fourth purplish-white layer (Prothero and Swisher, 1992; Prothero, 1996). In the Toadstool Park area, the fourth purplish-white layer, the Chadron B^2/B^3 boundary of Schultz and Stout (1955, Fig. 3, section 2), occurs at the top of the Peanut Peak Member (Fig. 4).

Data from Ostrander (1985) suggest that a medial Chadronian local fauna, the Raben Ranch local fauna, is present within sediments that we classify as the Big Cottonwood Creek Member. This local fauna occurs ~4.8 km (3 mi) north-northwest of Toadstool Park within the lower half of the Big Cottonwood Creek Member. Prothero's (1996) magnetostratigraphic section of the Raben Ranch, which follows Ostrander's (1985) measured section, is interpreted as Chron C16n.1 and C15r. Ostrander (1985) also described several other medial Chadronian local faunas 9.7 km (6 mi) to the southeast of Toadstool Park within what we classify as the lower half of the Big Cottonwood Creek Member. These data, along with the small geographic distance and lithologic similarity between these local faunas and Toadstool Park, suggest that the short normal magnetozone at the base of the

Toadstool Park section of Prothero and Swisher (1992), and Prothero (1996) may be Chron 15n.

Based on volcanic ash correlation data (Evanoff et al., 1992; Larson and Evanoff, this volume; Evanoff and Larson, unpublished data), ashes within the upper part of the Big Cottonwood Creek Member in reference sections 6 and 9 would likely occur within the lowest part of Chron C13R (Fig. 4, 9), whereas strata within the lowest part of these sections would likely occur within Chron C15 or C16. Based on our lithostratigraphic revision of the Chadron Formation in northwestern Nebraska, we recognize a gap in the existing paleomagnetic coverage of the Chadron Formation. According to Prothero and Swisher's (1992) measured section at Toadstool Park, the Peanut Peak Member and the lowermost 20 m of the Big Cottonwood Creek Member occurring within paleovalleys (reference sections 3, 4, and possibly 6 through 9) would not have been sampled for paleomagnetics.

Genesis

Initial deposition of the Big Cottonwood Creek Member occurred within paleovalleys incised into the underlying Peanut Peak Member. In some areas, these paleovalleys within the Peanut Peak Member remained a stable landscape for a period of time sufficient to allow pedogenic modification along the paleovalley walls. Within the paleovalleys, fluvial and lacustrine deposition of the Big Cottonwood Creek Member occurred either rapidly, producing only weak paleosols, or slowly and cyclically, producing paleosols that were then inundated by shallow, ephemeral, possibly spring-fed lakes. Within the paleovalleys, fluvial deposits were restricted to overbank claystones and silty claystones. With the filling of the paleovalleys, fluvial and lacustrine deposition proceeded on the former upland areas.

On the uplands, fluvial and lacustrine deposition continued, but at different rates in different locations. Fluvial deposition dominated to the south and west, and produced isolated, trough cross-bedded, multistory channel sandstones and overbank silty claystones representing mixed-load streams. In these areas deposition slowed to a rate that allowed more extensive pedogenic modification of the fluvial deposits (see the following discussion). To the north and east, cyclic sedimentation and lacustrine deposition dominated. These cycles consisted of a period of fluvial deposition, followed by pedogenic modification and inundation by lakes.

As in the Big Badlands of South Dakota (Retallack, 1983), ancient pedogenic modification within the White River Group in the Toadstool Park area has produced characteristic variegated color banding (Figs. 5A–D, 6A). Schultz and Stout (1955, 1980) and Schultz et al. (1955) discussed the presence of paleosols within the White River Group, and suggested that they marked regional unconformities of allostratigraphic and biostratigraphic significance. In contrast to Schultz and Stout (1980), who recognized only six "major" and several minor paleosols within the White River Group in northwestern Nebraska, we have recognized numerous paleosols throughout the Big Cottonwood Creek Member.

Twenty-two superimposed paleosol profiles are preserved within the 29 m of upland deposits exposed in the type area of the Big Cottonwood Creek Member (Terry, 1995). Paleosol profiles vary in their degree of development, ranging from weakly developed A–C profiles to well-developed A–B–C profiles with extensive horizons of pedogenic calcrete. The variability in the degree of profile development may be due to either the proximity of soil profiles to former rivers, position on the ancient landscape, or the period of time between depositional events. Identifiable macroscopic soil features include drab-haloed, clay-filled, and crystalline root traces, peds, cutans, slickensided cutans, and carbonate nodules. Based on field comparisons, paleosols within the Big Cottonwood Creek Member developed under less humid conditions than paleosols within the underlying Peanut Peak Member of the Chadron Formation. This change in soil-forming conditions is apparent in the modern erosional relief, lithic character, and paleosols of these two members. The underlying Peanut Peak Member is more smectitic, and weathers into low haystack mounds and hummocks. The overlying Big Cottonwood Creek Member weathers into steep slopes and cliffs. This change is likely the result of a reduced rate of chemical weathering of volcaniclastic sediments within the Big Cottonwood Creek Member and an increased silt content. Paleosols within the Peanut Peak Member only occasionally contain pedogenic calcrete, whereas paleosols within the Big Cottonwood Creek Member commonly contain pedogenic calcrete, likely reflecting progressively less humid conditions during deposition of the Big Cottonwood Creek Member.

In the Pinnacles and Dillon Pass areas of Badlands National Park, the stratigraphic position of the Big Cottonwood Creek Member of the Chadron Formation is marked either by a disconformity of approximately 400,000 yr (Prothero and Whittlesey, this volume) between the Peanut Peak Member of the Chadron Formation (sensu Terry, this volume) and the Scenic Member of the Brule Formation, or occasionally by lacustrine limestone beds (Terry, this volume, Fig. 9; Fig. 10a). Welzenbach (1992) referred to outcrops of these beds near Wall, South Dakota, as the "Bloom Basin limestone bed." According to Evans and Welzenbach (this volume), this period of increased carbonate accumulation within the upper portion of the Chadron Formation may be the result of the interaction of nearly equal rates of sediment accumulation and basin subsidence, establishment of a Late Eocene regional groundwater system related to the uplift and unroofing of the Black Hills, and paleoclimate.

It is our interpretation that the unconformity recognized by Prothero and Whittlesey (this volume) between the Chadron and Brule Formations, and the formation of numerous lacustrine deposits within the upper portion of the Chadron Formation, reflects a period of little or no fluvial deposition in the Big Badlands of South Dakota during the late Eocene. The presence of a similar lacustrine limestone and gypsum beds at and near the top of the Peanut Peak Member across northwestern Nebraska (Evans and Welzenbach, this volume; Terry, this volume; LaGarry and LaGarry, 1997) suggests that the region may have experienced a

period of reduced fluvial sedimentation during the late Eocene. For unknown reasons, possibly the continued or renewed uplift of the Black Hills or Chadron Dome, subsequent fluvial deposition and reworking of sediments (Big Cottonwood Creek Member) was concentrated in the northwestern Nebraska area.

CONCLUSIONS

Based on detailed lithostratigraphic correlation of rock units from northwestern Nebraska to the Big Badlands of South Dakota, we propose that the upper two-thirds of Schultz and Stout's (1955) Chadron B and the entire Chadron C subdivisions of the Chadron Formation, along with their Orella A subdivision of the Brule Formation, be combined to form a new lithostratigraphic unit that we herein designate as the Big Cottonwood Creek Member of the Chadron Formation. This revision is justified on the basis of the following: (1) the stratigraphic hierarchy erected by Schultz and Stout (1955) is based on the position of marker beds presumed to correspond with cycles of deposition, not lithologic character, as specified in the North American Stratigraphic Code (NACSN, 1983); and (2) the lithologies of the upper two-thirds of the Chadron B, Chadron C, and Orella A are demonstrably the same. The Big Cottonwood Creek Member has no lithologic counterpart in the Big Badlands of South Dakota. As of this report, the northernmost documented outcrop of the Big Cottonwood Creek Member is at Limestone Butte, near Oelrichs, South Dakota.

Our introduction of the Big Cottonwood Creek Member of the Chadron Formation, including a previously unrecognized 20- to 22-m paleovalley sequence, contributes to our understanding of the geologic history of the northern Great Plains. The newly recognized volcanic ashes within the Big Cottonwood Creek Member may serve as a means of correlating the White River Group across the Great Plains, as suggested by the preliminary correlations between Flagstaff Rim and Douglas, Wyoming, and the Toadstool Park area. As the Big Cottonwood Creek Member is traced westward, the lithologic relationships of exposures in Nebraska and Wyoming should become clearer. In addition, the stratigraphic interval represented by the Big Cottonwood Creek Member possibly records at least 400,000 yr (Prothero and Whittlesey, this volume) that was not preserved in the Big Badlands of South Dakota. The Big Cottonwood Creek Member provides an opportunity to closely examine climatic changes and vertebrate evolution during this period in proximity to the Eocene/Oligocene boundary, and to refine our concepts of late Eocene magneto- and biostratigraphy.

Further study may help to constrain the various tectonic pulses of the Black Hills and Chadron Dome, and to determine their effects on depositional patterns of the White River Group during the late Eocene. The Eocene/Oligocene transition is regarded as recording a gradual climatic shift from warm and humid to cooler and drier (Retallack, 1983; Evanoff et al., 1992; Prothero, 1994). Based on previous stratigraphic hierarchies and assumed correlations for the White River Group, this gradual change was thought to have begun in the west and progressed eastward (Evanoff et al., 1992). Our new lithostratigraphic model suggests that a reanalysis of this climatic model may be in order.

ACKNOWLEDGMENTS

We thank Bob Hunt, David Loope, Jim Swinehart, and Jim Evans for discussions of the finer points of the North American Stratigraphic Code; Ed Larson, Emmett Evanoff, Jim Evans, and Leigh Anne LaGarry for contributing ideas and insights on Nebraska stratigraphy and for sharing unpublished data; and the staffs of Badlands National Park and the Nebraska National Forest for their contributions of time and resources. We thank Bob Emry, Jim Honey, and Bob Hunt for reviewing this chapter in its manuscript stage. Terry's research was supported by the Nebraska National Forest, the University of Nebraska—Lincoln Geology Department Shell Fellowship, the Nebraska Geological Society Yatkola-Edwards Research Grant, the University of Nebraska—Lincoln Center for Great Plains Studies, and a Sigma Xi Grant-in-Aid. LaGarry's research was supported by Challenge Cost-Share Agreements 02-07-91-013, 02-07-92-025, 02-07-93-011, and 02-07-94-015 between the Nebraska National Forest and the University of Nebraska—Lincoln, and by the University of Nebraska Conservation and Survey Division state map project.

REFERENCES CITED

Bjork, P. R., 1967, Latest Eocene vertebrates from northwestern South Dakota: Journal of Paleontology, v. 41, p. 227–236.

Bump, J. D., 1956, Geographic names for members of the Brule Formation of the Big Badlands of South Dakota: American Journal of Science, v. 254, p. 429–432.

Clark, J., 1937, The stratigraphy and paleontology of the Chadron Formation in the Big Badlands of South Dakota: Carnegie Museum Annals, v. 25, p. 261–350.

Clark, J., 1954, Geographic designation of the members of the Chadron Formation in South Dakota: Carnegie Museum Annals, v. 33, p. 197–198.

Clark, J., 1975, Controls of sedimentation and provenance of sediments in the Oligocene of the central Rocky Mountains, *in* Curtis, B. F., ed., Cenozoic history of the southern Rocky Mountains: Geological Society of America Memoir 144, p. 95–117.

Clark, J., Beerbower, J. R., and Kietzke, K. K., 1967, Oligocene sedimentation, stratigraphy, paleoecology, and paleoclimatology in the Big Badlands of South Dakota: Fieldiana Geology Memoirs, v. 5, 158 p.

Darton, N. H., 1899, Preliminary report on the geology and water resources of Nebraska west of the one hundred and third meridian: U.S. Geological Survey 19th Annual Report, Pt. 4, p. 719–814.

Emry, R. J., 1973, Stratigraphy and preliminary biostratigraphy of the Flagstaff Rim area, Natrona County, Wyoming: Smithsonian Contributions to Paleobiology, v. 18, p. 1–43.

Emry, R. J., Bjork, P. R., and Russell, L. S., 1987, The Chadronian, Orellan, and Whitneyan North American land mammal ages, *in* Woodburne, M. O., ed., Cenozoic mammals of North America: geochronology and biostratigraphy: Berkeley, University of California Press, p. 118–152.

Evanoff, E., 1990, Late Eocene and early Oligocene paleoclimates as indicated by the sedimentology and nonmarine gastropods of the White River Formation near Douglas, Wyoming [Ph.D. thesis]: Boulder, University of Colorado, 440 p.

Evanoff, E., Prothero, D. R., and Lander, R. H., 1992, Eocene-Oligocene climatic

change in North America: the White River Formation near Douglas, east-central Wyoming, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 116–130.

Evans, J. E., and Terry, D. O., Jr., 1994, The significance of incision and fluvial sedimentation in the Basal White River Group (Eocene-Oligocene), Badlands of South Dakota, U.S.A.: Sedimentary Geology, v. 90, p. 137–152.

Harksen, J. C., and Macdonald, J. R., 1969, Type sections for the Chadron and Brule Formations of the White River Oligocene in the Big Badlands of South Dakota: South Dakota Geological Survey Report of Investigations 99, 23 p.

Harvey, C., 1960, Stratigraphy, sedimentation, and environment of the White River Group of the Oligocene of northern Sioux County, Nebraska [Ph.D. thesis]: Lincoln, University of Nebraska, 151 p.

Hatcher, J. B., 1893, The Titanotherium beds: American Naturalist, v. 27, p. 204–221.

Kihm, A. J., 1987, Mammalian paleontology and geology of the Yoder Member, Chadron Formation, east-central Wyoming, *in* Martin, J. E., and Ostrander, G. E., eds., Papers in vertebrate paleontology in honor of Morton Green: Dakoterra, v. 3, p. 28–45.

LaGarry, H. E., 1995, Provisional revision of lithostratigraphic boundaries of the Brule Formation in northwestern Nebraska: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A66–67.

LaGarry-Guyon, H. E., and Hunt, R. M., Jr., 1992, 1991 Paleontological resource survey of the Oglala National Grassland (Roundtop 7.5× Quadrangle), Sioux County, Nebraska: U.S. Department of Agriculture Forest Service/University of Nebraska—Lincoln Challenge Cost-Share Agreement No. 02-07-91-013, 26 p.

LaGarry-Guyon, H. E., and Hunt, R. M., Jr., 1993, 1992 Paleontological resource survey of the Oglala National Grassland (Roundtop, Horn, and Orella 7.5′ Quadrangles), Sioux and Dawes Counties, Nebraska: U.S. Department of Agriculture Forest Service/University of Nebraska—Lincoln Challenge Cost-Share Agreement No. 02-07-92-025, 65 p.

LaGarry, H. E., and LaGarry, L. A., 1997, Geology of the Montrose, Orella, Wolf Butte, Roundtop, and Horn 7.5× US Geological Society Quadrangles, Sioux and Dawes Counties, Nebraska: University of Nebraska Conservation and Survey Division Open-File Report No. 48, 159 p.

Larson, E. E., and Evanoff, E., 1995, Tephrostratigraphy of the White River sequence in Nebraska, Wyoming, and Colorado: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A67.

Leonard, B. F., 1957, Oligocene stratigraphy of the Douglas area, Converse County, Wyoming [M.S. Thesis]: Lincoln, University of Nebraska, 48 p.

Lisenbee, A. L., and DeWitt, E., 1993, Laramide evolution of the Black Hills uplift, *in* Snoke, A. W., Steidtmann, J. R., and Roberts, S. M., eds., Geology of Wyoming: Geological Survey of Wyoming Memoir 5, p. 374–412.

Luebke, L. O., 1964, Oligocene stratigraphy of the Lance Creek area, Wyoming [M.S. thesis]: Lincoln, University of Nebraska, 78 p.

Martin, J. E., 1985, Introduction to the geology and paleontology of western South Dakota and northwestern Nebraska, *in* Martin, J. E., ed., Fossiliferous Cenozoic deposits of western South Dakota and northwestern Nebraska: Dakoterra, v. 2, pt. 2, p. 1–11.

Meek, F. B., and Hayden, F. V., 1857, Descriptions of new species and genera of fossils, collected by Dr. F. V. Hayden in Nebraska Territory; with some remarks on the Tertiary and Cretaceous Formations of the northwest, and the parallelism of the latter with those of other portions of the United States and Territories: Proceedings, Academy of Natural Science of Philadelphia for 1857, p. 117–148.

Meek, F. B., and Hayden, F. V., 1861, Descriptions of new Lower Silurian (primordial), Jurassic, Cretaceous, and Tertiary fossils, collected in Nebraska, by the exploring expedition under the command of Capt. Wm. F. Raynolds, U.S. topographical engineer; with some remarks on the rocks from which they were obtained: Proceedings, Academy of Natural Science of Philadelphia, v. 13, p. 415–447.

Murphy, E. C., Hoganson, J. W., and Forsman, N. F., 1993, The Chadron, Brule, and Arikaree Formations in North Dakota: North Dakota Geological Survey Report of Investigations 96, 144 p.

NACSN (North American Commission on Stratigraphic Nomenclature), 1983, North American Stratigraphic Code: American Association of Petroleum Geologists Bulletin, v. 67, p. 841–875.

Obradovich, J. D., Evanoff, E., and Larson, E. E., 1995, Revised single-crystal laser-fusion $^{40}Ar/^{39}Ar$ ages of Chadronian tuffs in the White River Formation of Wyoming: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A77.

Osborn, H. F., 1929, Titanotheres of ancient Wyoming, Dakota, and Nebraska: U.S. Geological Survey Monograph 55, 953 p.

Ostrander, G. E., 1985, Correlation of the early Oligocene (Chadronian) in northwestern Nebraska, *in* Martin, J. E., ed., Fossiliferous Cenozoic deposits of western South Dakota and northwestern Nebraska: Dakoterra, v. 2, p. 205–231.

Pettyjohn, W. A., 1966, Eocene paleosol in the northern Great Plains: U.S. Geological Survey Professional Paper 550C, p. 61–65.

Prothero, D. R., 1985, Chadronian (early Oligocene) magnetostratigraphy of eastern Wyoming: implications for the age of the Eocene-Oligocene boundary: Journal of Geology, v. 93, p. 555–565.

Prothero, D. R., 1994, The Eocene-Oligocene transition: paradise lost: New York, Columbia University Press, 291 p.

Prothero, D. R., 1996, Magnetostratigraphy of the White River Group in the High Plains, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, U.K., Cambridge University Press, p. 262–277.

Prothero, D. R., and Swisher, C. C., III, 1992, Magnetostratigraphy and geochronology of the terrestrial Eocene-Oligocene transition in North America, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 46–73.

Retallack, G. J., 1983, Late Eocene and Oligocene paleosols from Badlands National Park, South Dakota: Geological Society of America Special Paper 193, 82 p.

Schlaikjer, E. M., 1935, Contributions to the stratigraphy and paleontology of the Goshen Hole area, Wyoming, Part III, A new basal Oligocene formation: Bulletin of the Museum of Comparative Zoology, v. 76, no. 3, p. 71–93.

Schultz, C. B., and Falkenbach, C. H., 1968, The phylogeny of the oreodonts. Parts 1–2: Bulletin of the American Museum of Natural History, v. 139, 498 p.

Schultz, C. B., and Stout, T. M., 1938, Preliminary remarks on the Oligocene of Nebraska: Geological Society of America Bulletin, v. 49, p. 1921.

Schultz, C. B., and Stout, T. M., 1945, The Pleistocene loess deposits of Nebraska: American Journal of Science, v. 243, no. 5, p. 231–244.

Schultz, C. B., and Stout, T. M., 1948, Pleistocene mammals and terraces of the Great Plains: Geological Society of America Bulletin, v. 59, no. 6, p. 553–588.

Schultz, C. B., and Stout, T. M., 1955, Classification of Oligocene sediments of Nebraska: Bulletin of the University of Nebraska State Museum, v. 4, p. 17–52.

Schultz, C. B., and Stout, T. M., 1977, Drought and the model of a Quaternary terrace-cycle: Transactions of the Nebraska Academy of Science, v. 4, p. 191–201.

Schultz, C. B., and Stout, T. M., 1980, Ancient soils and climatic changes in the central Great Plains: Transactions of the Nebraska Academy of Science, v. 8, p. 187–205.

Schultz, C. B., Tanner, L. G., and Harvey, C., 1955, Paleosols of the Oligocene of Nebraska: Bulletin of the University of Nebraska State Museum, v. 4, p. 1–16.

Singler, C. R., and Picard, M. D., 1979, Petrography of the White River Group (Oligocene) in northwest Nebraska and adjacent Wyoming: Contributions to Geology, University of Wyoming, v. 18, no. 1, p. 51–67.

Singler, C. R., and Picard, M. D., 1980, Stratigraphic review of Oligocene beds in northern Great Plains: Wyoming Geological Association Earth Science

Bulletin, v. 13, no. 1, p. 1–18.

Singler, C. R., and Picard, M. D., 1981, Paleosols in the Oligocene of northwest Nebraska: Contributions to Geology, v. 20, no. 1, p. 57–68.

Stanley, K. O., and Benson, L. V., 1979, Early diagenesis of High Plains Tertiary vitric and arkosic sandstone, Wyoming and Nebraska, *in* Scholle, P. A., and Schluger, P. R., eds., Aspects of Diagenesis: Tulsa, Society of Economic Geologists and Paleontologists Special Publication 26, p. 401–423.

Stone, W. J., 1972, Middle Cenozoic stratigraphy of North Dakota, *in* Ting, F. T. C., ed., Depositional environments of the lignite-bearing strata in western North Dakota: North Dakota Geological Survey Miscellaneous Series 50, p. 123–132.

Stout, T. M., 1971, Quaternary classification and sedimentary model [abs.]: Union Internationale pour l'étude du Quaternaire (INQUA), VIIIe Congrès, Paris, 1969, études sur le Quaternaire dans le Monde, v. 2, p. 790.

Stout, T. M., 1973, Revised classification of the Quaternary and Medial to Late Tertiary, based upon the standard for the Great Plains (U.S.A.) and intercontinental correlations [abs.]: International Union for Quaternary Research (INQUA) Ninth Congress, Christchurch, New Zealand, p. 347–348.

Stout, T. M., 1977, Frontiers of Nebraska geology: Transactions of the Nebraska Academy of Science, v. 4, p. 117–120.

Stout, T. M., 1978, The comparative method in stratigraphy: the beginning and end of an ice age: Transactions of the Nebraska Academy of Science, v. 6, p. 1–18.

Swenson, D. B., 1959, The heavy mineral assemblages of the Chadron and Brule Formations in northwestern Nebraska [M.S. thesis]: Lincoln, University of Nebraska—Lincoln, 66 p.

Swinehart, J. B., Souders, V. L., Degraw, H. M., and Diffendal, R. F., Jr., 1985, Cenozoic paleogeography of western Nebraska, *in* Flores, R. M., and Kaplan, S., eds., Cenozoic paleogeography of the west central United States: Special Publication, Rocky Mountain Section, Society of Economic Paleontologists and Mineralogists, p. 209–229.

Swisher, C. C., III, and Prothero, D. R., 1990, Single-crystal $^{40}Ar/^{39}Ar$ dating of the Eocene-Oligocene transition in North America: Science, v. 249, p. 760–762.

Tedford, R. H., Swinehart, J. B., Swisher, C. C., III, Prothero, D. R., King, S. A., and Tierney, T. E., 1996, The Whitneyan-Arikareean Transition in the High Plains, *in* Prothero, D. R., and Emry, R. S., eds., The Terrestrial Eocene-Oligocene Transition in North America: Cambridge, Cambridge University Press, p. 312–334.

Terry, D. O., Jr., 1995, Paleopedology of the Big Cottonwood Creek Member of the Eocene Chadron Formation, White River Group, badlands of northwestern Nebraska [abs.]: Proceedings, 105th Annual Meeting, Nebraska Academy of Science, p. 57.

Terry, D. O., Jr., and Evans, J. E., 1994, Pedogenesis and paleoclimatic implications of the Chamberlain Pass Formation, Basal White River Group, Badlands of South Dakota: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 110, p. 197–215.

Terry, D. O., Jr., and LaGarry, H. E., 1994, Lithostratigraphic correlation and provisional redefinition of the Eocene-Oligocene White River Group, Nebraska and South Dakota: Geological Society of America Abstracts with Programs, v. 26, no. 7, p. A249.

Terry, D. O., Jr., and LaGarry, H. E., 1995, Provisional redescription and redefinition of the upper part of the Chadron Formation in northwestern Nebraska: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A89.

Terry, D. O., Jr., LaGarry, H. E., and Wells, W. B., 1995, The White River Group revisited: vertebrate trackways, ecosystems, and lithostratigraphic revision, redefinition, and redescription, *in* Flowerday, C. A., ed., Geologic field trips in Nebraska and adjacent parts of Kansas and South Dakota: University of Nebraska Conservation and Survey Division Guidebook 10, p. 43–53.

Vondra, C. F., 1958, The stratigraphy of the Chadron Formation in northwestern Nebraska [M.S. thesis]: Lincoln, University of Nebraska, 138 p.

Vondra, C. F., 1960, Stratigraphy of the Chadron Formation in northwestern Nebraska: Compass, v. 37, no. 2, p. 73–90.

Wanless, H. R., 1922, Lithology of the White River sediments: American Philosophical Society Proceedings, v. 62, p. 663–669.

Wanless, H. R., 1923, The stratigraphy of the White River beds of South Dakota: American Philosophical Society Proceedings, v. 62, p. 191–267.

Ward, F., 1922, The geology of a portion of the Badlands: South Dakota Geological and Natural History Survey Bulletin, v. 11, 80 p.

Wells, W. B., Terry, D. O., Jr., and Nixon, D. A., 1994, Multiple cut and fill paleovalley sequences within Oligocene sediments (White River Group) of northwestern Nebraska: Geological Society of America Abstracts with Programs, v. 26, no. 7, p. A492.

Wells, W. B., Terry, D. O., and LaGarry, H. E., 1995, Stratigraphic implications of a fluvial origin for the "nodular zones," Brule Formation (Orella Member), northwestern Nebraska: Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A95.

Welzenbach, L. C., 1992, Limestones in the Lower White River Group (Eocene-Oligocene), Badlands of South Dakota: depositional environment and paleoclimatic implications [M.S. thesis]: Bowling Green, Ohio, Bowling Green State University, 131 p.

MANUSCRIPT ACCEPTED BY THE SOCIETY DECEMBER 5, 1997

Printed in U.S.A.

Geological Society of America
Special Paper 325
1998

Magnetic polarity stratigraphy and correlation of the Arikaree Group, Arikareean (late Oligocene–early Miocene) of northwestern Nebraska

Bruce J. MacFadden
Florida Museum of Natural History, University of Florida, Gainesville, Florida 32611
Robert M. Hunt, Jr.
Division of Vertebrate Paleontology, University of Nebraska, Lincoln, Nebraska 68588

ABSTRACT

The volcaniclastic sediments of the Arikaree Group of northwestern Nebraska are of fundamental importance to the calibration of late Oligocene–early Miocene time throughout North America using fossil land-mammals. The defining fauna of the Arikareean land-mammal age is derived from this rock sequence. This chapter presents the results of a comprehensive magnetostratigraphic study of the Arikaree Group in Sioux County, Nebraska, including the Gering, Monroe Creek, and Harrison Formations and Upper Harrison beds (and includes the lower part of the Runningwater Formation, the oldest unit of the overlying Hemingford Group). Paleomagnetic samples were collected and analyzed from 266 horizons within nine principal measured sections within the Arikaree Group.

The lower part of the Arikaree Group, including the Gering and lower Monroe Creek Formations, consists of an R-N (reversed-normal) sequence that correlates to chron C9 between ~28.2 and 27 Ma. The middle and upper part of the Monroe Creek and overlying ?Harrison Formation contain only three polarity zones (R-N-R); these cannot be unambigously correlated to the Magnetic Polarity Time Scale (MPTS) other than that they are between 27 and 22 Ma. Above this hiatus the lower part of the Harrison Formation consists of a short (N-R-N-R) sequence that correlates to the late part of Chron C6A and early part of Chron C6r, or alternatively, to Chron C6AAn and part of Chron C6A: this latter correlation is consistent with a radioisotopic age determination of 21.9 Ma for the base of this sequence. The middle and upper part of the Harrison Formation and lowermost part of the Upper Harrison beds consist of a single, long normal polarity zone that, when used in conjunction with an interbedded ash of 19.2 ± 0.5 Ma near the top of this zone, correlates unambiguously to the prominent normal zone C6n between 20.1 and 19 Ma on the MPTS. The world-famous, exceedingly rich Agate waterhole bonebed occurs at the top of Chron C6, and in conjunction with the upper dated ash, indicates an age of about 19.2 Ma. The overlying Upper Harrison beds are predominantly of reversed polarity and correlate to Chron C5Er. Above the Arikaree Group, the lowermost Runningwater Formation at Agate National Monument correlates to the latest part of Chrons C5Er, C5En, and the early part of C5Dr. Using these correlations, the local duration for the lower part of the Runningwater Formation is between ~18.8 and 18.0 m.y.

As typified in Sioux County, Arikareean faunas occur within Chrons C9–C6,

MacFadden, B. J., and Hunt, R. M., Jr., 1998, Magnetic polarity stratigraphy and correlation of the Arikaree Group, Arikareean (late Oligocene–early Miocene) of northwestern Nebraska, *in* Terry, D. O., Jr., LaGarry, H. E., and Hunt, R. M., Jr., eds., Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America): Boulder, Colorado, Geological Society of America Special Paper 325.

extend into Chron C5Er, and span a duration between 28.2 and 18.8 m.y. Earliest Hemingfordian faunas occur in the upper part of Chrons C5Er and C5En and are younger than 18.8 Ma. The extrapolated age for the Arikareean/Hemingfordian boundary is therefore 18.8 Ma.

INTRODUCTION

Upper Oligocene–lower Miocene volcaniclastic sediments of the Arikaree Group are widely exposed over ca. 3,000 km^2 in a region encompassing northeastern Colorado, southeastern Wyoming, southwestern South Dakota, and western Nebraska. Fossil mammals have been reported from these sediments for more than a century, and these taxa represent some very important transitional faunas between mammals of archaic, and those of more modern, aspect. The biochronology of these extinct mammals from this region, in particular those of Sioux County in northwestern Nebraska, is the fundamental basis for our understanding of the Arikareean North American Land-Mammal Age (NALMA) (Wood et al., 1941; Tedford et al., 1987). Important localities assigned to the age are found in Florida, the midcontinent, Montana, Wyoming, New Mexico, California, and the John Day Formation of the Columbia River Plateau, Oregon.

Although the other recognized NALMA type localities previously have been characterized by their magnetic polarity stratigraphy (Opdyke, 1990; Opdyke and Channell, 1996), and this framework has been of immense utility in more refined chronological calibrations, the sediments from the Arikaree Group in Sioux County have heretofore remained unstudied for either their paleomagnetic signature or correlation to the Magnetic Polarity Time Scale (MPTS). Accordingly, this chapter presents the results of a magnetostratigraphic study of the typical Arikaree Group as it is exposed in Sioux County, northwestern Nebraska. The magnetostratigraphic results from this study are important because they provide: (1) a framework for local chronologic control and an independent test of previous litho- and biostratigraphic correlations of the Arikaree Group, and (2) the basis for a refined understanding of the correlation of Arikareean faunas throughout North America.

BACKGROUND AND PREVIOUS GEOLOGIC INVESTIGATIONS

Early investigations

Arikaree sediments of western Nebraska and southeastern Wyoming were first studied at the turn of the century (1897–1909) by N. H. Darton, J. B. Hatcher, and O. A. Peterson, who developed a regional concept of Arikaree lithostratigraphy and paleontology that remains the basis for more recent geological investigations.

Darton mapped the region in 1897. In an 1899 (p. 747) report on the geology of western Nebraska, he provided the first description of the "Arikaree formation" as fine gray sands (with layers of dark gray concretions) that "contain a large amount of volcanic ash, mainly as a constituent intermixed with the sand. . . ." He also observed that Arikaree sediments formed a broad tableland extending south from the Pine Ridge to the North Platte Valley. Darton recognized the distinctive sheetlike regional geometry and volcaniclastic character of the Arikaree, two of its most important hallmarks. He identified a maximum thickness of about ~210 m (700 ft) of fine-grained gray Arikaree sandstones and siltstones in outcrops along the Pine Ridge in northwestern Nebraska, and >240 m (800 ft) in southeastern Wyoming (Darton 1899, 1905).

During his mapping of western Nebraska, Darton (1899) distinguished a series of valley fills comprising fluvial and eolian volcaniclastic sandstones locally present below the Arikaree as the Gering Formation. The Gering was sometimes separated from the Arikaree by unconformity but was otherwise continuous with it. Darton (1905, p. 177) stated that "the Gering beds represent the first deposits of Arikaree times, separable only in areas where they consist of coarse material laid down in channels." The Gering beds were incorporated in the Arikaree by Adams (1902) during a study of Arikaree rocks in southeastern Wyoming, and by the paleontologist J. B. Hatcher (1902), who examined the Arikaree in northwest Nebraska.

Until recently, the term Gering Formation has been indiscriminately employed for most bedded fluvial and eolian sediments at the base of the Arikaree Group in western Nebraska and southeastern Wyoming. Here we follow the currently proposed restriction of Darton's Gering Formation to pumice-rich volcaniclastic fluvio-eolian sediments in the Wildcat Ridge area of the southern Nebraska panhandle (Tedford et al., 1996). Fluvial and interbedded eolian sediments of the basal Arikaree Group along the Pine Ridge escarpment discussed in this chapter that were previously termed Gering Formation by Darton are excluded by us from the Gering Formation. The term "Gering" placed in quotation marks in this chapter indicates the basal fluvio-eolian volcaniclastic sandstones and stream-deposited arkosic sands and gravels of the lower part of the Arikaree Group along the Pine Ridge escarpment of northwest Nebraska.

In 1902 Hatcher divided the Arikaree into two "horizons," the Monroe Creek and Harrison beds. His principal study area was the Pine Ridge escarpment, in particular the canyon of Monroe Creek, where ~210 m (700 ft) of the Arikaree section are exposed. Here Hatcher attempted to distinguish Monroe Creek from overlying Harrison sediments using both lithic and faunal criteria, but these distinctions are inadequate and not clearly identified. According to Hatcher, the Monroe Creek and Harrison lacked any sharp contact between them, grading insensibly into each other.

Hatcher (1902, p. 117) recognized above the Harrison Formation "a series of buff-colored sandstones . . . with occasional

layers of siliceous (not calcareous) grits. . . . These beds are rich in vertebrate fossils . . . [and] are represented at various localities along the Niobrara River, south of Harrison, Nebraska. . . ." O. A. Peterson collected fossil mammals from this unit in the Niobrara Canyon. Peterson (1907, 1909) named these mammal-producing rocks the "Upper Harrison" beds based on the outstanding exposures in the canyon. He was the first to recognize that these rocks were not referable to the Nebraska Beds of Scott (1893), as Hatcher (1902) had supposed, but that they were in fact distinct.

Physiography and regional distribution of the Arikaree Group

The exposures of the Arikaree Group in northwestern Nebraska and adjacent Wyoming were the focus of the initial studies by Darton, Hatcher, and Peterson:

Pine Ridge escarpment. Pine Ridge (Fig. 1) is a steep east-west–trending linear escarpment in northwestern Nebraska and southeastern Wyoming. This prominent topographic feature forms the northern border of the Hartville Table (Hunt, 1990, Fig. 1) and the High Plains physiographic province.

As exemplified in Monroe Canyon, the Pine Ridge Arikaree sediments begin with cross-stratified fluvial sandstones and gravels filling the axis of an Arikaree paleovalley (initially considered the Gering beds by Darton in 1899 but currently regarded as a geographically disjunct, penecontemporaneous valley fill, relative to the type Gering sediments of the southern Nebraska panhandle; see Tedford et al., 1996). These sandstones and gravels in the canyon and for a considerable distance east and west along the escarpment are the only fluvial sediments present in the local Arikaree section. Fluvial deposits grade upward into several hundred meters of mas-

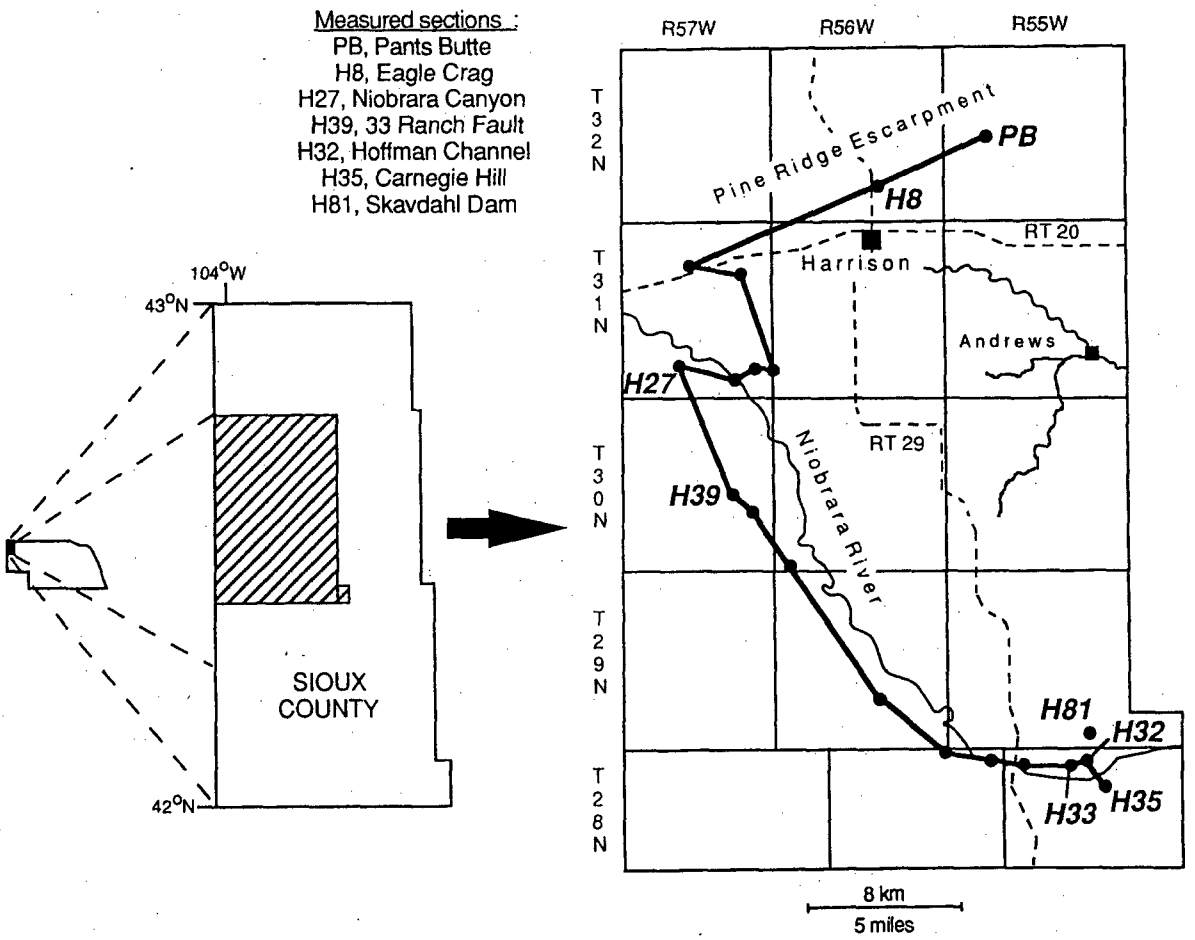


Figure 1. Map of western Nebraska showing the location of our study area in Sioux County and the position of the measured sections used to collect paleomagnetic samples during this study. The section numbers refer to the designations published in Hunt (1990). Although not one of our principal sections, H33 indicates the exact location of the Agate Ash that was dated by Evernden et al. (1964).

sive eolian siltstones and sandstones (Monroe Creek and Harrison beds of Hatcher). These eolian sediments are structureless in some parts of the section, suggesting accumulation as a volcaniclastic loess, and are vaguely cross-stratified elsewhere, with foreset bedding reminiscent of dunelike bedforms.

Along the high rim of the escarpment at the head of Monroe Canyon is a prominent paleosol 4.6m (15 ft) thick developed on the Harrison Formation. Recognizable by its densely concentrated rhizoliths and burrows, including the helical burrow fills (*Daemonelix*) of the beaver *Palaeocastor*, and with an upper surface sometimes cemented as a silcrete, this paleosol once formed a continuous paleosurface marking the end of Harrison Formation deposition. Remnants of this regional surface occur on small flat-topped hills and buttes along the summit of the ridge from Monroe Canyon westward to the Nebraska-Wyoming state boundary. This geomorphic surface is a useful marker horizon separating the Harrison Formation from the overlying rock unit, the Upper Harrison beds of O. A. Peterson.

Niobrara Canyon to Agate. At the head of Monroe Canyon and other localities along the Pine Ridge in Sioux County the terminal Harrison paleosol is overlain by a thin veneer 3 to 17 m (10–55 ft) of buff to pale brown very fine grained volcaniclastic sandstones containing occasional sheetlike lacustrine micrites and silcrete paleosols. This locally thin lithostratigraphic unit can be traced from the Pine Ridge escarpment 14.5 km (9 mi) southwest to the Niobrara Canyon (Hunt, 1990, Fig. 7) where it thickens to 45 m (150 ft) and can be identified as the Upper Harrison beds of Peterson (1907, 1909). Earlier the Upper Harrison had been referred to the Nebraska Beds by Hatcher (1902, p. 117) based on exposures "along the Niobrara River, south of Harrison. . . ." O. A. Peterson (1907) realized that this rock unit could not be referred to the Nebraska Beds of Scott (1893) and abandoned Hatcher's name. (Peterson used the name Lower Harrison for Hatcher's Harrison Formation, explaining his use of the term Upper Harrison for the overlying unit.)

The thick section of the Upper Harrison beds is the most prominent lithostratigraphic unit along the south wall of the canyon where it is particularly rich in fossil mammals, yielding the Niobrara Canyon local fauna (Hunt, 1985, 1990). Today the section exposed in the south wall is regarded as the type area for this lithostratigraphic unit (Peterson, 1907, 1909; Hunt, 1985, 1990). Petrographic studies of the Upper Harrison beds demonstrate their compositional similarity to the Pine Ridge Arikaree units, indicating these beds can be confidently assigned to the Arikaree Group (Hunt, 1990). Adding 45 m (150 ft) of Upper Harrison beds to the Pine Ridge Monroe Canyon section of ~189 m (620 ft) of "Gering," Monroe Creek, and Harrison rocks yields a composite thickness of 235 m (770 ft) for the Sioux County Arikaree section.

Recent investigations

In 1971 Hunt initiated a study of the Arikaree Group in northwest Nebraska, including the classic Arikaree sections at Monroe Canyon, the Niobrara Canyon, and Agate National Monument. Upper Arikaree sediments at the monument include the world-famous waterhole bonebed that produced remarkably preserved population samples of early Miocene rhinoceroses and chalicotheres (Matthew, 1923; Hunt, 1990). East of the monument the Arikaree section eventually disappears beneath younger rocks, and south of the monument the section is lost to downcutting by Whistle Creek, a small tributary of the Niobrara River. To the west the Arikaree beds onlap the granites and Paleozoic limestones of the Hartville Uplift in Wyoming.

Hunt's work (1978, 1985, 1990) recognizes three primary depositional intervals or units from the Pine Ridge to the Niobrara Canyon and Agate Monument: (1) "Gering"–Monroe Creek, (2) Harrison, and (3) Upper Harrison. In Monroe Canyon, the classic section reveals a late Oligocene valley fill sequence that includes "Gering" fluvial sandstones and arkosic gravels, grading upward to eolian Monroe Creek siltstones and sandstones. This sequence reflects a pattern of sedimentation that seems to be generally characteristic of regional Arikaree deposition (Hunt, 1990, Fig. 14): Arikaree valley fills include two components—a lower unit comprising laterally restricted, fluvially reworked, fine-grained volcaniclastic sediments overlain by an upper unit made up of thick volcaniclastic loess and other eolian facies that complete the filling of the valleys and extend beyond them to mantle the interfluves.

The difficulty mentioned by Hatcher (1902) in distinguishing Monroe Creek sediments from those of the Harrison Formation is attributed by us to the superposition of fine-grained eolian Harrison loess on fine-grained Monroe Creek loess; the two volcaniclastic units are similar in mineralogy and grain size. However, these eolian Harrison sediments, when traced southward to the Niobrara Canyon and to Agate, are laterally equivalent in the Agate area to fluvial and eolian sediments of diverse lithofacies containing abundant fossil mammals (Hunt, 1990). These fluvio-eolian Harrison sediments are incised into rocks of the White River Group at Agate, and the entire lower Arikaree ("Gering"–Monroe Creek) interval is missing. Thus, Harrison sediments appear to fill a wide, shallow paleovalley situated ~32 km (20 mi) south of the larger, more deeply incised lower Arikaree paleovalley at Monroe Canyon. From this paleovalley axis, Harrison sediments extend northward as a thin eolian blanket to mantle the lower Arikaree deposits of the Pine Ridge escarpment.

Early Miocene Upper Harrison sediments constitute the third and final depositional interval of the Arikaree in Sioux County. At Agate National Monument, fluvial sediments of the Upper Harrison fill a shallow paleovalley axis incised into the Harrison Formation (Hunt, 1990), attaining a maximum thickness of ~15 m (50 ft). These Upper Harrison fluvial sediments laterally intertongue with, and are overlain by, very fine grained pale brown sandstones with interbedded silcrete paleosols, interpreted as airfall volcaniclastic loess periodically stabilized by soil development. If the Upper Harrison beds are traced from the Agate area northward toward the Niobrara Canyon, fluvial sediments immediately disappear and only loess deposits with silcrete paleosols remain. In the canyon, as much as 45m (150 ft) of loess was pre-

served in a down-faulted block—this is the thickest section of Upper Harrison rocks currently known, explaining its designation by Peterson (1909) as his type section for the formation.

In the Niobrara Canyon, below the thick volcaniclastic Upper Harrison loess, are occasional lenses of sandy calcareous white tuff at the base of the formation. These waterhole sediments are often the source of mammalian bonebeds (Hunt, 1978, 1990). To the north of the Niobrara Canyon, the Upper Harrison has been removed by erosion along much of the Pine Ridge, but where present it includes volcaniclastic loess with silcrete paleosols, thin lacustrine micrites, claystone, and vitric tuffs.

PALEOMAGNETISM

Field and laboratory methods and analyses

We selected the Pine Ridge and Niobrara Canyon sections of the Arikaree Group for paleomagnetic sampling because of the thick, continuous, vertical outcrops, superposition of key units, ability to demonstrate lateral continuity from Pine Ridge to Niobrara Canyon, and the historical association with earlier studies. These rocks have also produced important collections of fossil mammals. North of the village of Harrison, near the summit of Pine Ridge, the Eagle Crag Ash (19.2 ± 0.5 Ma), calibrates the paleomagnetic zonation (Hunt et al., 1983). These northern sections correlate to those at Agate National Monument by walking out nearly continuous outcrops from the Niobrara Canyon to Agate (Hunt, 1985, 1990). At Agate National Monument the fossil mammal quarries and Agate Ash (21.9 Ma) (Evernden et al., 1964) provide biostratigraphic and radiometric control.

Three separately oriented paleomagnetic samples were collected from each of 266 horizons within the Arikaree Group of Sioux County, Nebraska. All sites are stratigraphically located within our measured sections (Fig. 1). The vertical spacing between superposed sites varied depending on the individual section; in the more densely sampled sections it was ~1 m, it averaged ~1.5 m, and in no case was it >5 m.

Paleomagnetic samples were cut into 2.5-cm cubes for analysis in the Paleomagnetics Laboratory at the University of Florida. Most of the samples were analyzed in an SCT or 2G cryogenic magnetometer (some relatively strongly magnetized samples were measured in a Schonstedt spinner magnetometer). In order to determine the dominant carrier of the remanence, 20 samples, each from a different site and representative of the variety of lithologies collected, were subjected to isothermal remanence (IRM) experiments in fields up to 3 tesla (T). Individual samples were subjected to stepwise alternating field (AF) or thermal demagnetization, and depending on the individual sample, from 3 to 19 incremental steps of demagnetization.

In order to perform Fisher (1953) statistics on each site, one of two procedures was employed to choose the individual sample characteristic paleomagnetic direction. For samples demagnetized with numerous (ca. > 10) steps between 100° and 550° C, principal component analysis (Kirschvink, 1980) determined the characteristic direction of magnetization. For samples with few (ca. three to five) stepwise measurements, or those in which principal component analysis had unsatisfactory data scatter (i.e., MAD values ≥20), a single magnetic direction, usually between 225° and 550°C, was chosen to represent the characteristic magnetization.

Results

Magnetic mineralogy. Several petrographic and sedimentologic studies of the Arikaree Group are relevant to an understanding of the magnetic mineralogy and potential extent of diagenetic alteration of these sediments. Based on petrographic analyses, accessory or heavy minerals represent a minor (ca. 10% or less) component of the Arikaree sediments (Hunt, 1990). Magnetite is a variable portion of this latter fraction, whereas hematite is rare. One remarkable aspect of the Arikaree Group is the relatively unaltered, or only slightly altered, nature of the majority of the sediments. When diagenetic alteration of the sediments appears to be present, it almost always consists of carbonate and siliceous cements, which in later stages form caliches and silcretes. Hematitic staining of the outcrops and/or the presence of secondary hematitic cements in thin-section are uncommon (Hunt, 1990; Stanley and Benson, 1979). These previous studies suggest that detrital magnetite is the dominant carrier of the remanence.

The results of almost every IRM experiment in the laboratory for the 20 samples analyzed (see representative samples in Fig. 2A–C), indicate that magnetic saturation is acquired in applied fields of <1 T. These samples further indicate that magnetite is the primary carrier of the remanence. In a few samples, a very slight increase in magnetization (e.g., Fig. 2D), indicates the possible presence of minor amounts of hematite or goethite in certain samples.

Paleomagnetic data. Representative examples of AF and thermal demagnetization indicate two dominant patterns of demagnetization behavior represented in the Arikaree sediments: (1) In the simple case, after removal of a low coercivity or low unblocking temperature component, there is a steady and linear decay of remanence indicating what is interpreted to be the characteristic component of the NRM (e.g., Figs. 3B, C, 4A, B, 5A, B); and (2) a multicomponent magnetization also can be identified in which isolation of a characteristic component is usually interpreted to occur between about 20 and 50 mT or 200° to 575°C (e.g., Figs. 3A, 4C, D, 5C, D). For this kind of demagnetization at temperatures above 575°C, the vector decay was usually of the same polarity and general directions as the characteristic component. Above the 575°C demagnetization step in many samples (e.g., Fig. 4C), which is above the Curie Point of magnetite, there frequently is a significant drop in magnetic intensity. Above the temperature, (i.e., 590°–630°C), the magnetization remaining is usually between 1 and 0.1% of the NRM intensity; this further indicates that hematite contributes a minor component of the total magnetization.

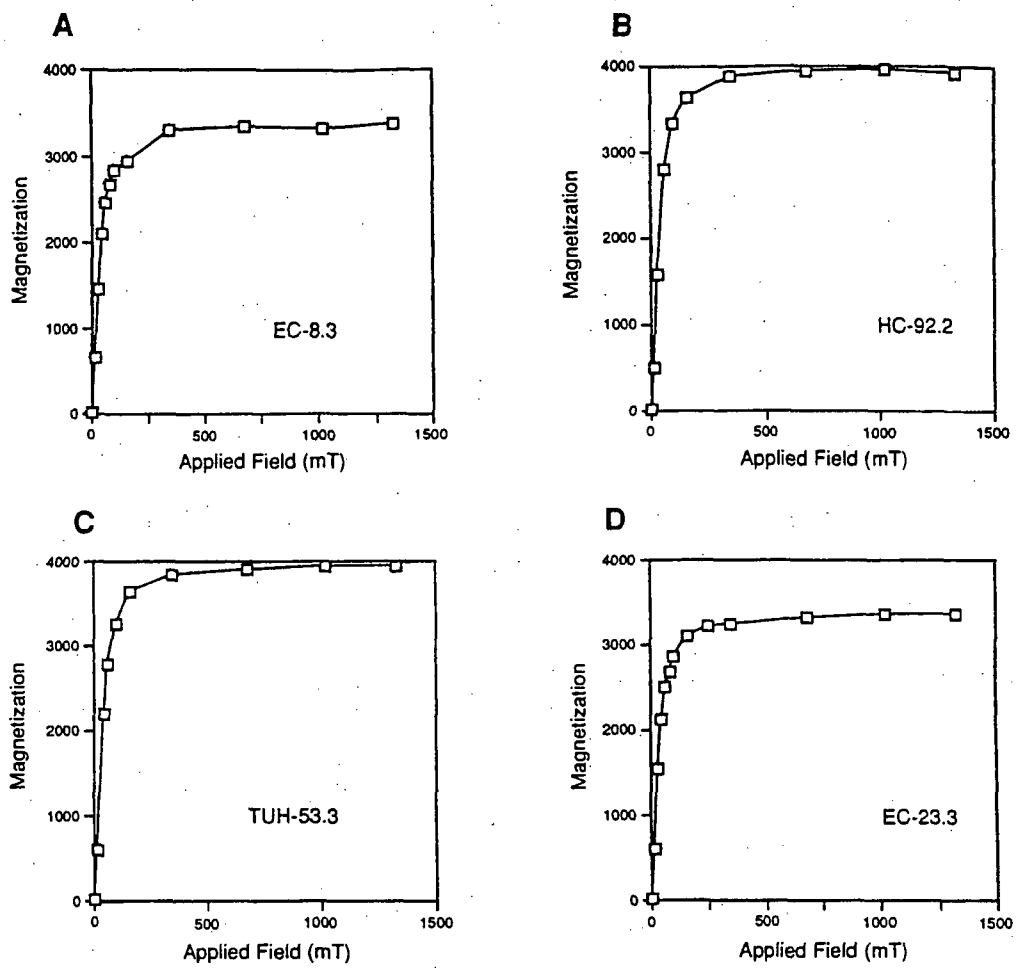


Figure 2. Isothermal remanent magnetization (IRM) curves for four sediment samples from four different paleomagnetic sites. These data indicate that the dominant magnetic mineral of the Arikaree Group in the region sampled is magnetite.

In order to determine which demagnetization regime more effectively isolated a stable remanence, AF and thermal demagnetization was done for different samples from each site. In the best-case results, the polarity and characteristic component of magnetization were similar for both the AF and thermal samples (Fig. 5). However, in numerous cases AF demagnetization was ineffective in removing a normal overprint on sites that thermal results indicated to be of reversed polarity. These latter results are interpreted to represent a high-coercivity secondary magnetization. Accordingly, the majority of those samples originally AF demagnetized were also subsequently thermally demagnetized in about four to six steps of temperatures between 200°-500°C. In general, thermal demagnetization was more effective at isolating a stable, characteristic component of the NRM. This is consistent with many other studies of the magnetostratigraphy of continental clastic sediments (Opdyke, 1990; Butler, 1992).

The mean NRM intensity for the Arikaree Group is 8.97 A/m (N = 427); those for selected demagnetization steps are 0.26 A/m (N = 7) at 40 mT, 4.25 A/m (N = 13) at 300°C, 1.07 A/m at 500°C (N = 51), and 0.09 A/m (N = 10) at 600°C. The great drop in magnetic intensity at temperatures above 575°C (i.e., the magnetization at 600°C is characteristically between 0.1 and 0.01% of the total NRM intensity) further confirms the IRM interpretation that hematite is only a minor carrier of the remanence.

An analysis of the paleomagnetic data is presented in Table 1. After Fisher statistical analysis and site quality classification (Opdyke et al., 1977), 216 of the 266 (i.e., 81%) originally sampled horizons were used to interpret the magnetostratigraphy.

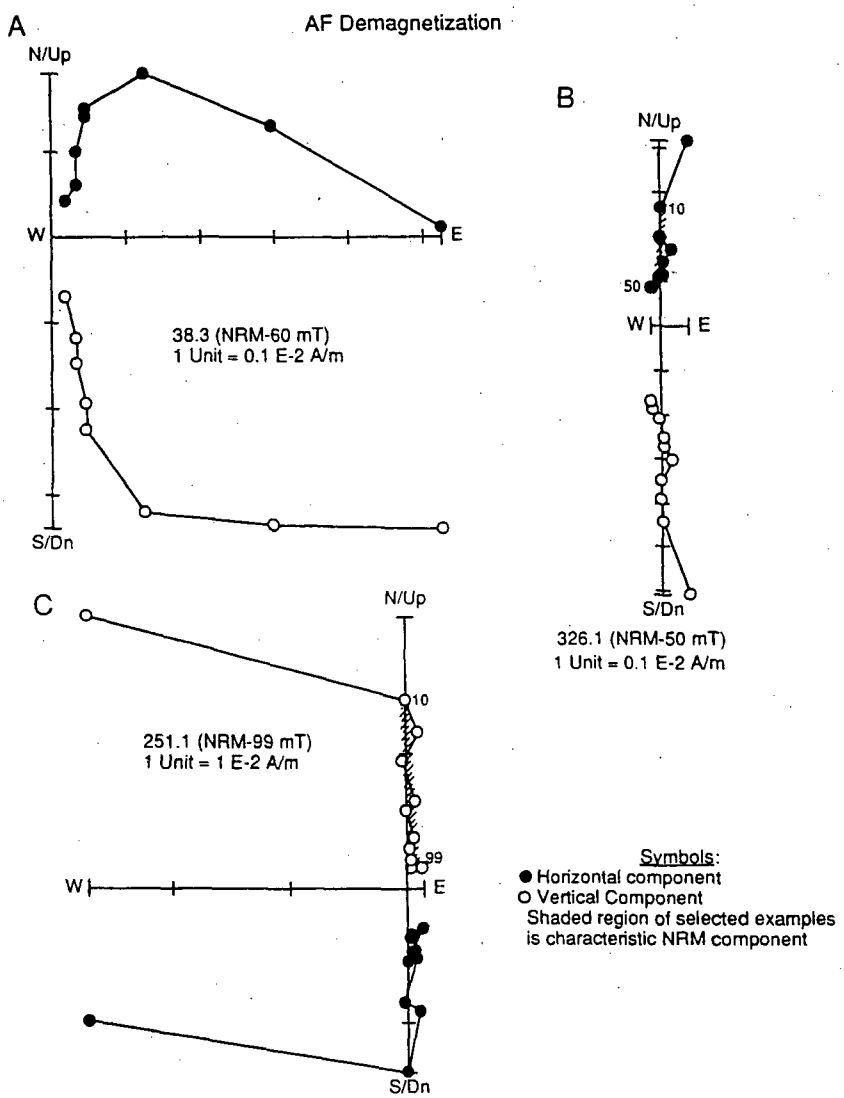


Figure 3. Orthogonal vector demagnetization (Zijderveld) diagrams showing the decay of remanence and interpretation of characteristic component of magnetization for samples from the Arikaree Group treated by AF demagnetization.

This includes 150 Class 1 sites, 53 Class 2 or 3 sites, and 13 Class 4 sites (*sensu* Opdyke et al., 1977). Class 1 sites (three samples with concordant directions) provide the strongest indication of polarity, whereas Class 2 (two samples available) or Class 3 (two of three samples with concordant directions) sites are only used in conjunction with superposed sites of the same polarity to define polarity zones, that is, a single polarity zone is never defined on one Class 2 or 3 site. Class 4 sites are the weakest indicators of polarity. In this chapter, Class 4 sites are used for reversed sites from which one or two samples trend toward, but are not definitely, reversed (this is interpreted by the presence of negative inclinations after demagnetization). Class 4 R sites are used only to further confirm the polarity of the nearest superadjacent or superposed sites. Class 4 N sites are not used for polarity information because the tendency for a site to be normal in many cases could indicate equally well a secondary overprint of the original magnetization.

The Class 1 site mean NRM data (Fig. 6A) show that the individual directions are relatively dispersed, lie mostly in the Northern Hemisphere, and are mostly of positive inclination.

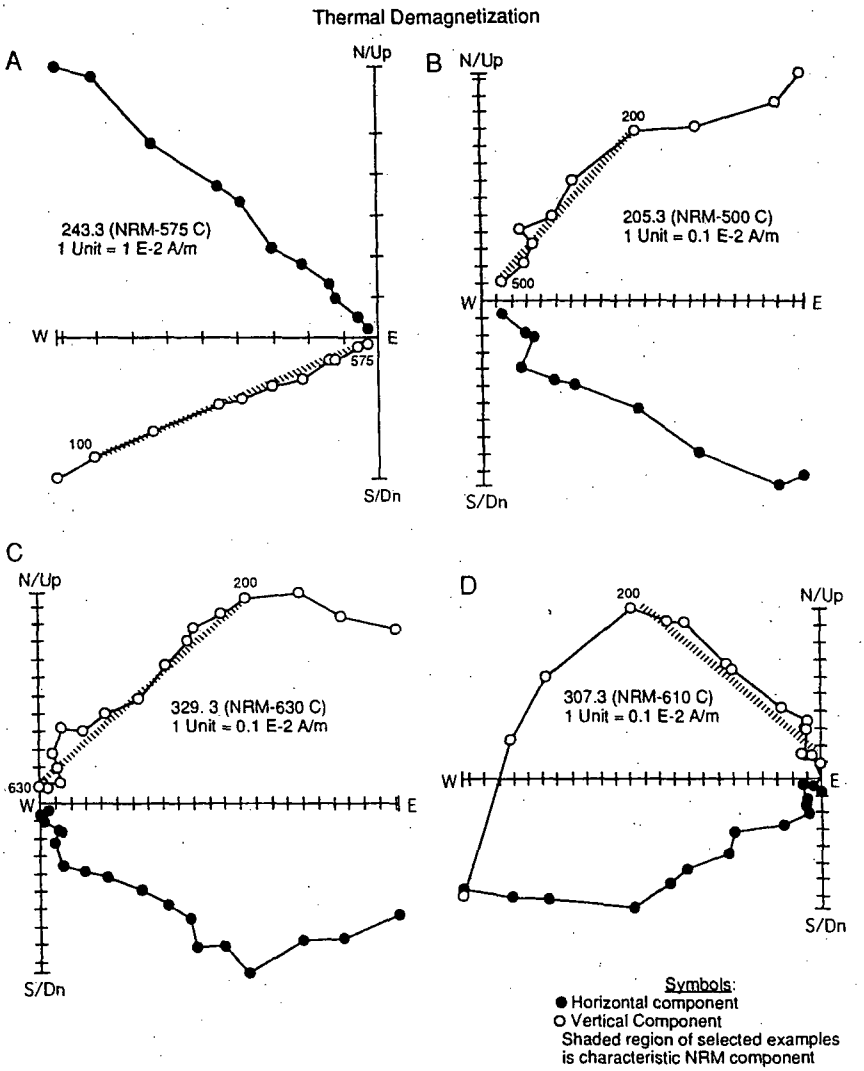


Figure 4. Orthogonal vector demagnetization (Zijderveld) diagrams showing the decay of remanence and interpretation of characteristic component of magnetization for samples from the Arikaree Group treated by thermal demagnetization.

After demagnetization (Fig. 6B), the site means fall into two populations, one in the Northern Hemisphere with positive inclinations (interpreted as normal-polarity sites) and a second group lying in the Southern Hemisphere with negative inclinations (interpreted as reversed-polarity sites). The mean value of these 137 sites (with 13 sites of intermediate directions excluded) after demagnetization (with the reversed sites inverted to the Northern Hemisphere), is declination 356.1°, inclination 59.1° (R = 121, α_{95} = 4.4°, $\underline{k}$ = 8.6). The mean of the normal sites (declination = 2.0°, inclination = 66.2°, N = 90, R = 84.3, α_{95} = 3.9°, $\underline{k}$ = 15.6) and mean of the reversed sites (declination = 169.0°, inclination −43.3°, N = 47, R = 39.2, α_{95} = 9.4°, $\underline{k}$ = 5.9), although roughly antipodal, do not overlap (Fig. 7). The most probable reason for this is because the reversed sites are of shallower inclination than either the normal sites or predicted ancient dipole axis calculated for this latitude (61.4°, indicated by cross in Fig. 7). We interpret these results to indicate that the reversed sites exhibit incomplete removal of an overprint that resulted in the relatively shallower inclinations in that data set.

Opdyke (1990) and Opdyke and Channell (1996) proposed a set of criteria that rate the overall quality of magnetostratigraphic studies. Of the 10 criteria, our study of the Arikaree Group

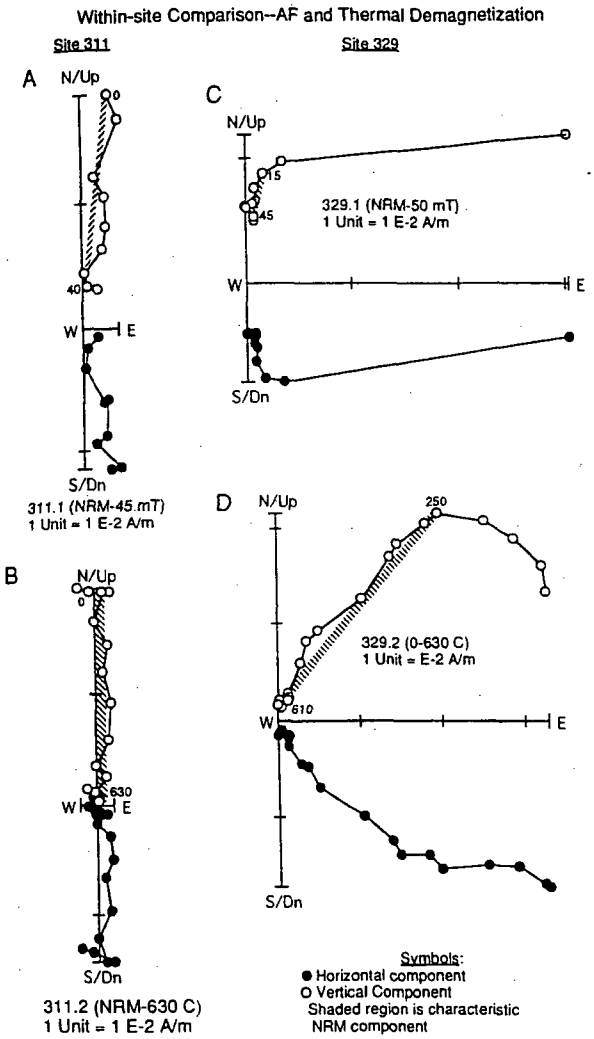


Figure 5. Orthogonal vector demagnetization (Zijderveld) diagrams showing a comparison of the decay of remanence and interpretation of characteristic component of magnetization for samples that have been AF demagnetized and thermally demagnetized for two paleomagnetic sites from the Arikaree Group.

meets 7: (1) stratigraphic age known to the level of Cenozoic stage, (2) sampling localities placed in a measured stratigraphic section, (3) complete thermal or AF demagnetization performed and vector analysis carried out using orthogonal plots, (4) data published completely, (5) magnetic mineralogy determined, (6) associated paleontology presented adequately, and (7) radiometric ages are available in the stratigraphic sections. We are unable to satisfy two criteria: field tests for stability and antipodal reversals. This results from what seems to be an incomplete removal of a normal overprint for the reversed population (also see the discussion above). The tenth criterion—directions determined from principal component analysis—is only partially satisfied in some of our sites. Opdyke (1990, p. 622) stated that: "ratings of 5 [or more] should be achieved by modern magnetostratigraphic studies." Accordingly, the seven criteria fulfilled in our study, despite the negative reversals test, satisfy the acceptability criteria proposed by Opdyke (1990) and Opdyke and Channell (1996).

RADIOISOTOPIC AND BIOSTRATIGRAPHIC FRAMEWORK

Radioisotopic age determinations

Two radiosiotopic age determinations are currently available for tuffs from the Arikaree Group in Sioux County; both of these are located in our measured sections. The first date published was a ^{40}K-^{40}Ar (biotite) age determination of 21.3 Ma (Evernden et al., 1964: KA sample 481) that has subsequently been corrected to 21.9 Ma based on the new decay constants (Dalrymple, 1979). (Note that no standard error margin was published with the orginal date.) This tuff was sampled by M. F. Skinner at an exposure 3 m (10ft) thick on the north side of the entrance road to Agate National Monument (NW4, SW4, SE4, SW4, sec. 4, T. 28 N., R. 55 W.). Along this road are a number of laterally continuous outcrops of the Agate Ash in sec. 4, T. 28 N., R. 55 W. (Agate and Whistle Creek NW 7.5′ Quadrangles). The outcrop sampled by Skinner for radioisotopic dating and the outcrop in our Hoffman Channel section (Fig. 12, below) are taken from the same laterally continuous bed of Agate Ash (see Hunt, 1990, Fig. 7, sections H32–H33).

The second age determination using the fission-track method (zircon), yielded an age of 19.2 ± 0.5 Ma from the Eagle Crag Ash (Hunt et al., 1983). The sampled outcrop of the Eagle Crag Ash is located in a roadcut on the east side of the Monroe Canyon Road, 2.64 km (1.65 mi) north of the town of Harrison in the W1/2, NW4, SW4, SE4, sec. 27, T. 32 N., R. 56 W., Harrison West 7.5′ Quadrangle. This tuff is located ~0.6 m (2 ft) above the contact between the Harrison Formation and Upper Harrison beds and ~50 m (162 ft) above the base of our Eagle Crag measured section (Fig. 9). Fission-track dating was carried out on a population of uniformly sized euhedral zircon crystals lacking any indication of detrital zircons in the samples.

Biochronologic framework

Fossil mammals are present throughout the Arikaree Group of western Nebraska and southeastern Wyoming. Concentrations of fossils are most common in fluvial sediments and waterhole paleoenvironments, whereas the massive volcaniclastic sandstones and siltstones interpreted as eolian loess yield fewer remains (Hunt, 1985). Nevertheless, continued collecting of mammals over many years demonstrates a succession of faunas within the Arikaree strata of the region that permits the establishment of a reliable biochronology for the late Oligocene–early Miocene of the North American midcontinent.

TABLE 1. PALEOMAGNETIC DATA AND SITE QUALITY ANALYSIS FOR THE MEASURED SECTIONS AND PALEOMAGNETIC TRANSECTS MEASURED AND SAMPLED FROM THE ARIKAREE GROUP, SIOUX COUNTY, NEBRASKA

Measured	Sites Total	Class 1 Sites*	Class 2 or 3 Sites	Class 4 Sites	Sites Used Total and (%)	Polarity Transitions	Sampling Density
Pants Butte	43	24	12	2	38 (88)	4	0.11
Eagle Crag (H8)	42	35	3		38 (90)	1	0.03
Niobrara Canyon (H27)	52	29	8	2	39 (75)	≥5	0.13
33 Ranch Fault (H39)	18	7	4		11 (61)	0	.
Hoffman Channel (H32)	25	12	6	2	20 (80)	≥5	0.26
Carnegie Hill (H35)	24	12	5	3	20 (83)	≥3	0.16
Skavdahl Dam (H81)	18	12	1		13 (72)	≥2	0.17
Others (including H26, H28/9, H31, and H33)	44	19	14	4	37 (84)		
All section totals	266	150	53	13	216 (81)		

*Includes 13 sites of intermediate (easterly or westerly) declinations and usually very shallow inclinations (usually less than −10) not used to calculate formational mean statistics nor are these 13 sites used in Figures 6 and 7.

Using the formula of Johnson and McGee, 1983, where the probability sampling density $p = R/(N-1)$ where R is the number of polarity transitions encountered and the N = number of sites used to construct a magnetic polarity zonation within a particular section. The lower the p value, the higher the probability that all polarity transitions have been sampled within a given section.

In the past the term "Gering" has been applied to fluvial sediments at the base of the Arikaree Group along the length of the Pine Ridge escarpment in northwest Nebraska, despite the fact that the type Gering Formation of Darton (1899) is restricted to the Wildcat Ridge of the southern Nebraska panhandle. The Gering sediments of Wildcat Ridge are relatively rich in fossil mammals, resulting in a preliminary biostratigraphic zonation (Tedford et al., 1985, 1996), whereas the "Gering" sediments of the lower Arikaree paleovalley of the Pine Ridge in Sioux County are not known to be particularly fossiliferous. Most "Gering" fossil mammals from the Pine Ridge escarpment come from basal Arikaree fluvial sediments near the town of Chadron in Dawes County, east of the study area. Only a few oreodonts were recorded by Peterson (1907) near Chadron, but two recently discovered quarry sites south and west of the town have yielded mammals in stream sediments in some abundance, indicating at least an approximate age equivalence of the Pine Ridge "Gering" to the Wildcat Ridge Gering beds.

The lower Arikaree fluvial deposits at Pine Ridge in Sioux County grade upward into massive volcaniclastic loess. The loess is Hatcher's (1902) Monroe Creek Formation, and along the Pine Ridge it has produced isolated mammalian remains, most often oreodonts and camels, as well as infrequent rodents, carnivores and rhinoceroses. These lower Arikaree "Gering" and Monroe Creek beds in Sioux County represent a single depositional interval comprising fluvial volcaniclastics grading upward to, and intertonguing with, eolian volcaniclastics, and are characterized by a sparse Early Arikareean fauna (Peterson, 1907).

The Harrison Formation of the Pine Ridge in Sioux County is entirely eolian. Mammals in such depositional settings are scarce, but the loessic lithofacies overprinted by paleosols at the head of Monroe Canyon has produced rare remains of the chalicothere *Moropus*, the rhinoceros *Menoceras*, and the camel *Stenomylus*, characteristic taxa of the Late Arikareean fauna of Sioux County. To the west near the town of Van Tassell, Wyoming, the Pine Ridge outcrops of the Harrison Formation yielded oreodonts such as *Promerycochoerus*, *Merychyus*, and *Hypsiops* from loess deposits. More abundant samples of mammals from the Harrison Formation occur in the mixed fluvial-eolian lithofacies of Agate National Monument and adjacent sites along the Niobrara River in Sioux County. These localities produce chalicothere, stenomyline camels, the rhinoceroses *Menoceras* and *Diceratherium*, equids, temnocyonine beardogs, the oreodont *Hypsiops*, and small canids, but lack Early Arikareean last occurrences such as leptauchenine oreodonts, the giant oreodont *Megoreodon*, nimravid cats, hyaenodonts, the amphicyonid *Paradaphoenus*, and archaeothere entelodonts. From the Pine Ridge to Agate, the Harrison Formation contains the helical burrows (also called *Daemonelix*, or "devil's corkscrews") of the fossorial beaver *Palaeocastor*. These burrows often occur in communities that collectively establish the location of ancient land surfaces within the Harrison volcaniclastic loess, demonstrating the episodic nature of loess sedimentation over time.

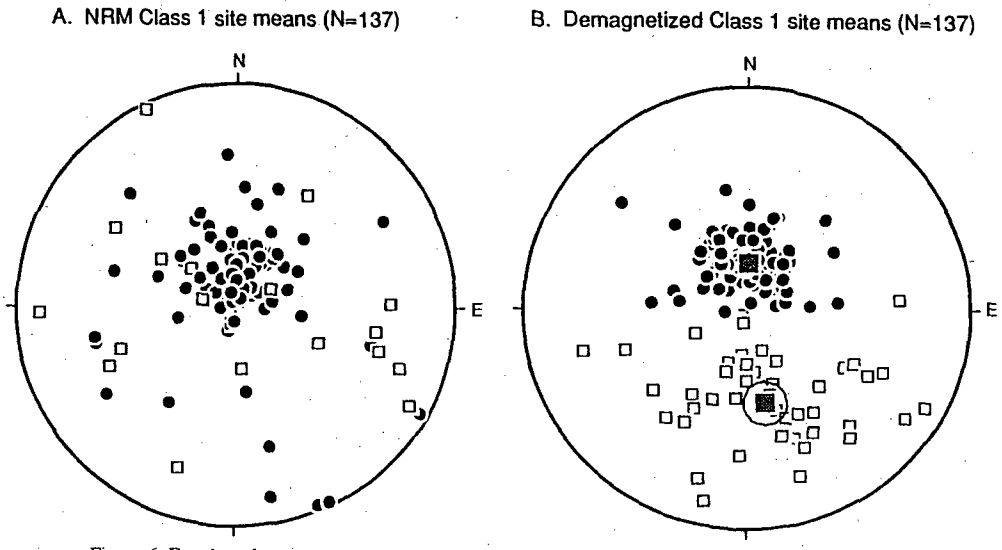


Figure 6. Equal-angle stereographic projections of 137 Class 1 site mean data for the natural remanent magnetization (NRM) (A), and after demagnetization (B). (Thirteen sites that satisfy the Class 1 criteria in Table 1, i.e., statistically significant but of intermediate direction and/or low inclination (less than ±10°), are not included in this plot). In (B), the square in the northern hemisphere represents the mean of all N-polarity sites (α_{95} = 3.9° is too small to see); the square in the southern hemisphere represents the mean of all R-polarity sites, and the circle surrounding the mean is the 95% cone of confidence (α_{95} = 9.4°). Black circles represent positive inclination; open squares, negative inclination.

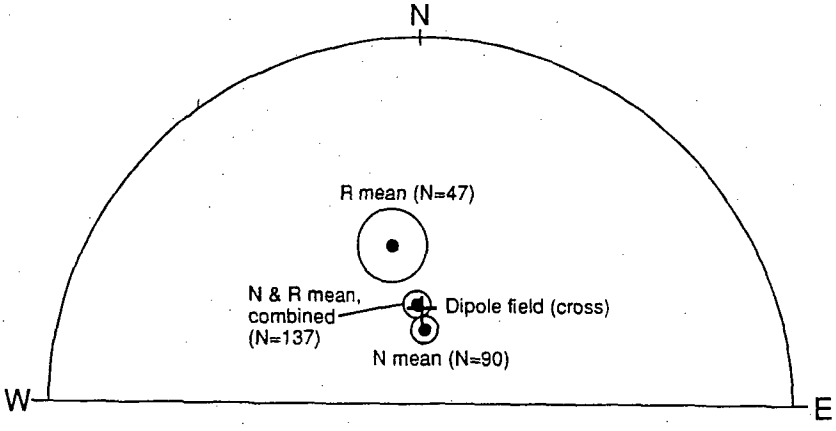


Figure 7. Stereographic projection of the mean for all Class 1 sites, all Class 1 R sites, all Class 1 N sites, and the inclination of the dipole field calculated for the collecting latitude (42.5°N).

The Upper Harrison beds of the Niobrara Canyon yield a prolific fauna of latest Arikareean mammals. The principal outcrops along the south rim of the canyon produce isolated remains from massive volcaniclastic loess and silcrete paleosols. Stratigraphically below these deposits, basal Upper Harrison sediments in the canyon (Hunt, 1978) contain rich bonebeds (Harper Quarry) that accumulated in ancient waterholes comparable to the Agate bonebed in basal Upper Harrison sediments at the national monument 40 km (25 mi) to the south. Whereas oreodonts and camels are the most common mammals in the loess and paleosols, the waterholes contain diverse carnivores, horses, chalicothere, entelodont, and rhinoceroses, with rarer remains of camels, oreodonts, and small mammals. Upper Harrison mammals along the Niobrara River at Agate National Monument and vicinity are largely confined to the waterhole settings and to a unique carnivoran den site (Hunt et al., 1983; Hunt, 1990).

Overlying the Upper Harrison beds at the national monument are the lowermost deposits of the Runningwater Formation, the basal unit of the Hemingford Group. These sediments in the district northeast of Agate post office have produced an Early Hemingfordian mammal fauna of oreodonts, camels, horses, rhinoceroses, rodents, and carnivorans as isolated occurrences.

Mammals from the Arikaree and basal Hemingford sediments studied by us during paleomagnetic sampling fall into faunal groups that correspond closely to depositional intervals bounded by erosional unconformities. The fauna of each of these units displays a uniformity that contrasts with faunas from rock units above and below. Thus at present one can recognize a fauna from the "Gering"–Monroe Creek (lower Arikaree) interval, one from the Harrison Formation, a third from the Upper Harrison beds, and a fourth from the lower part of the Runningwater Formation northeast of Agate. These mammals supply a necessary biochronologic framework for the magnetostratigraphic data reported in this chapter.

MAGNETIC POLARITY STRATIGRAPHY

Paleomagnetic site-polarity results

Pants Butte section. The Pants Butte section is very important in this study because it consists of a relatively long section, 98m (320 ft) thick, with no apparent structural complexity that spans the lower part of the Arikaree Group on the Pine Ridge escarpment (Figs. 1, 8). It was chosen instead of the excellent exposures along Monroe Creek Canyon ~8km (5mi) west of Pants Butte because the canyon section contains a prominent fault (Whetstone Fault of Hunt, 1978). The lowest sediments in the section do not record the Arikaree–White River contact but are representative of lower Arikaree ("Gering") fluvial sands that fill a paleovalley axis trending west-east parallel to the Pine Ridge. These gray fluvial sediments fine upward into buff silty sandstone. In the section at 1,393 m (4,570 ft: elevations in feet are presented to key with exact elevations on U.S. Geological Survey topographic maps used to provide control on all measured sections), these deposits are overlain by compact buff Monroe Creek eolian siltstone that continues to ~1,402m (4,600 ft). From this point to 1,420m (4,660 ft) the Monroe Creek Formation comprises intertonguing lenses of compact buff to gray silty sands and sandy silts. From 1,420m (4,660 ft) to about 1,460 m (4,790 ft), the upper part of the Monroe Creek is a lithically monotonous loess: a very fine silty gray sandstone that appears to be the sparsely fossiliferous unit termed "upper Monroe Creek" by Peterson (1907). From 1,460 m (4,790 ft) to 1,468m (4,815 ft) fine-grained gray sandstone is overprinted by a prominent paleosol with dense concentrations of siliceous rhizoliths and burrows that may represent the Harrison Formation of Hatcher (1902).

Fortunately, the quality of the paleomagnetic site data (Table 1) from the Pants Butte section is very good, with a high sampling density (*sensu* Johnson and McGee, 1983) (Table 1) and 88% of the orginally sampled sites ultimately being used for polarity determination. These high-quality paleomagnetic data probably relate to the lack of obvious bioturbation or extensive paleosol development that are seen in some of the other measured sections containing lower quality data. The Pants Butte section consists of five well-defined (i.e., each by five or more superposed sites of the same polarity) magnetic polarity zones, R1-N1-R2-N2-R3 (Fig. 8) represented in our paleomagnetic transect. A single Class 2/3 site (no. 416), lying about midway within zone N1, is considered spurious and not used for polarity interpretation.

Eagle Crag section (H8). The Eagle Crag section is laterally continuous with outcrops at the head of Monroe Canyon used by Hatcher (1902) to typify the Harrison Formation. In addition, this section, 58m (190 ft) thick, is important because it is the longest section available to us along the Pine Ridge that spans the transition from the Harrison Formation to the Upper Harrison beds (Fig. 9). It also includes the Eagle Crag Ash in the upper part of the section.

The Harrison Formation in the Eagle Crag section is entirely eolian and includes two lithofacies. From the base of the section to 1,497 m (4,910 ft), massive fine-grained gray sandstone represents a volcaniclastic loess: its upper part from 1,482 to 1,497 m (4,862–4,910 ft) is overprinted by a paleosol with abundant rhizoliths and burrows and the entire lithofacies contains the helical burrow fills of *Palaeocastor*. From 1,497 to 1,500 m (4,911–4,919 ft) is an eolian sheet sand (dikaka), richly infiltrated by rhizoliths and burrows; this lithofacies is always the terminal unit of the Harrison Formation in its type area along Pine Ridge and its upper 0.3 to 0.6 m (1–2 ft) are often cemented into a silcrete forming the terminal Harrison paleosurface.

In the Eagle Crag section the terminal Harrison paleosurface is overlain by pale brown volcaniclastic sandstones of the Upper Harrison beds interbedded with thin lacustrine micrites and the Eagle Crag Ash. This 9 m (30 ft) interval at Eagle Crag begins with a fine-grained volcaniclastic sandstone, overlain by two micrites (with abundant aquatic pulmonate gastropods) separated by 2 m (5–6 ft) of intervening claystone, succeeded by pale brown volcaniclastic sandstones that continue upward to the top of a prominent hill above Eagle Crag, a local landmark north of the village of Harrison. On the southwest side of this hill, the Eagle Crag Ash occurs 0.6 m (2 ft) above the Harrison–Upper Harrison contact. This friable vitric tuff, containing numerous well-preserved euhedral crystals of zircon and apatite, is found at the top of the long normal-polarity interval in our Eagle Crag section and has been dated by the fission-track method (zircon) at 19.2 ± 0.5 Ma (Hunt et al., 1983).

Unfortunately, an exact lithostratigraphic correlation between the Pants Butte and Eagle Crag sections currently cannot be established. Although there is topographic overlap between the sections, the presence of the Harrison Formation remains uncertain at the top of the Pants Butte section: it seems that a hiatus of unknown duration occurs between the top of the Pants Butte transect and the base of the Eagle Crag sampled interval.

Two magnetic polarity zones, a relatively long N1 and short R1, are represented in the Eagle Crag section. This section is

characterized by high-quality data, that is, 90% of the orginally sampled sites were used to establish the magnetic polarity zonation and the sampling density is high (Table 1). The Eagle Crag Ash, dated at 19.2 ± 0.5 Ma (Hunt et al., 1983), is located in the top of zone N1, about 3 m (10 ft) below the N1-R1 transition.

Niobrara Canyon section (H27). This section (Fig. 10) from the south wall of the canyon includes the type Upper Harrison beds of Peterson (1907, 1909) that comprise the interval from 1,473 to 1,509 m (4,834–4,952 ft). The Upper Harrison beds here are represented by massive pale brown volcaniclastic loess with interbedded silcrete paleosols (nine are recorded in H27). The lower 1.8 to 3 m (6–10 ft) of the Upper Harrison beds in this section display more varied lithologies including thin lacustrine micrite, tuff, and gray sandstone with paleosols having well-developed rhizolith-burrow complexes. In the Niobrara Canyon the contact between the Harrison and Upper Harrison can be recognized as a surface of nondeposition or erosion above the terminal Harrison paleosol. At several localities in the Niobrara Canyon, channels and depressions incised in the Harrison paleosol are filled with fine-grained calcareous tuff. At Boggs Butte and Harper Quarry these waterhole tuffs contain abundant Late Arikareean fossil mammals (Hunt, 1978). Below the eolian sand sheet at the top of the Harrison Formation are intervals of foreset bedding indicating the presence of dune bedforms; below the dunes the massive Harrison volcaniclastic loess lithofacies also present at Eagle Crag continues to the base of the section.

The 61-m-long (200ft) Niobrara Canyon section consists of a paleomagnetic sampling transect with six relatively well-defined magnetic polarity zones: N1-R1-N2-R2-N3-R3 (Fig. 10). There is a short interval (consisting of sites 51–53) between zones N1 and R1 in which the paleomagnetic data yielded uninterpretable results. The contact between the Harrison Formation and Upper Harrison beds occurs in magnetic zone N2 and the top of zone N2 correlates with the top of N1 at the Eagle Crag section. Sites 255 and 77 in the middle of the uppermost magnetic polarity zone R3 are of normal polarity. Site 255 was collected from a heavily weathered paleosol and, given the dominantly reversed signature for the upper part of the Upper Harrison beds in the other measured sections, it is not used for polarity interpretation. Likewise, site 77 of normal polarity within a predominantly reversed-polarity zone of the Upper Harrison beds is here considered to have a spurious normal overprint. Given the episodic sedimentation of the Arikaree Group, another possible explanation is that sites 255 and 77 record a short-period normal-polarity zone not recorded elsewhere in the dominantly reversed Upper Harrison beds.

The normal magnetozone N3 (samples 249, 70–73, 252) does not appear in any other sampled section of the Upper Harrison beds. Possibly the very thick Niobrara Canyon section (H27) records a normal zone not represented by sediments elsewhere in the formation (we find it surprising, however, that this normal interval is not recorded in the apparently equivalent section H39 where lithostratigraphic correspondence suggests it should occur). Zone N3 in the Niobrara Canyon is only ~6 m (20 ft) thick: it is possible that it may occur within an interval of nondeposition in other sections.

33 Ranch Fault section (H39). This section closely corresponds in thickness and lithofacies to the type Upper Harrison section (H27) in the Niobrara Canyon. It consists almost entirely of pale brown volcaniclastic loess with interbedded silcrete paleosols similar in number and sequence to those of the canyon. The Harrison–Upper Harrison contact is identified by the terminal paleosol of the Harrison Formation overlain by low-angle cross-strata that mark the beginning of Upper Harrison deposition. Although the pale brown loess in the canyon is especially fossiliferous, the loess here is generally barren of fossils.

The paleomagnetic data from this 44-m-thick (143 ft) measured section (Fig. 11) are of relatively lower quality; only 61% of the originally sampled sites were used for polarity interpretation (Table 1). The entire section, including the transition from the Harrison Formation to the Upper Harrison beds, appear to be of reversed polarity, and are designated zone R1. While the reversed signature of the Upper Harrison beds is consistent with other sections (except H27), the reversed polarity below and above the Harrison–Upper Harrison contact is different from the polarity seen at the four other sections that sample this interval. Given the overall low quality of the data in the 33 Ranch Fault section, and the fact that the lower two sites within the Harrison Formation in the 33 Ranch Fault section are Class 2/3 R, this polarity is considered spurious and it is not used below to interpret the composite magnetic polarity stratigraphy of the Arikaree Group.

Hoffman Channel section (H32). The lower 10.7 m (35 ft) of the section are fluvial and eolian beds of the Harrison Formation that contain the Agate Ash (Evernden et al., 1964). The ash is a ponded vitric tuff with few crystals and it varies in thickness from a feather edge to 3 m (10 ft) locally, filling an irregular paleotopography. The Harrison Formation in section H32 at Agate National Monument is incised by a wide, shallow stream channel filled with about 5.5 m (18 ft) of Upper Harrison fluvial sandstone, tuff and air-fall units overprinted by paleosols. Overlying the channel fill from 1,368 to 1,378 m (4,490–4,520 ft) is pale brown air-fall loess capped by the Agate Limestone, here a lacustrine micrite 1.2m (4 ft) thick that serves as a local marker bed. This section contains a succession of stratigraphic units nearly identical and presumably time-equivalent to those in the Carnegie Hill section.

This 26-m-thick (84 ft) section consists of six magnetic polarity zones, N1-R1-N2-R2-N3-R3 (Fig. 12). One of the zones, R2, is defined by only one Class 1 reversed site, which would not normally be used to define a polarity zone. However, the presence of this zone in the closely correlative Carnegie Hill section, which is located just 1.75 km (1.1 mi) to the southeast (Fig. 1), corroborates its validity. The Agate Ash, dated at 21.3 Ma using ^{40}K-^{40}Ar (biotite) by Evernden et al. (1964) and corrected to 21.9 Ma (Dalrymple, 1979), is located 1.2 m (4 ft) above the base of the measured section in magnetic polarity zone N1. The contact between the Harrison Formation and the Upper Harrison beds occurs at the transition between zones R2 and N3. The

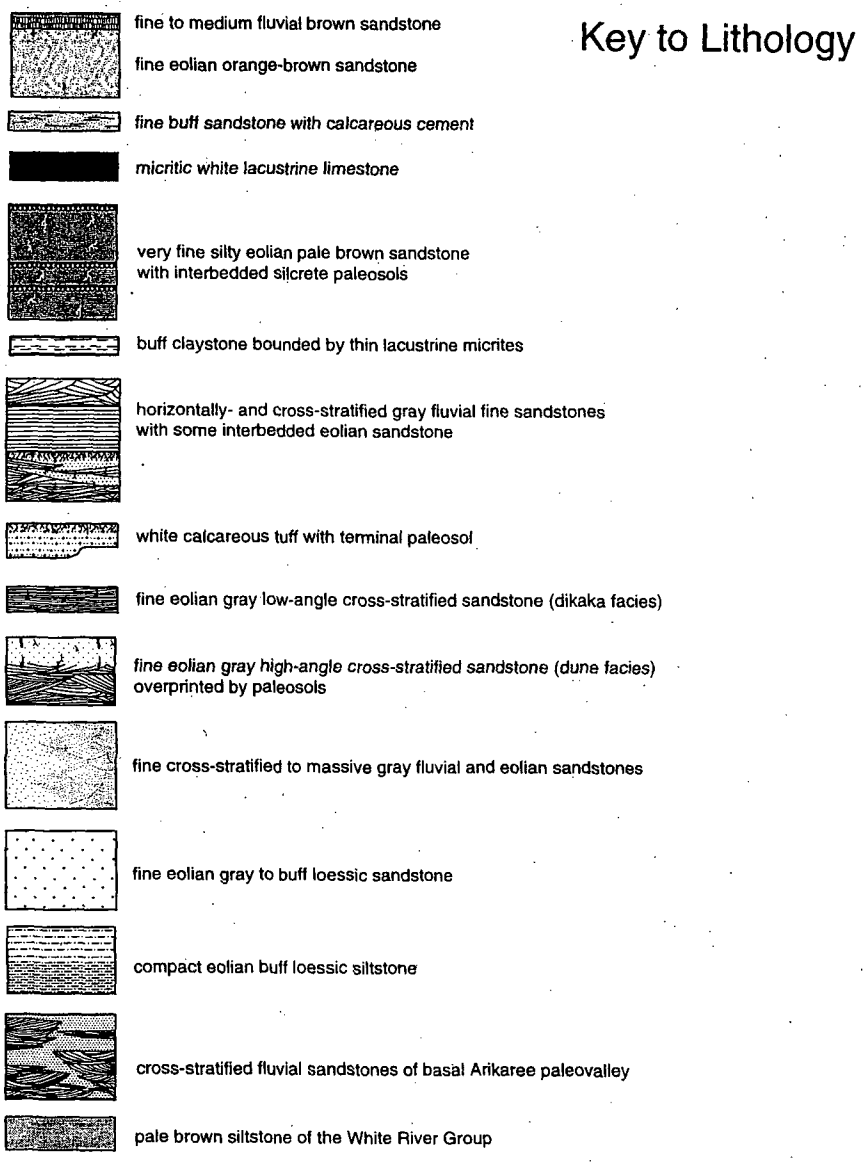
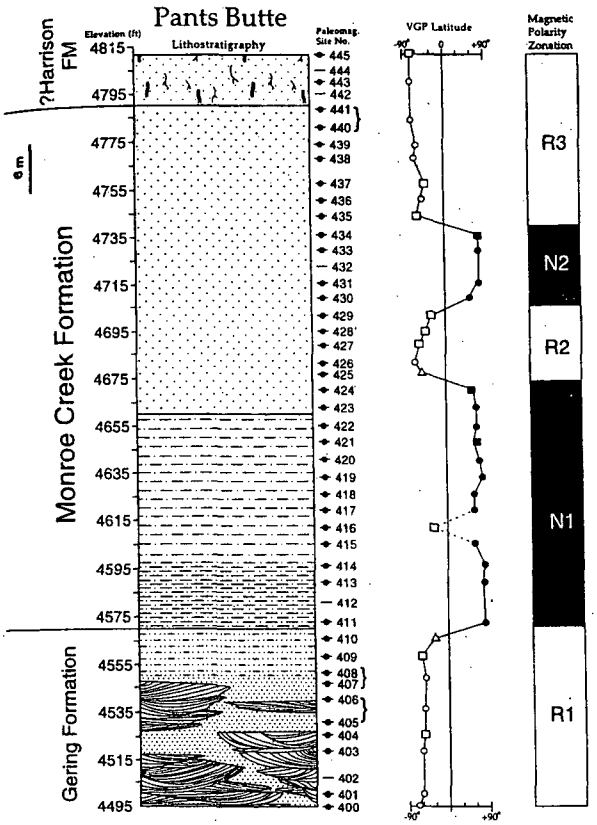


Figure 8 (on this and facing page). Litho- and magnetostratigraphy of the Pants Butte section. See Figure 11 for symbols.

lower part of the Upper Harrison beds is characterized by normal polarity overlain by a zone of indeterminant polarity between sites 92 and 95 (the latter mostly consists of either Class 2/3 or 4 sites and/or of shallow VGP latitudes). As is also the case in the Eagle Crag, 33 Ranch Fault, and Carnegie Hill sections (Fig. 15), the remainder of the overlying Upper Harrison beds are characterized by a relatively long zone of reversed polarity, i.e., R3 in the Hoffman Channel section. (For site 99 in the Hoffman Channel section, see the discussion below on the Agate Limestone.)

Carnegie Hill section (H35). This section (Fig. 13) unambiguously correlates to the Hoffman Channel section 1.75 km (1.1 mi) northwest of Carnegie Hill. The two sections are offset by 24 to 30 m (80–100 ft) due to faulting and uplift of the Carnegie Hill section. Sediments of the Harrison Formation at Carnegie Hill are dark gray sandstones lithically indistinguishable from those at Hoffman Channel and are similarly incised by a wide, shallow channel filled with Upper Harrison calcareous tuff and fine sand that contains the Agate waterhole bonebed (Hunt, 1990). The bonebed tuff from 1,384 to 1,386 m (4,543–4,548 ft) terminates in a paleosol that is overlain by about 6 m (20 ft) of unfossiliferous



cross- and horizontally laminated fluvial sandstones that complete the filling of the wide Upper Harrison paleovalley. These fluvial sediments are overlain by ~9 m (30 ft) of pale brown volcaniclastic loess capped by the Agate Limestone.

This 18.5-m-thick (56 ft) section consists of what are interpreted below to represent at least four magnetic polarity zones, N1-R1-N2-R2 (and possibly a fifth, ?R2 or R3 in Fig. 13) with an interval of indeterminant polarity for paleomagnetic sites 109 and 110 (both of which are Class 4 sites with shallow VGP latitudes). As is the case with the Hoffman Channel section (Figs. 12,15), the contact between the Harrison Formation and Upper Harrison beds occurs at an R to N polarity transition (here R1 to N2). The incredibly rich Agate waterhole bonebed also lies just above the Harrison/Upper Harrison contact at the transition between magnetic zones R1 and N2. As far as the data indicate (i.e., considering the indeterminant part of the section), the Upper Harrison above zone N2 is dominantly of reversed polarity.

The Harrison Formation south of the Niobrara River at the national monument rests with erosional disconformity on brown volcaniclastic siltstone of the White River Group.

Skavdahl Dam section (H81). Harrison fluvial and eolian sediments are incised here by the northern channel margin of the same shallow stream present at Carnegie Hill and Hoffman Channel. However Upper Harrison fluvial sandstones are only ~1 to 4 m thick (3-13 ft) at most and are overlain by an abbreviated thickness of pale brown volcaniclastic loess capped by the Agate Limestone. Above the limestone is ~1.8 m (6 ft) of calcareous sandstones that are interpreted as the terminal unit of the Upper Harrison beds. From 1,378 to 1,397 m (4,523–4,582 ft) the youngest pre-Quaternary stratigraphic unit of the national monument is represented by orange-brown volcaniclastic sandstones (loess) and interbedded fluvial sheet sandstones of the lower part of the Runningwater Formation, the basal unit of the Hemingford Group. This formation is fossiliferous northeast of Agate, where ~30 m (100 ft) of section occurs over a wide area of about two townships.

This 42-m-thick (137 ft) section (Fig. 14) includes the upper part of the Harrison Formation, the Upper Harrison beds, and the overlying Runningwater Formation, although our paleomagnetic transect includes only samples from the upper two units. This section is lithostratigraphically important because

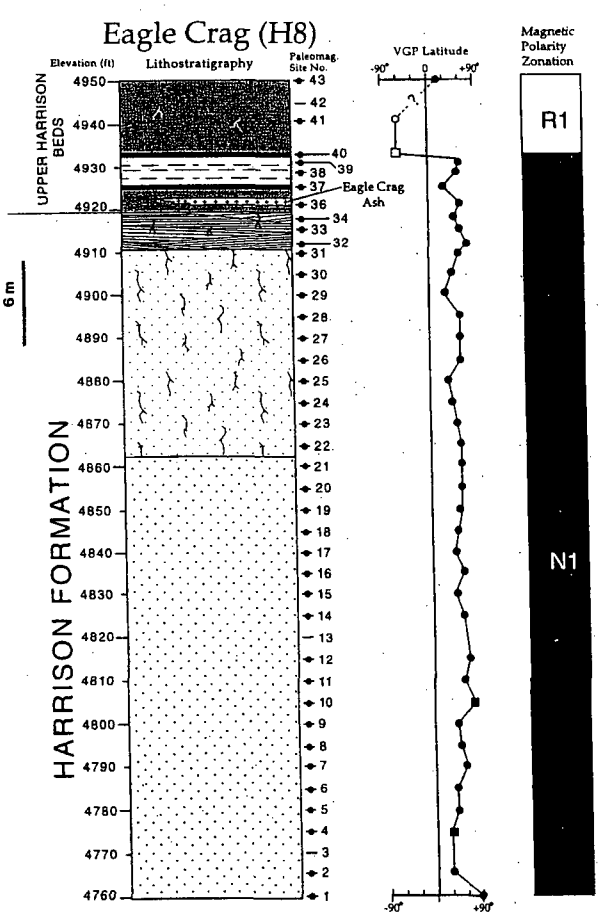


Figure 9. Litho- and magnetostratigraphy of the Eagle Crag section (H8). See Figure 11 for symbols.

it contains the contact between the highest stratigraphic unit (Upper Harrison beds) of the Arikaree Group and the lowest unit (Runningwater Formation) of the Hemingford Group. In addition, a local marker bed, the Agate Limestone, provides a tie-point for measured sections within the monument in which this bed occurs. The co-occurrence of the Agate Limestone in the Hoffman Channel, Carnegie Hill, and Skavdahl Dam sections (Fig. 15) establishes a reference horizon separating the pale brown volcaniclastic sandstone of the Upper Harrison beds below from the orange-brown volcaniclastic sandstone of the Runningwater Formation above.

In the Skavdahl Dam section, the lowermost zone ?R1 is defined by only a single site below a zone of indeterminate polarity. Two more reversed sites above the indeterminate zone are represented by zone R1 (Fig. 14). Zone R1 could either continue the same polarity from below, or represent a separate polarity event (neither of these possibilities can be ruled out given the existence of the indeterminate zone,

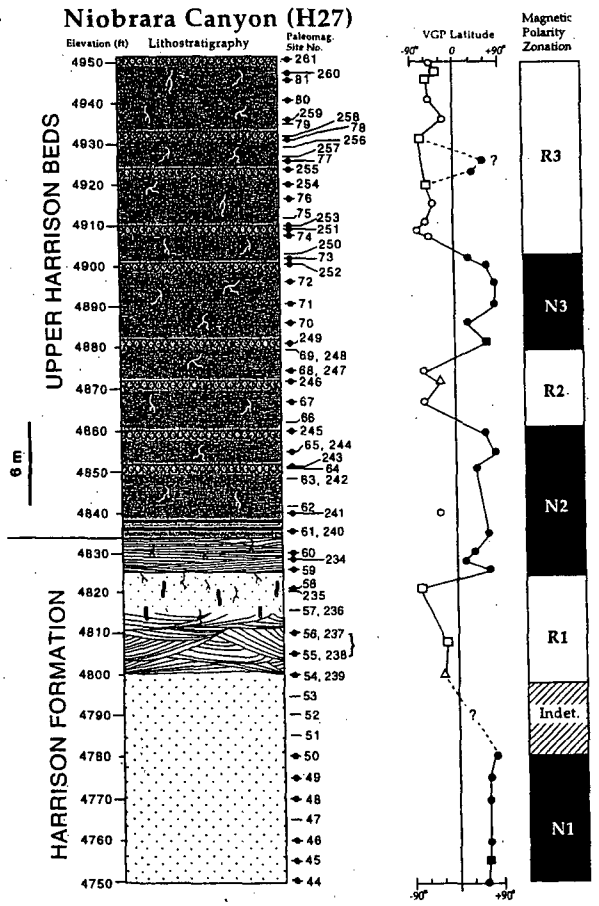


Figure 10. Litho- and magnetostratigraphy of the Niobrara Canyon section (H27). See Figure 11 for symbols.

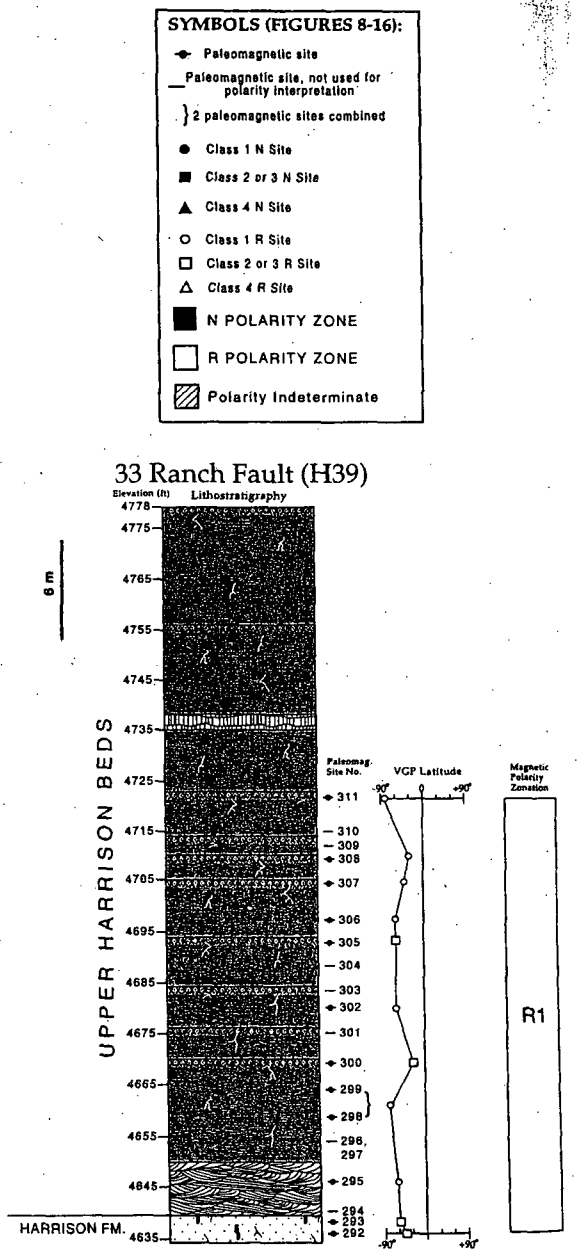


Figure 11. Litho- and magnetostratigraphy of the 33 Ranch section (H39).

although the former seems more likely). Although single sites (312 in this case) usually are not used for polarity interpretation, the reversed signature of zone ?R1 is considered valid in the Skavdahl Dam section because of its magnetostratigraphic correlation to the single, long reversed zone in the Upper Harrison beds of the Carnegie Hill and Hoffman Channel sections (Fig. 15). The two normal-polarity sites (314 and 318) are contained within a zone of poor magnetization (site 313 and sites 315–317 are uninterpretable) and given the polarity of adjacent sections, the normal sites are considered spurious overprints and hence indeterminate (Fig. 14). Above R1 a relatively long (between sites 321 and 326) and well-defined (6 Class 1 sites) magnetic polarity zone N1 characterizes more than half of the local outcrop of the Runningwater Formation, and this is overlain by a short zone R2 at the top of this unit.

Other sections (including H26, H28/29, H31, and H33). Other paleomagnetic samples, consisting of one or a few paleomagnetic sites, were collected to characterize the polarity of two important marker beds within the Agate area. The Agate Ash is highly variable in thickness along strike because it represents a water-lain tuff filling an irregular paleotopography. We collected paleomagnetic samples of this tuff at five separate geographic locations of the Agate Ash and five sites for the Agate Limestone. The paleomagnetic polarity data for the Agate Ash indicate a mixture of 2 N (Class 1) and 3 R (1 Class 1, 1 Class 3, 1 Class 4) sites. These unexpected data may result from postdepositional bioturbation (numerous invertebrate and vertebrate burrows) of the ash because it is unlikely that the thicker outcrops of the Agate Ash were deposited over a considerable period of time.

The five sites of the Agate Limestone yielded 1 N (Class 1) and 4 R (2 R3, 2 R4) polarities. Given their stratigraphic position in otherwise long, well-defined R zones (see below), we interpret the R data to indicate the polarity of the Agate Limestone (i.e., site 99 in the Hoffman Channel section is con-

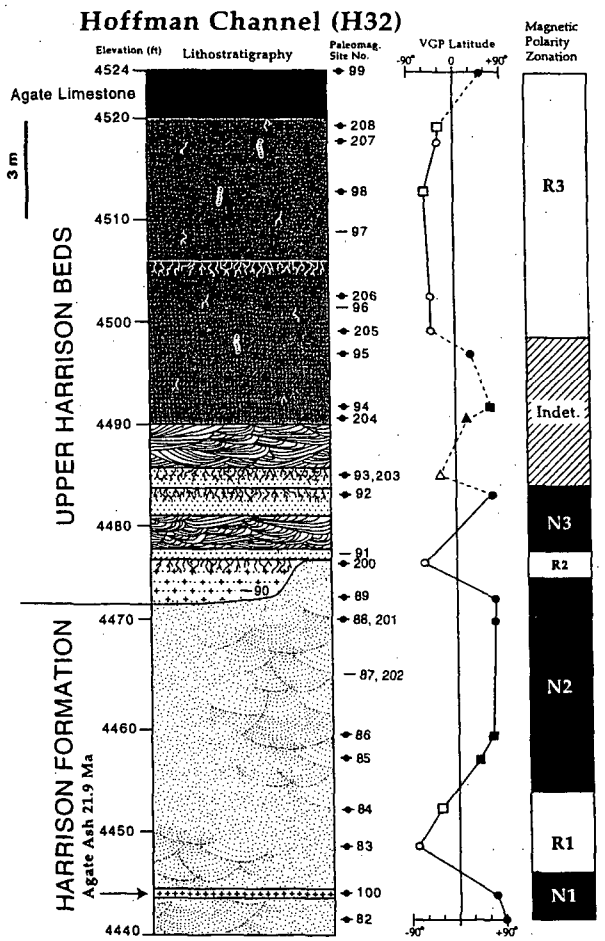


Figure 12. Litho- and magnetostratigraphy of the Hoffman Channel section (H32). See Figure 11 for symbols.

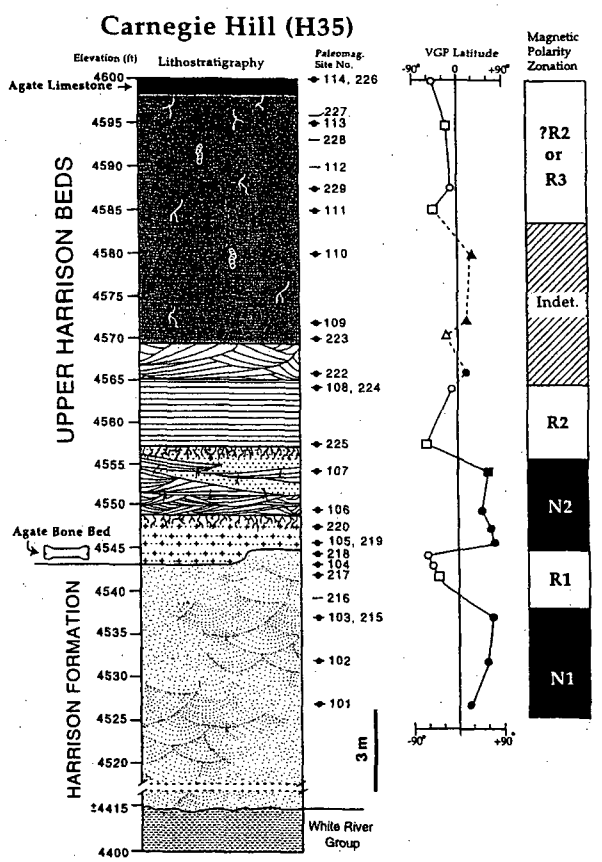


Figure 13. Litho- and magnetostratigraphy of the Carnegie Hill section (H35). See Figure 11 for symbols.

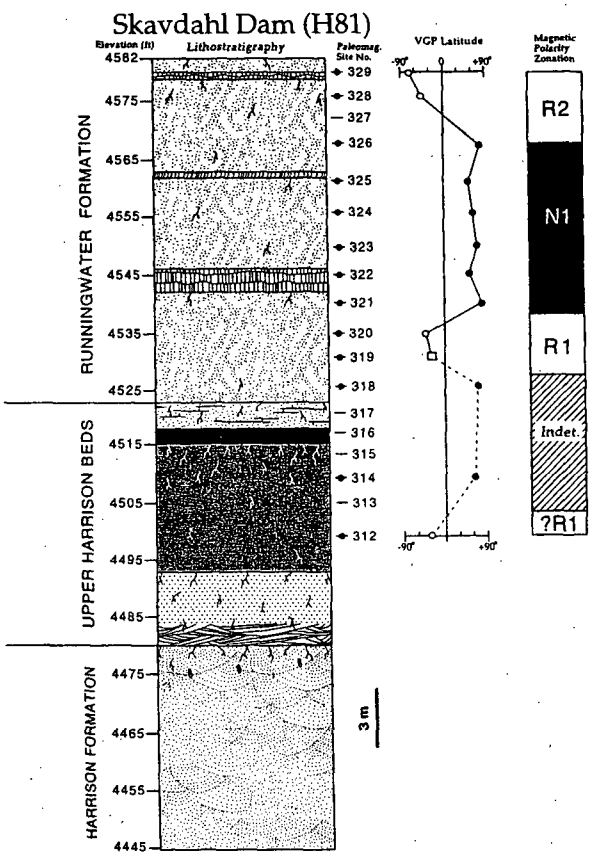


Figure 14. Litho- and magnetostratigraphy of the Skavdahl Dam section (H81). See Figure 11 for symbols.

sidered spurious). We also note the numerous burrows throughout this micrite limestone that may have influenced the paleomagnetic data.

Composite magnetic polarity stratigraphy and correlation to the Magnetic Polarity Time Scale

A composite section for the Arikaree Group in Sioux County, Nebraska, using the magnetic polarity data from the seven principal measured sections is presented (Fig. 15) and then correlated to the Magnetic Polarity Time Scale (MPTS) (Fig. 16) of Berggren et al. (1995). The lower part of the Arikaree Group, including the "Gering," Monroe Creek, and ?Harrison Formations as represented in the Pants Butte section, is characterized by a sequence of five magnetic polarity zones, designated R1-N1-R2-N2-R3 (Figs. 15, 16).

In their study of the lower Arikaree Group exposed at Wildcat Ridge in the southern part of the Nebraska panhandle, Tedford et al. (1996) use an R/N sequence calibrated with new, high-resolution $^{40}Ar/^{39}Ar$ age determinations to correlate the Gering/Monroe Creek formational transition to Chron C9 on the MPTS (Berggren et al., 1995). We interpret our R1/N1 transition in Sioux County to be the same interval as that at Wildcat Ridge and therefore correlate the base of our composite section to C9 in the late Oligocene part of the MPTS (Fig. 16). This correlation implies a local age range for the upper "Gering" through lower Monroe Creek of between ~28.2 and 27 m.y. ago. Above magnetic polarity zone N1 (which correlates to C9n at Wildcat Ridge) there are three polarity zones, R2, N2, and R3, locally representing what is lithologically the middle–upper Monroe Creek Formation and the overlying ?Harrison Formation. Given the age of earliest known Harrison faunas, which are approximated by an age of ~22 to 23 Ma (Tedford et al., 1987), the lack of diagnostic fossils in our Pants Butte section, and the numerous normal and reversed polarity zones in the MPTS in Chrons C8–C6AA between 27 and 22 Ma, the exact correlation of our three polarity zones (R2-N2-R3) (Fig. 16) in the Arikaree Group composite is uncertain.

Above R3 (Fig. 16) there seems to be a hiatus in our composite section for the following reason: the uppermost part of the Pants Butte section (Figs. 8, 15), which is lithologically suggestive of the lower part of the ?Harrison Formation, is of reversed polarity. In contrast, the lowermost parts of each of the other sections that begin in the Harrison Formation are of normal polarity (Eagle Crag and Niobrara Canyon) (Figs. 9, 10, 15), or mixed N-R polarity (Hoffman Channel and Carnegie Hill) (Figs. 12, 13, 15). The exact duration of this hiatus is uncertain and could span several hundred thousand, to a few million, years within the interval between 27 and 22 m.y.

In our study area the most complete and best-exposed section of the Harrison Formation occurs near the head of Monroe Canyon at Eagle Crag where Hatcher (1902) initially described the formation. Here the type Harrison Formation consists of a 48-m-thick (160 ft) zone of fine-grained gray sandstones showing normal polarity. The formation ends in a prominent regional paleosol, also of normal polarity, that can be traced southward in other measured sections (Fig. 15).

In the southern and eastern areas of the Agate National Monument, a broad, shallow Upper Harrison stream valley cuts into the Harrison Formation, removing the terminal paleosol and the upper part of the formation. This downcutting would seem to explain the absence of the long, normal polarity interval (N1 of the Eagle Crag section) in the abbreviated Carnegie Hill (H35) and Hoffman Channel (H32) sections. In fact, the appearance of reversed intervals in these latter sections may result from downcutting of the formation to stratigraphic levels low enough to record reversed magnetozones. Consequently, the N1-R1-N2-R2 zones of the Hoffman Channel section (designated as the thin magnetic polarity zones N3, R4, N4, and R5) (Fig. 16) are placed below the long normal zone (designated N5) (Fig. 16) of the Eagle Crag section. It is possible that the reversed interval (upper R3, between 4,793 and 4,813 ft) at the top of the Pants Butte section correlates with one of the reversed zones in the Harrison Formation at Hoffman Channel and Carnegie Hill.

Using this sequence in the middle to upper part of the

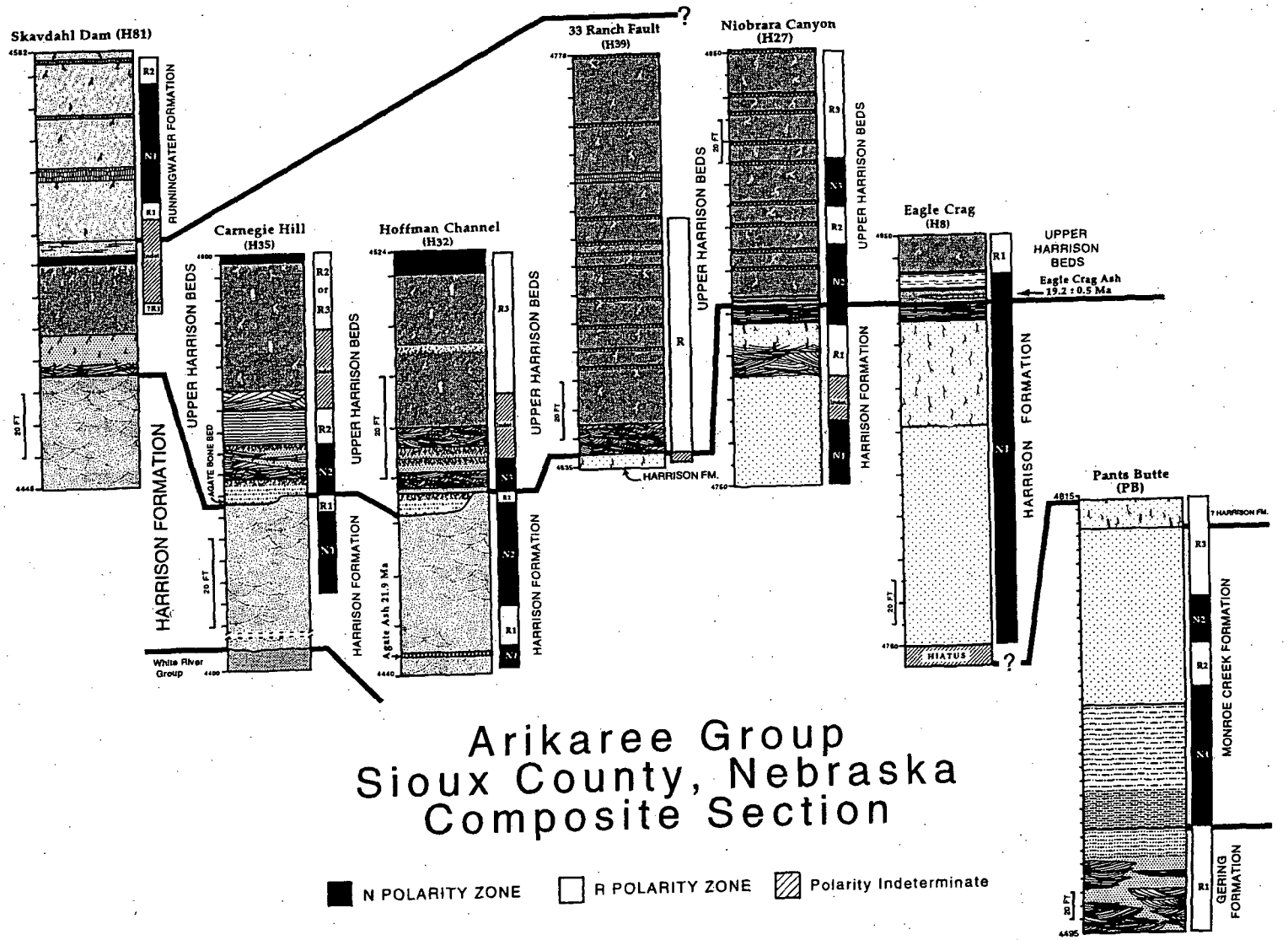


Figure 15. Composite correlation of the litho- and magnetostratigraphy of the Arikaree Group and lower Runningwater Formation in Sioux County, Nebraska. See Figure 11 for symbols.

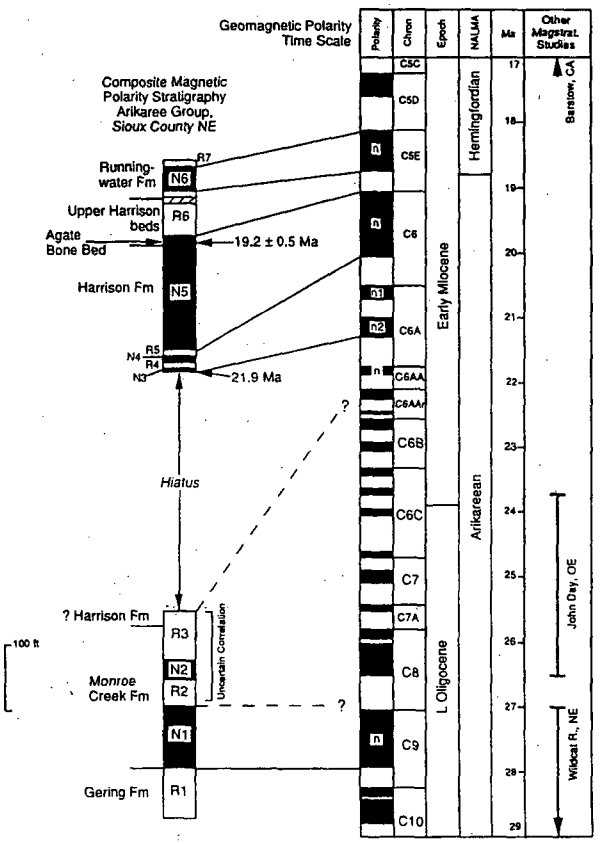


Figure 16. Correlation of the Arikaree Group and lower Runningwater Formation to the Magnetic Polarity Time Scale (Berggren et al., 1995).

Arikaree Group composite and the two available radioisotopic age determinations, the sequence of local magnetic polarity zones N3–N5 (Fig. 16) possibly correlates to the upper half of Chron C6A and the lower third of Chron C6. Given the uncertainty (i.e., lack of reported analytical error) in the age of the Agate Ash, a correlation to Chron C6AAn and the greater part of C6A also appears plausible. In the Hoffman Channel section, the N1-R2 zones (N3-R5 zones) (Fig. 16) span only ~10.7 m (35 ft), and either correlation proposed here would involve about the same amount of time for the deposition of these sediments, 1.15–1.20 m.y. of the MPTS.

The prominent local magnetic polarity zone N5 (Fig. 16) unambiguously correlates to relatively long duration and distinctive Chron C6n on the MPTS. The 19.2 ± 0.5 Ma fission-track (zircon) age determination at the top of local magnetic polarity zone N5 is virtually identical to the extrapolated age for the N5–R6 boundary that correlates to the C6n to C5Er boundary at 19.1 Ma on the MPTS. The six sections (i.e., Eagle Crag, Niobrara Canyon, 33 Ranch Fault, Hoffman Channel, Carnegie Hill, and Skavdahl Dam) that contain the Upper Harrison beds and/or the important lithostratigraphic marker Agate Limestone are dominantly of reversed polarity in their upper parts (except for zone N3 in the Niobrara Canyon section) (Figs. 10, 15). Zone R6 within the Upper Harrison beds is correlated to Chron C5E between 19.1 and 18.75 m.y. on the MPTS. This correlation requires us to interpret N3 in the Niobrara Canyon section (Fig. 10) as a brief normal episode apparently missing in the MPTS. Zone N3 of the Niobrara Canyon section is absent in other Upper Harrison sections in the study; we interpret it as a valid normal-polarity interval probably represented by nondeposition elsewhere in the formation. An alternate interpretation would suggest that zone N3 should correspond to Chron C5En, a span of ~500,000 yr; the Upper Harrison beds would then include the uppermost part of Chron C6n, all of C5E, and all or part of C5Dr, and would extend in time from 19.2 Ma to as young as about 17.7 Ma. However, based on associated faunal and radioisotopic data, we consider the former correlation most probable and the latter unlikely.

As represented in the Skavdahl Dam section (H81), the lower part of the Runningwater Formation (the only part of the formation present at this locality) is of mostly normal-polarity (local zone N1 in Fig. 14, composite zone N6 in Fig 16). Given its position in the composite litho- and magnetic polarity stratigraphy, this zone plausibly correlates to Chron C5En on the MPTS between 18.75 and 18.3 m.y. The uppermost magnetic polarity zone in our composite section is near the top of the local Runningwater section and is designated R2 in the Skavdahl Dam section (Figs. 14, 15) and R7 in our Sioux County composite (Fig. 16). This reversed magnetic polarity zone then correlates to Chron C5Dr on the MPTS and indicates that this part of the section in Sioux County is younger than 18.3 Ma and older than 17.6 m.y. (i.e., the latter represents the C5Dr–C5Dn transition, which is not encountered in our section). However, if the alternate interpretation noted above is accepted, zone N1 of the Skavdahl Dam section (N6) (Fig. 16) would correspond to Chron C5Dn between 17.25 and 17.6 Ma, suggesting that the Runningwater Formation in the Skavdahl Dam section most likely falls between ~17 and 17.7 Ma. Here we consider this latter alternate interpretation improbable.

Although there are several other possibilities, we prefer the correlation to the MPTS indicated in Figure 16 in which the Upper Harrison beds and lower Runningwater Formation correspond for the most part to Chron C5E. This is based on faunal correspondence of mammals from the Runningwater Formation (Skavdahl section and laterally equivalent rocks) to age-equivalent mammals of the Hackberry Local Fauna, eastern Mojave province, California, which occur in superposition with radiometrically dated volcanics (Reynolds et al., 1995). Lacustrine sediments containing the Hackberry mammals are overlain by volcanics of the Woods Mountains volcanic center dated by the $^{40}Ar/^{39}Ar$ method at 17.6 to 18.5 Ma (McCurry et al., 1995). The Hackberry Local Fauna also lies directly beneath precaldera volcanic rocks (Hackberry Spring Volcanics of McCurry) at Hackberry Mountain that were erupted from 17.8

to ~18.5 Ma. This indicates that the correlative fauna of the lower part of the Runningwater Formation in our study area northeast of Agate post office can be dated to ~18.5 Ma or somewhat older. Because the Runningwater sediments northeast of Agate are predominantly of normal polarity, mammals dating to ~18.5 Ma would fall in Chron C5En between 18.75 and 18.3 m.y. Therefore we conclude that the Runningwater rocks northeast of Agate most likely date from ~18.8 to ~18 Ma.

DISCUSSION AND SIGNIFICANCE

Correlations to other middle Tertiary North American magnetostratigraphic studies containing land mammals

A global review of all published Cenozoic magnetostratigraphic studies containing land-mammals (Opdyke, 1990; Opdyke and Channell, 1996) indicates that, in contrast to results available for the later Miocene through Pleistocene, the late Oligocene and early Miocene interval between 30 and 20 m.y. is poorly characterized for its polarity signature. The new data for the Arikaree Group presented here therefore fill a major gap in middle Tertiary magnetochronology.

Three other early Miocene magnetostratigraphic studies of North American land-mammal faunas are relevant here (Fig. 16). Tedford et al. (1996) presented paleomagnetic and radioisotopic data for the Whitneyan/Arikareean boundary as it is represented in the upper White River Group and lower Arikaree Group at Wildcat Ridge in Morrill County, Nebraska. With regard to the upper part of their sequence, high-resolution $^{40}Ar/^{39}Ar$ age determinations and the match of their R-N transition with our local section in Sioux County allow calibration of the lower part of the Arikareean within Chron C9 between 28.2 and 27 m.y.

Prothero and Rensberger (1985) presented the magnetostratigraphy of the Turtle Cove and Kimberly members of the John Day Formation of Oregon. This sequence is of fundamental importance to middle Tertiary land-mammal chronology because it represents a relatively long sequence spanning the late Oligocene–early Miocene and it contains some classic faunas. The magnetic polarity zonation of the John Day, which contains important Arikareean mammals from the Columbia River Plateau region, correlates from late Chron C8r to early Chron C6r on the MPTS. Taking these chrons and using the timescale of Berggren et al. (1995), this indicates a duration between 27 and 24.5 m.y. (Note that the original study of Prothero and Rensberger [1985] had a lower boundary for this sequence at 28.5 Ma, i.e., about 1 m.y. older than using the revised correlations). The John Day beds analyzed by Prothero and Rensberger (1985) therefore occur within part of the temporal span of the hiatus represented by our work in Sioux County (Fig. 16).

MacFadden et al. (1990) presented the magnetostratigraphy of the Barstow sequence from the Mojave Desert in southern California. Known for diagnostic land-mammals since the early part of this century, the Barstow sequence is of importance because it contains the typical Barstovian (middle Miocene) fauna (Wood et al., 1941; Tedford et al., 1987). Late Hemingfordian land mammals occur in the lower Barstow Formation. These are followed in direct superposition above by typical Barstovian mammals in the middle and upper parts of the Barstow Formation. Using radioisotopic and paleomagnetic data, the Hemingfordian faunas from the lower Barstow Formation correlate to Chron C5C between 17.3 and 16 m.y. on the timescale of Berggren et al. (1995). These correlations indicate that the late Hemingfordian mammals of the lower Barstow Formation are about 1 to 2 m.y. younger than those of the earliest Hemingfordian as represented in the lower Runningwater Formation (about 18.8–18 Ma) in Sioux County.

Arikareean-Hemingfordian boundary

Prior to the present study, the boundary between the Arikareean and Hemingfordian land-mammal ages had been somewhat uncertain. In their detailed correlation chart, Tedford et al. (1987, Fig. 6.2) tentatively indicate this boundary as a dashed line at 20 Ma. The sequence of sediments, land-mammals, and polarity signatures in Sioux County bear directly on the precise age of this boundary. Latest Arikareean mammals occur in all but the uppermost part of a reversed polarity zone (R6 in Fig. 16) in the Upper Harrison beds; earliest Hemingfordian mammals (northeast of Agate Local Fauna; 18.8–18 Ma) occur in the succeeding normal polarity zone (N6) and in the lowest part of the overlying reversed zone (R7 in Fig. 16), both within the lower Runningwater Formation. Zone R6 correlates to the Chron C5Er on the MPTS, which has lower and upper age boundaries of 19.05 and 18.75 Ma (Berggren and et al., 1995). Given the transition between Arikareean to Hemingfordian faunas within the latest part of Chron C6r, an extrapolated age for this boundary is 18.8 Ma, slightly younger than indicated by Tedford et al. (1987).

SUMMARY AND CONCLUSIONS

The late Oligocene to early Miocene was an important interval in North American land-mammal evolution because it includes the transition from faunas of archaic to those of more modern aspect. This time also records the dispersal of numerous biochronologically important immigrant taxa from the Old World. As indicated in the Miocene correlation chart of Tedford et al. (1987, Fig. 6.2), this interval is characterized by a paucity of faunas contained within long stratigraphic sequences amenable to precise paleomagnetic and radioisotopic dating methods. The Arikaree and Hemingford Groups of Sioux County consist of relatively long stratigraphic sequences that include the typifying land-mammal faunas for the Arikareean and Hemingfordian ages. Furthermore, this stratigraphic sequence preserves a well-defined boundary between Arikaree and superposed Hemingford sediments (the correspondence of the lithostratigraphic boundary between the rock units and the biochronologic boundary between

the Arikareean and Hemingfordian land mammal ages in this case is coincidental). As such, this study represents a useful addition to our overall knowledge of middle Tertiary mammalian biochronology for North America.

As described above, the paleomagnetic signatures of the various measured sections in our study vary from excellent to marginal quality. The most probable reason for the poor quality data is the presence of intervals of sediments that have been heavily bioturbated (i.e., burrowed) and are characterized by paleosols. The other, more pristine, parts of the Arikaree Group give excellent paleomagnetic results.

The ability to correlate a local magnetic polarity zonation to the MPTS requires either a unique pattern with one or more distinctive polarity intervals, excellent biochronologic data, and/or interbedded high-resolution radioisotopic age determinations. Between 30 and 20 m.y. ago the geomagnetic history of the earth's main field is relatively "busy" with numerous polarity transitions during this interval. We are fortunate in the current study of the Arikaree Group because Chron C6n, which has the longest duration of a single polarity (1.05 Ma using Berggren et al., 1995) within this interval on the MPTS, is a distinctive polarity signature for much of the Harrison Formation and the lowermost part of the Upper Harrison beds. In conjunction with a fission-track age determination of 19.2 ± 0.5 Ma for the Eagle Crag Ash, the correlation of this part of the Arikaree Group to the MPTS is solid and unambiguous.

ACKNOWLEDGMENTS

We thank J. Daniel Bryant and Josh Kaufman for help in collecting samples in the field and Robb Browning, J. Daniel Bryant, and Vicky Norton for assistance in the University of Florida Paleomagnetic Laboratory. We also thank the ranchers of Sioux County and the National Park Service staff at Agate Fossil Beds National Monument for access to outcrops on their land and their support of this project. This is University of Florida Contribution to Paleobiology No. 457.

APPENDIX A: LOCATION OF MEASURED SECTIONS IN SIOUX COUNTY, NEBRASKA, PROVIDING PALEOMAGNETIC SAMPLES

Pants Butte: The section begins in S2, NW4, NE4, SW4, sec. 17, T. 32 N, R. 55 W., and continues through SW4, NE4, SW4, sec. 17, into NE4, SW4, SW4, sec. 17, and through W2, SE4, SW4, SW4, sec. 17, ending in W2, NE4, NW4, NW4, sec. 20, T. 32 N., R. 55. W.

H8: The section begins in center, SE4, SW4, NE4, sec. 26, T. 32 N., R. 56 W., and continues southwest along drainage in NW4, NW4, SW4, sec. 26, to SE4, NE4, SE4, sec. 27, ending in NW4, SE4, SE4, sec. 27, T. 32 N., R. 56 W.

H27: The section begins in SE4, sec. 28. T. 31 N., R. 57 W., and continues into NE4, sec. 33, T. 31 N., R. 57 W.

H32: The section is at center, W2, SE4, sec. 4, T. 28 N., R. 55 W.

H33: The section is at W2, SE4, SW4, sec. 4, T. 28 N., R. 55 W.

H35: The section is at NW4, NW4, NW4, sec. 10, T. 28 N., R. 55 W.

H39: The section is at W2, NE4, sec. 23, T. 30 N., R. 57 W.

H81: The section begins at center, NE4, NW4, SE4, sec. 35, and continues to NE4, NE4, NE4, SE4, sec. 35, T. 29 N, R. 55 W.

REFERENCES CITED

Adams, G. I., 1902, Geology and water resources of the Patrick and Goshen Hole Quadrangles in eastern Wyoming and western Nebraska: U.S. Geological Survey Water Supply and Irrigation Paper 70, p. 1–41.

Berggren, W. A., Kent, D. V., Swisher, C. C. , and Aubry, M. P., 1995, A revised Cenozoic geochronology and chronostratigraphy, *in* Berggren, W. A., Kent, D. V., Aubry, M. P., and Hardenbol, J., eds., Geochronology, time scales and global stratigraphic correlation: Tulsa, SEPM Special Publication 54, p. 129–212.

Butler, R. F., 1992, Paleomagnetism: magnetic domains to geologic terranes: Boston, Blackwell, p. 1–319.

Dalrymple, B. G., 1979, Critical tables for conversion of K-Ar ages from old to new constants: Geology, v. 7, p. 558–560.

Darton, N. H., 1899, Preliminary report on the geology and water resources of Nebraska west of the 103rd meridian: U.S. Geological Survey 19th Annual Report 1897–98; pt. 4, Hydrography: Washington, D. C., U.S. Geological Survey, p. 719–785.

Darton, N. H., 1905, Preliminary report on the geology and underground water resources of the central Great Plains: U.S. Geological Survey Professional Paper, v. 32, p. 1–433.

Evernden, J. F., Savage, D. E., Curtis, G. H., and James, G. T., 1964, Potassium-argon dates and the Cenozoic mammalian chronology of North America: American Journal of Science, v. 262, p. 145–198.

Fisher, R. A., 1953, Dispersion on a sphere: Royal Society of London Proceedings, v. 217, p. 295–305.

Hatcher, J. B., 1902, Origin of the Oligocene and Miocene deposits of the Great Plains: Proceedings of the American Philosophical Society, v. 41, p. 113–131.

Hunt, R. M., Jr., 1978, Depositional setting of a Miocene mammal assemblage, Sioux County, Nebraska: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 24, p. 1–52.

Hunt, R. M., Jr., 1985, Faunal succession, lithofacies, and depositional environments in Arikaree rocks (lower Miocene) of the Hartville Table, Nebraska and Wyoming, *in* Martin, J. E., ed., Fossiliferous Cenozoic deposits of western South Dakota and northwestern Nebraska: Dakoterra, v. 2, no. 2, p. 155–204.

Hunt, R. M., Jr., 1990, Taphonomy and sedimentology of Arikaree (lower Miocene) fluvial, eolian, and lacustrine paleoenvironments, Nebraska and Wyoming: A paleobiota entombed in fine-grained volcaniclastic rocks: Geological Society of America Special Paper 244, p. 69–111.

Hunt, R. M., Jr., Xue, X. X., and Kaufman, J., 1983, Miocene burrows of extinct beardogs: Indication of early denning behavior of large mammalian carnivores: Science, v. 221, p. 364–366.

Johnson, N. M., and McGee, V. E., 1983, Magnetic polarity stratigraphy: stochastic properties of data, sampling sets and the evaluation of interpretations: Journal of Geophysical Research, v. 88, p. 1213–1221.

Kirschvink, J. L., 1980, The least-squares line and plane and the analysis of paleomagnetic data: Royal Astronomical Society Geophysical Journal, v. 62, p. 699–718.

MacFadden, B. J., Swisher, C. C., III, Opdyke, N. D., and Woodburne, M. O., 1990, Paleomagnetism, geochronology, and possible tectonic rotation of the middle Miocene Barstow Formation, Mojave Desert, southern California: Geological Society of America Bulletin, v. 102, p. 478–493.

Matthew, W.D., 1923, Fossil bones in the rock: Natural History, v. 23, no. 4, p. 358–369.

McCurry, M., Lux, D.R., and Mickus, K.L., 1995, Neogene structural evolution of the

Woods Mountains Volcanic Center, East Mojave National Scenic Area: San Bernardino County Museum Association Quarterly, v. 42, no. 3, p. 75–80.

Opdyke, N. D., 1990, Magnetic stratigraphy of Cenozoic terrestrial sediments and mammalian dispersal: Journal of Geology, v. 98, p. 621–637.

Opdyke, N. D., and Channell, J. E. T., 1996, Magnetic stratigraphy: San Diego, California, Academic Press, 346 p.

Opdyke, N. D., Lindsay, E. H., Johnson, N. M., and Downs, T., 1977, The paleomagnetism and magnetic polarity stratigraphy of the mammal-bearing section of Anza-Borrego State Park, California: Quaternary Research, v. 7, p. 316–329.

Peterson, O. A., 1907, The Miocene beds of western Nebraska and eastern Wyoming and their vertebrate faunae: Carnegie Museum Annals, v. 4, no. 1, p. 21–72.

Peterson, O. A., 1909, A revision of the Entelodontidae: Carnegie Museum Memoir 4, no. 3, p. 41–158.

Prothero, D. R. and Rensberger, J. M., 1985, Preliminary magnetostratigraphy of the John Day Formation, Oregon, and the North American Oligocene-Miocene boundary: Newsletters on Stratigraphy, v. 15, p. 59–70.

Reynolds, R. E., Hunt, R. M., Jr., and Albright, B., 1995, Rhinoceros in Lanfair Valley: San Bernardino County Museum Association Quarterly, v. 42, no. 3, p. 107–110.

Scott, W. B., 1893, The later Tertiary lacustrine formations of the west: Geological Society of America Bulletin, v. 5, p. 594–595.

Stanley, K. O.., and Benson, L. V., 1979, Early diagenesis of High Plains Tertiary vitric and arkosic sandstone, Wyoming and Nebraska: Tulsa, SEPM Special Publication 26, p. 410–423.

Tedford, R. H., Swinehart, J., Hunt, R. M., and Voorhies, M. R. 1985, Uppermost White River and Lowermost Arikaree rocks and faunas, White River Valley, northwestern Nebraska, and their correlation with South Dakota: Dakoterra, v. 2, no. 2, p. 334–352.

Tedford, R. H. (chairman), and 8 others, 1987, Faunal succession and biochronology of the Arikareean through Hemphillian interval (late Oligocene through earliest Pliocene epochs) in North America, *in* Woodburne, M. O., ed., Cenozoic mammals of North America: Berkeley, University of California Press, p. 153–210.

Tedford, R. H., Swinehart, J., Swisher, C., Prothero, D., King, S., and Tierney, T., 1996, The Whitneyan-Arikareean Transition in the High Plains, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: Cambridge, U. K., Cambridge University Press, p. 312–334.

Wood, H. E., II (chairman), Chaney, R. W., Clark, J., Colbert, E. H., Jepsen, G. L., Reeside, J. B., Jr., and Stock, C., 1941, Nomenclature and correlation of the North American continental Tertiary: Geological Society of America Bulletin, v. 52, p. 1–48.

MANUSCRIPT ACCEPTED BY THE SOCIETY DECEMBER 5, 1997

Printed in U.S.A.

Geological Society of America
Special Paper 325
1998

The Arikareean Land Mammal Age in Texas and Florida: Southern extension of Great Plains faunas and Gulf Coastal Plain endemism

L. Barry Albright III
Florida Museum of Natural History, University of Florida, Gainesville, Florida 32611

ABSTRACT

It has long been recognized that early Miocene faunas of the Gulf Coastal Plain represented a distinct biogeographic province from those better known assemblages of similar age in the northern Great Plains. Yet the age of Gulf Coast faunas was based on the general resemblance, particularly at the generic level, to faunas from the geochronologically better developed and more fossiliferous midcontinent stratigraphic sequences. Thus, whereas the early Miocene environments of the Gulf Coastal Plain and the Great Plains were certainly different, barriers or filters to dispersal between the two regions were only weakly imposed. A significant new fauna from the Texas Coastal Plain, the Toledo Bend Local Fauna, firmly establishes the presence of Arikareean land mammals in the region and provides information that aids in constraining the timing of midcontinent–Gulf Coastal Plain biogeographic disparity. On the one hand the mammalian taxa from Toledo Bend provide evidence for a filtered corridor of dispersal between the rich faunas of the northern Great Plains and the Gulf Coastal Plain, but on the other hand they support recognition of earliest Miocene Gulf Coastal Plain endemism. Taxa shared by both regions include the anthracothere *Arretotherium,* a small protoceratid similar to *Protoceras,* the tapir *"Miotapirus" marslandensis,* a large entelodont, a hypertragulid, and the rhinoceros *Diceratherium.* Gulf Coastal Plain endemic taxa, on the other hand, include the dwarf amphicyonid *Daphoenodon notionastes,* a dwarf chalicothere, a probable new species of dwarf rhinoceros, a small floridatragulid camelid, and the rodent *Texomys.* These taxa additionally provide faunal continuity with Arikareean faunas in Florida, which are also reviewed. Development, differentiation, and decline of a Gulf Coastal Plain early Miocene fauna has also been discussed in terms of a northward and eastward expansion of the Neotropical realm. The Toledo Bend Local Fauna also aids in timing the early phase of this event.

INTRODUCTION

The Arikareean North American Land Mammal "Age" (NALMA) is poorly represented in the Texas Coastal Plain and in the Gulf Coastal Plain in general. In contrast, the next younger interval, the Hemingfordian, is much better represented primarily because of the well-known Garvin Gully Fauna of Texas and the Thomas Farm Local Fauna of Florida; both were originally considered Arikareean in age (see Patton, 1969a, p. 205–210, for a discussion). In Florida there are eight published localities with mammals indicative of an Arikareean age. In Texas there are at least two local faunas plus a few other minor, single taxon, sites. Summaries of most of these sites can be found in MacFadden and Webb (1982), Tedford and Hunter (1984), Tedford et al. (1987), and most recently Morgan (1989, 1993, 1994). Another site in Escambia County, Alabama, where Westgate (1992) reported

Albright, L. B., III, 1998, The Arikareean Land Mammal Age in Texas and Florida: Southern extension of Great Plains faunas and Gulf Coastal Plain endemism, *in* Terry, D. O., Jr., LaGarry, H. E., and Hunt, R. M., Jr., eds., Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America): Boulder, Colorado, Geological Society of America Special Paper 325.

three teeth of a large entelodont (*Dinohyus*) from the "Catahoula Sandstone," may also be Arikareean in age. The Catahoula Formation of Texas, Louisiana, Mississippi, and Alabama is variably considered latest Oligocene to earliest Miocene in age.

Tedford and Hunter (1984) and Morgan (1993, 1994) addressed Florida mammal faunas recovered from nearshore marine depositional settings and their importance with regard to mammalian/invertebrate biochronologic correlation. Faunas found in wholly terrestrial settings, on the other hand, particularly those collected from karst infillings, were correlated to their respective land mammal age on the basis of characterizing taxa and, in some cases, stage of evolutionary development. Localized infilling of the karst fissures containing these faunas precludes attempts at lithostratigraphic correlation between widely distributed sites, and even between those within the same quarry.

In Texas, stratigraphic control is somewhat better where a long belt of fluvial sediments that parallels the present Texas coastline from near the Mexican border to the Sabine River records a relatively "continuous" sequence of early to late Miocene land mammal faunas. Nevertheless, refined biochronologic control is lacking for earliest Miocene mammal sites in this region because associations of mammal fossils with marine invertebrates have not yet been discovered to the extent that they have in Florida. Because sites in Texas are rare, most of our knowledge about the Arikareean in the Gulf Coastal Plain is based on Florida assemblages. As is discussed in the following section, however, a significant new vertebrate assemblage from eastern Texas, the Toledo Bend Local Fauna (Albright, 1990, 1991, 1994, 1996; Manning, 1990), adds considerably to our understanding of the Arikareean Land Mammal Age in the southeastern United States.

Early Miocene mammals from the Texas Coastal Plain have long been recognized from a series of localities that occur at various stratigraphic levels within the laterally equivalent (in part) Oakville and Fleming Formations. Only recently has a mammal fossil (entelodont) been reported from the underlying Catahoula Formation (Westgate, 1993). Although Hay (1924) described several mammalian taxa from this sequence, it was Hesse (1943) who first placed the various localities and faunas known at that time into a biostratigraphic context. Wilson (1956, p. 2233; 1957, 1959, 1960) refined the mammalian biostratigraphy of the Texas Coastal Plain and noted that, although the northern Great Plains were so "spectacularly lucrative" paleontologically, the Texas Coastal Plain, with its "fragmentary and poorly preserved" fossils was nonetheless "extremely important stratigraphically. . . ." To complicate matters, however, the base of the Fleming Formation, and therefore the outcrop pattern of the Catahoula Formation, is mapped differently in eastern Texas than in western Louisiana resulting in different interpretations as to whether certain faunas, particularly the Toledo Bend Local Fauna, actually occur in the lower Fleming Formation or the upper Catahoula Formation (see below). The heterogeneity of the Fleming Formation in this region (the Toledo Bend Local Fauna is found within 400 m of the state border) is such that accurate correlation, and therefore age determination, based on lithostratigraphy is impractical. Therefore, even though the early Miocene stratigraphy of this area is more straightforward and understood to a greater extent than that in Florida, the age of the Toledo Bend Local Fauna is still based primarily on characterizing taxa. This does not, however, detract from the significance of the fauna: the Toledo Bend Local Fauna is the most diverse Arikareean assemblage yet discovered in the Gulf Coastal Plain; it extends the geographic range of certain Florida taxa of similar age farther west, thereby defining the geographic boundaries of Arikareean Gulf Coastal Plain endemism; and many of its mammalian taxa were previously known only from the Great Plains. The shared Great Plains/Coastal Plain mammals are particularly germane to this report because they contribute evidence about the filtered affiliation between Great Plains and Gulf Coastal faunas and about Gulf Coastal Plain endemism.

Whether the Arikareean land mammal age was represented in the Texas Coastal Plain prior to discovery of the Toledo Bend Local Fauna has historically been a point of controversy. The well-known and important assemblages from the Oakville and lower Fleming Formations in Grimes, Washington, Walker, and San Jacinto Counties, collectively referred to as the Garvin Gully Fauna, were originally considered late Arikareean (Quinn, 1952, 1955; Wilson, 1956, 1957, 1959, 1960). More recent work, however, (Patton, 1969a; Forsten, 1975a,b; Tedford et al., 1987) indicates that the Garvin Gully Fauna is early Hemingfordian in age, correlative with faunas of the Runningwater Formation, Nebraska, and that only a very few assemblages placed within it, such as the Cedar Run Local Fauna of Washington County (Wood and Wood, 1937) and, perhaps, the Aiken Hill Local Fauna of Walker County, are late Arikareean. The latter two are small, rather poor assemblages and in this report are not considered correlative. Evidence presented in this chapter suggests that the Cedar Run Local Fauna is equivalent in age to the Toledo Bend Local Fauna, i.e., "medial" Arikareean, whereas the Aiken Hill locality is here considered latest Arikareean or earliest Hemingfordian. The informal term medial Arikareean is used in this report for those faunas containing taxa of both early and late Arikareean affinity. The designations of early early (Ar1), late early (Ar2), early late (Ar3), and late late (Ar4) follow the division of the Arikareean LMA proposed by Tedford et al. (1987). In this context, the Toledo Bend Local Fauna is considered early late Arikareean (Ar3).

Figure 1 shows all published Arikareean localities in the Gulf Coastal Plain. The early Hemingfordian Thomas Farm and Garvin Gully faunas are included because they are often mentioned in this chapter. Although references cited above provide summaries of Gulf Coastal Plain Arikareean faunas, no single publication reviews them all. Therefore, the first part of this chapter serves as an updated summary of all published faunas in the Gulf Coastal Plain of appropriate age. Following these summaries, problems particular to the region with respect to correlation and age assignment are discussed. Finally, a section on the significance of the Toledo Bend Local Fauna and implications

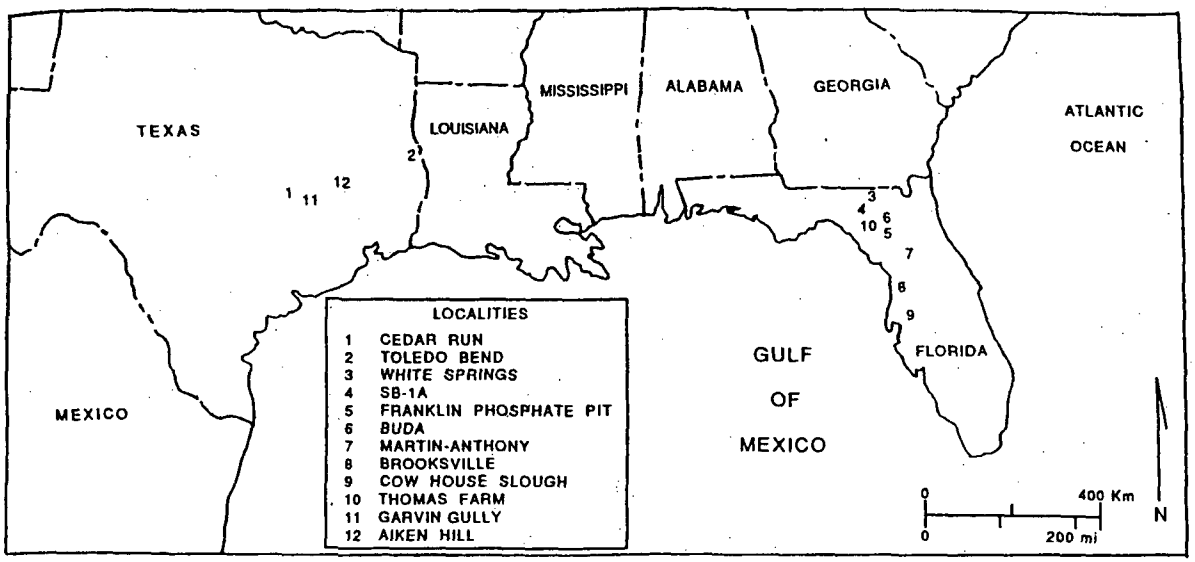


Figure 1. Location of major Arikareean localities in the Gulf Coastal Plain, plus the early Hemingfordian Garvin Gully and Thomas Farm localities.

regarding the relationship between the Gulf Coastal Plain and the northern Great Plains biogeographic province is presented.

ARIKAREEAN LAND MAMMAL AGE IN FLORIDA

There are eight published sites in Florida with mammals representative of the Arikareean age: the Martin-Anthony oreodont site and the Martin-Anthony Local Fauna in Marion County (MacFadden, 1980); the Brooksville Local Fauna of Hernando County (Patton, 1967); the Franklin Phosphate Pit No. 2 in Alachua County (Simpson, 1930); the SB-1A Local Fauna of Suwannee County (Frailey, 1978); the Buda Local Fauna of Alachua County (Frailey, 1979); the White Springs Local Fauna of Columbia and Hamilton Counties (Morgan, 1989, 1993, 1994); and the Cow House Slough Local Fauna from Hillsborough County (Morgan, 1989, 1993). Table 1 lists the taxa from each. A ninth, very rich site producing the Brooksville 2 Local Fauna is currently under study by G. Hayes and colleagues at the Florida Museum of Natural History. This fauna will be mentioned only to a limited extent pending publication by Hayes. The Brooksville 2 Local Fauna surpasses the Buda Local Fauna in mammalian diversity and will contribute substantially to an understanding of the Arikareean LMA in the Gulf Coastal Plain.

Cow House Slough Local Fauna. The Cow House Slough Local Fauna may be the oldest Arikareean assemblage in Florida. Morgan (1989, 1993) suggested a late early Arikareean age based on the presence of *Palaeolagus* and *Miohippus*. More recent study of the lagomorph by G. Hayes (personal communication, 1997) indicates that it most closely resembles *Megalagus*. The presence of the insectivore *Centetodon* lends additional support (G. Morgan, personal communication, 1995) for an early Arikareean age as does the castorid *Agnotocastor* (G. Hayes, personal communication, 1997). In the Great Plains *Miohippus*, *Centetodon*, and *Agnotocastor* are last recorded in the early early Arikareean (Lillegraven et al., 1981; Woodburne, 1987; Tedford et al., 1996; Xu, 1996) and *Megalagus* last occurs at the Orellan/Whitneyan boundary (Prothero and Emry, 1996). The latter "holdover" taxon provides early evidence of the refugial aspect of the Gulf Coastal Plain noted by Webb et al. (1995), which became particularly evident in the latest Miocene when several Barstovian and Clarendonian taxa from the Great Plains persisted into the Hemphillian along the forested, subtropical Gulf Coast.

Additional taxa from Cow House Slough include an undetermined species of sciurid, a new large eomyid rodent, an entoptychine rodent, a tiny, probably new camelid that closely resembles *Nothokemas waldropi* with the exception of its even smaller size, an oreodont, and a small peccary resembling *Cynorca*.

White Springs Local Fauna. Terrestrial mammals from White Springs include the marsupial *Herpetotherium*, the rabbit *Archaeolagus*, the beaver *Neatocastor* (G. Hayes, personal communication, 1997), indeterminate species of a shrew and squirrel, the new eomyid noted above, the primitive cricetid rodent *Leidymys*, three different species of heteromyid rodents tentatively referred to *Proheteromys* (Morgan, 1989), an indeterminate species of canid, a horse of primitive "*Parahippus*" grade here referred to *Anchippus texanus*, a merycoidodontid oreodont, and a small indeterminate species of camelid (Morgan, 1989). Morgan (1989) noted that the large eomyid from White Springs and Cow House Slough also occurs in the Buda and SB-1A local faunas; that the large- and medium-sized heteromyids are also

TABLE 1. MAMMALIAN TAXA FROM ARIKAREEAN SITES IN FLORIDA*

	White Springs	Cow House Slough	Brooksville	Franklin Phosphate Pit No. 2	SB-1A	Buda	Martin-Anthony Oreodont	Martin Anthony
Marsupialia								
Herpetotherium	X				X			
Insectivora								
Soricid, indet.	X							
Centetodon sp.		X				X		
Parvericius sp.						X		
Chiroptera								
Vespertilionid, undet.	X				X			
Lagomorpha								
Archaeolagus sp.	X							
Megalagus sp.		X						
Rodentia								
?*Neatocastor*	X							
?*Agnotocastor*		X						
Sciurud gen. and sp. indet.	X	X						
Protosciurus sp.					X			
Texomys sp. §						X		
Eomyid, new gen. and sp. §	X	X			X	X		
Proheteromys sp. cf *P. magnus*	X				X	X		
Proheteromys sp. undet.§	X				X	X		
Proheteromys sp. of *P. floridanus*§	X				X			
Entoptychine		X						
Leidymys sp.	X							
Carnivora								
Daphoenodon notionastes§			X	X		X		
Mammocyon sp. cf. *M. obtusidens*					X			
Temnocyon sp.								X
Bassariscops achoros§						X		
Cynarctoides sp.						X		
Phlaocyon sp.					X			
?*Mesocyon* sp.					X			
Mustelidae gen. and sp. undet.						X		
Paroligobunis frazieri§					X			
Palaeogale sp.					X			
Nimravidae gen. and sp. indet.						X		
Perissodactyla								
Anchippus texanus	X			X				
Miohippus sp.		X						
Archaeohippus blackbergi								
Equidae gen. and sp. undet.					X	X		
Moropus sp.§						X		
Tapirid, undet.			X					
Tapirid, indet.				X				
Menoceras arikarense								X
Rhinocerotidae, new gen. and sp. §			X	X				
Artiodactyla								
Dinohyus sp.				X				
Cynorca sp.		X				X		
Merycoidodontid, undet.[1]	X							
Phenacocoelinae, undet.[2]		X						
Phenacocoelinae, undet.[3]			X			X		
Phenacocoelus luskensis[4]							X	
Camelidae, gen. and sp. indet.						X		
Camelidae, sm. new gen. and sp. §	X	X		X		X		X
?*Oxydactylus* sp.			X	X				
Nothokemus waldropi §				X	X			
Nanotragulus loomisi						X		

*Taxa totals: 43 genera, 46 species. The indeterminate soricid and equid are excluded from count; three genera of oreodonts; four species: 1. White Springs; 2. Cow House Slough; 3. Buda-Brooksville; 4. Martin-Anthony.
Genera endemic to the Gulf Coastal Plain (5).
§Species endemic to the Gulf Coastal Plain (11).

found at Buda and SB-1A; and that the smallest of the three White Springs heteromyids also occurs at SB-1A. Morgan (1989) suggested an early Arikareean age for White Springs on the basis of *Leidymys*, which last occurs in Ar1 in the Great Plains (Tedford et al., 1996). Strontium isotopic work by Jones et al. (1993) supports this (see below).

Brooksville, Franklin Phosphate Pit No. 2, and Buda local faunas. The sparse faunas from Brooksville and Franklin Phosphate Pit No. 2, together with the diverse assemblage from the Buda locality, are considered correlative based primarily on the mutual occurrence of the small Gulf Coastal Plain endemic amphicyonid, or "bear-dog," *Daphoenodon notionastes*. The Brooksville Local Fauna (Patton, 1967) also includes a small rhino, an undetermined species of tapir, a phenacocoeline oreodont, and a camel tentatively assigned to *Oxydactylus*. Additional taxa from Franklin Phosphate Pit include a rhino and the tiny new camelid noted above. The latter was originally assigned to "cf. *Blastomeryx*" by Simpson (1930), and Tedford et al. (1987, p. 178) noted a late Arikareean age for the fauna partly on the basis of this referral. Frailey (1979), however, found more diagnostic material of the same taxon in the Buda Local Fauna and determined that it belonged to a new species of camelid rather than to the moschid *Blastomeryx*. Another camelid at Franklin Phosphate Pit is represented by a single phalanx that Simpson (1930) suggested might belong to *Oxydactylus*, and there is an upper molar indistinguishable from that of *Nothokemus waldropi*. There is also a large entelodont referred to *Dinohyus* sp. and a horse that Simpson referred to *Parahippus* sp., but noted was likely distinct from *P. leonensis*. Frailey (1979) referred this horse to "Anchitheriinae gen. et sp. indet." It is here assigned to *Anchippus texanus*, as is the horse from White Springs, based on comparisons with abundant material from the Toledo Bend site referred to that taxon (see following section). A fragment of an edentulous mandible of a tapirid allows no further identification.

The Buda Local Fauna (Frailey, 1979) was the most diverse Arikareean fauna in the Gulf Coastal Plain prior to discovery of the Toledo Bend Local Fauna (and the new Brooksville 2 Local Fauna). The Buda assemblage is notable for its high diversity of carnivores (5 of the 13 large mammal species), its scant equid material (a single lateral phalanx), and an abundance of material belonging to a dwarf chalicothere. Chalicotheres were unknown elsewhere in the Gulf Coastal Plain until additional fossils were found at Toledo Bend. The rare geomyoid rodent, *Texomys*, is also found at Buda and Toledo Bend, thus representing its earliest appearance. *Texomys* was previously known only from the Hemingfordian Cucaracha Formation of Panama, and from the early Barstovian Moscow, Trinity River Pit, and Town Bluff Local Faunas of the Texas Coastal Plain (Slaughter, 1981). Carnivores in addition to *D. notionastes* include the canids "*Bassariscops*" *achoros* and *Cynarctoides* sp., a large mustelid of indeterminate genus and species, and a nimravine also of indeterminate genus and species (Frailey, 1979). Ungulates other than the horse and chalicothere (Frailey referred the former to Anchitheriinae gen. et sp. indet. and the latter to *Moropus* sp.) include the peccary *Cynorca* sp., the tiny new camelid noted above, a larger camelid of indeterminate taxonomic status referred to by Frailey (1979, p. 5156) as "*Oxydactylus*-like," a phenacocoeline oreodont, and the hypertragulid *Nanotragulus loomisi*. The hedgehog from the Buda Local Fauna originally referred by Rich and Patton (1975) to *Amphechinus* has recently been reassigned to *Parvericius* (G. Hayes, personal communication, 1997).

SB-1A Local Fauna. The SB-1A Local Fauna (Frailey, 1978) is typically considered correlative with those discussed above (Tedford et al., 1987), but few, if any, of its taxa are shared with the others. Like the Buda Local Fauna, SB-1A is unusual in the high diversity of carnivores (five out of eight mammal species) including the amphicyonid *Mammacyon* cf. *obtusidens*, the canids *Phlaocyon* sp. and "?*Mesocyon*" (later referred by Frailey [1979] to "Canidae gen. et sp. indet." and more recently to ?*Leptocyon* by Hayes), an indeterminate carnivore of slightly larger size than *Phlaocyon*, the mustelid *Paroligobunis frazieri*, and the weasel-like *Palaeogale*. Also similar is the rarity of horses (a single metapodial), which stands in "sharp contrast to the Hemingfordian faunas of Florida, especially Thomas Farm, in which horses are abundantly represented" (Frailey, 1978, p. 13; 1979). Other mammals from SB-1A include the squirrel *Protosciurus* sp., the endemic camelid *Nothokemas waldropi*, the marsupial *Herpetotherium* (G. Hayes, personal communication, 1997), and a vespertilionid bat. Frailey (1979) and Tedford et al. (1987) assigned the Buda and SB-1A faunas a late Arikareean age, noting a likely correlation with faunas from the Harrison Formation, Nebraska (Ar3). G. Hayes (personal communication, 1997), however, indicates that SB-1A may be slightly older based on his current work on the new Brooksville 2 Local Fauna.

Martin-Anthony local faunas. The Martin-Anthony oreodont site and the Martin-Anthony Local Fauna are the youngest of the Florida Arikareean localities. They are from a single roadcut locality, but the former is stratigraphically lower than the latter and an unconformity separates them (MacFadden, 1980). As both the oreodont, *Phenacocoelus luskensis*, which was found embedded in a marine limestone, and the mammals found in the overlying clastic sediments are of late Arikareean affinity, MacFadden (1980) concluded that the hiatus represented by the unconformity between them was of little significance. Because the type specimen of *P. luskensis* is from the Harrison Formation, Morgan (1993) considered the oreodont site to be of early late Arikareean age (Ar3). The overlying Martin-Anthony Local Fauna includes a large temnocyonine amphicyonid and the common Great Plains rhino, *Menoceras arikarense*, which Tedford and Hunter (1984) considered to be characteristic of the late late Arikareean. MacFadden (1980) also noted the presence of the small camelid that Frailey (1979) reported from the Buda Local Fauna.

Table 1 lists all mammals recorded to date from Arikareean localities in Florida. Taxa from Florida that last occur in the early Arikareean of the Great Plains include *Centetodon*, *Leidymys*, *Mesocyon*, and *Miohippus* (Tedford et al., 1987, 1996). Ranges were determined primarily on the basis of radioisotopically calibrated faunal occurrences in the Great Plains and western

interior. Taxa that first appear in the early Arikareean in the Great Plains include *Parvericius*, *Archaeolagus*, *Protosciurus*, *Temnocyon*, *Moropus*, *Dinohyus*, *Phenacocoelus*, and *Nanotragulus*. Late Arikareean first appearances include *Daphoenodon*, "*Parahippus*" (i.e., *Anchippus*), *Menoceras*, *Cynarctoides*, and *Oxydactylus*. *Phlaocyon* and *Mammacyon obtusidens* were previously known from the Monroe Creek and Harrison formations of early to late Arikareean age, respectively; *Phenacocoelus luskensis* and *Paroligobunis* were previously recorded from the Harrison Formation. *Protosciurus*, *Temnocyon*, *Daphoenodon*, and *Nanotragulus* are restricted to the Arikareean (and last occur in the late Arikareean), whereas *Moropus*, *Parahippus*, *Dinohyus*, and *Phenacocoelus* range into the early Hemingfordian.

The species of primitive "*Parahippus*" from Franklin Phosphate Pit and White Springs appear to be the same and are referred to *Anchippus texanus* based on comparisons with horse material from Toledo Bend. The small camelid in the White Springs, Cow House Slough, Franklin Phosphate Pit, Buda, and Martin-Anthony Local Faunas also represents a single species and appears closely related to *Nothokemas waldropi*. Whether the Brooksville and Buda oreodonts are the same has not been determined, although MacFadden (1980, p. 99) noted that they and the oreodont from the Martin-Anthony roadcut were "all represented by the Phenacocoelinae." The oreodont from White Springs, however, is a separate species, as is that from Cow House Slough.

Problems in correlation

It is difficult to correlate early Miocene vertebrate assemblages across the Gulf Coastal Plain because fossiliferous outcrops are limited and sites are broadly scattered geographically. In Florida, there is no lithostratigraphic continuum between localities because assemblages are typically preserved in fissure infillings, nor is there a physically well-developed stratigraphic succession over this interval of time. It necessarily follows that magnetostratigraphy is also of limited utility, and the lack of volcanic ash beds precludes radioisotopic age analysis.

Whereas White Springs, Cow House Slough, and the Martin Anthony oreodont site occur in nearshore marine depositional paleoenvironments, refined correlation to marine invertebrate biochronologies (i.e., nannofossil or foraminiferal zones) has been reported only for White Springs. There, Morgan (1989, p. 35) suggested an indirect correlation to the upper N4 to lower N5 subtropical planktonic foraminiferal zones based on indicative foraminifera in the Porters Landing Member of the Parachucla Formation in southern Georgia thought to extend into the area of White Springs. This correlation results in an age of about 21 to 22 Ma based on the time scale of Berggren et al. (1995), and Jones et al. (1993) reported an $^{87}Sr/^{86}Sr$ date of 20.2 Ma for the Porters Landing Member at its type section in Georgia. However, Jones et al. (1993) also obtained an $^{87}Sr/^{86}Sr$ date of 24.4 Ma on mollusc shells from the White Springs locality, thus corroborating the late early to early late Arikareean age indicated by the mammalian fauna.

Adding to the problem of correlation and accurate age determination is the limited taxonomic overlap between these faunas. For example; the Buda, Franklin Phosphate Pit, and Brooksville assemblages are considered correlative because of the mutual occurrence of only one species, *Daphoenodon notionastes*, but they have also been considered correlative with the SB-1A Local Fauna because both SB-1A and Buda have taxa (at least at the generic level) that share close affinities with High Plains forms. Yet the Buda, Franklin Phosphate Pit, and Brooksville faunas share few taxa with SB-1A (Table 1). Differences may be due to an ecological discontinuity, although the Buda and SB-1A local faunas are separated by no more than 75 km, are both inland from the coast, and both include amphicyonids, canids, mustelids, horses, and camelids. A preservational bias is also unlikely; fossils are obviously preserved at both localities. More likely, the SB-1A Local Fauna is temporally inequivalent to the others.

MacFadden and Webb (1982) suggested that the Buda Local Fauna may be somewhat older than SB-1A and more recent work by G. Morgan seems to substantiate this, as he has determined that *Centetodon* is present in the former. The presence of *Centetodon* and *Nanotragulus loomisi*, and the more primitive stage of evolutionary development of *Daphoenodon notionastes* relative to the late Arikareean Great Plains *Daphoenodon superbus* (R. Hunt, personal communication, 1994), supports a late early Arikareean age for the Buda Local Fauna (and therefore Brooksville and Franklin Phosphate Pit), although other taxa suggest a slightly later age.

In summary, the Cow House Slough and White Springs local faunas appear to be of late early Arikareean age (Ar2); the Brooksville, Franklin Phosphate Pit, and Buda Local Faunas appear to be of late early to early late Arikareean age (Ar2-A3); the SB-1A Local Fauna and the Martin-Anthony oreodont site appear to be early late Arikareean (Ar3) (the latter slightly younger than the former); and the Martin-Anthony Local Fauna is of late late Arikareean age (Ar4) (Fig. 2). Refined temporal resolution in a region where lithostratigraphic, biostratigraphic, and geochronologic control is limited will result only when additional sites of similar age are discovered that share taxa with well-understood temporal ranges. The rich new Brooksville 2 Local Fauna, currently under study, is one such site that will substantially improve the basis for faunistic comparisons. Another option in Florida is to find more sites in which land mammals are directly associated with marine invertebrates. These provide, as Tedford and Hunter (1984, p. 147) noted, "tests of hypotheses of correlation as well as making possible a broader chronological linkage that transcends the geographic and environmental barriers posed by the ecology of the organisms themselves."

ARIKAREEAN LAND MAMMAL AGE IN TEXAS

In this chapter, two local faunas in the Texas Coastal Plain and one in Big Bend National Park are considered Arikareean in age. The Castolon Local Fauna (Stevens et al., 1969; Stevens, 1977) from the Big Bend region of west Texas is eco-

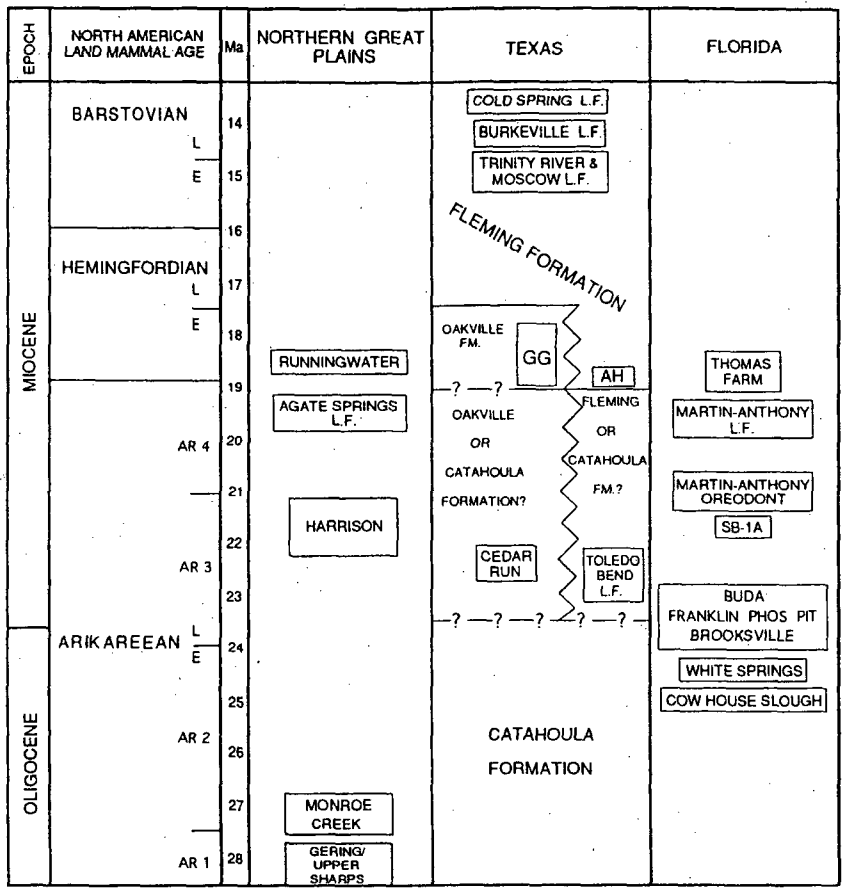


Figure 2. Correlation chart showing age and relative stratigraphic position of Arikareean and Hemingfordian local faunas of the Gulf Coastal Plain relative to others of similar age in the northern Great Plains. Some Garvin Gully local faunas, including Aiken Hill (AH) and Cedar Run, were previously reported as occurring near the base of the Oakville Formation and its lateral equivalent to the east, the lower Fleming Formation (top dashed line). In this report the Cedar Run Local Fauna is considered older than other Garvin Gully (GG) assemblages and similar in age to the Toledo Bend Local Fauna. Because the base of the Fleming Formation is mapped differently in western Louisiana than in eastern Texas, interpretations as to the placement of the Toledo Bend locality will differ accordingly (see text and Fig. 3). Hence the dashed lines with question marks. Placement of Great Plains faunas and mammal age boundaries and subboundaries follows Woodburne and Swisher (1995), Tedford et al. (1996), and MacFadden and Hunt (this volume).

logically separated from faunas of similar age along the Texas Coastal Plain and few, if any, species are shared. Stevens (1977) found the Castolon Local Fauna most similar to Arikareean faunas of the Southwest where expansion and evolution of a more xeric environment was occurring. Artiodactyls with moderately hypsodont dentitions dominate the fauna, including two oreodonts, five camelids, and a hypertragulid. A small rhino, *Moschoedestes delahoensis*, is also described as having subhypsodont teeth (Stevens et al., 1969). The absence of horses and the presence of the "Gila monster," *Heloderma*, further supports a xeric, rather than mesic, paleoenvironment (Stevens, 1977). Because this fauna occurs in Trans-Pecos Texas, its relevance to the Gulf Coastal Plain fauna is of limited utility with respect to the aims of this chapter, and thus is not considered further.

In the glossary of their report on the *Nomenclature and Correlation of the North American Continental Tertiary*, Wood et al. (1941) listed two local faunas from the "Oakville sandstone" of the Texas Coastal Plain: the Garvin Gully Local Fauna and the Cedar Run Local Fauna. Only the latter is currently considered Arikareean in age, whereas the Garvin Gully Local Fauna, as noted previously, represents the early Heming-

fordian. The second Arikareean fauna from the Texas Coastal Plain is the Toledo Bend Local Fauna.

The Garvin Gully Local Fauna includes fossils from localities in the vicinity of Navasota in Grimes and Washington Counties (Hay, 1924). Wood et al. (1941) placed this fauna in the "earliest Hemingfordian (or possibly, late Arikareean)," and their correlation chart shows it in the former. Realizing that fossil assemblages taxonomically similar to the Garvin Gully Local Fauna extended along the trace of the Oakville sandstone outcrop, Quinn (1952, p. 5) proposed the more inclusive term Oakville fauna, but in his 1955 publication changed the name back to the "Garvin Gulley fauna . . . to distinguish between biostratigraphic and stratigraphic units." Wilson (1956) discussed the concept of the terms local fauna and fauna and, in so doing, essentially formalized the use of the term Garvin Gully fauna by which these scattered assemblages within the Oakville Formation and its lateral equivalent to the east, the lower part of the Fleming Formation, have since been referred. Although Quinn (1952, 1955) and Wilson (1956, 1957, 1959, 1960) interpreted the mammals of the Garvin Gully Fauna to be indicative of the late Arikareean, Patton (1969a) found the artiodactyls to correlate best with those from the Hemingfordian-age Runningwater Formation of Nebraska and revised the age of the Garvin Gully Fauna accordingly. As Tedford et al. (1987) noted, the horses (Forsten, 1975a,b) and rhinos (Prothero and Manning, 1987) of the Garvin Gully Fauna also correlate best with those of the Runningwater Formation. Thus, the early Hemingfordian age of the Garvin Gully Fauna is accepted here.

Toledo Bend Local Fauna. The Toledo Bend Local Fauna (Albright, 1990, 1991, 1992, 1994, 1996; Manning, 1990) was recovered from a submerged paleochannel deposit in the lower Fleming Formation (as mapped in western Louisiana, see following section) near the Toledo Bend Reservoir, Newton County, Texas. Faunal affinities with the Great Plains and Florida suggest a medial Arikareean age. This assemblage of 17 lower vertebrate and at least 25 mammalian taxa differs from other early Miocene coastal plain faunas in the unusual abundance of riparian species and in the absence of grazing ungulates. Exceptional is the fact that three riparian forms, the hippo-like anthracothere *Arretotherium*, the tapir "*Miotapirus*" *marslandensis,* and a small protoceratid tentatively referred to *Protoceras*, account for the majority of mammalian remains recovered and none were known previously from the Gulf Coastal Plain. Also at Toledo Bend is the dwarf chalicothere *Moropus* sp. and the small bear-dog *Daphoenodon notionastes*, previously known from Arikareean sites in Florida, a new dwarf rhinoceros, and a giant entelodont. Oreodonts are absent, camels are exceptionally rare, and the predominant horse is the primitive "parahippine," *Anchippus texanus*.

Insectivores are rare with only a partial upper tooth having been recovered, and the only carnivore material other than that of *Daphoenodon notionastes* is a single, partial upper carnassial questionably referred to the small weasel-like *Palaeogale* (Albright, 1996). *Palaeogale* was previously reported in Florida at the Whitneyan-age I-75 locality (Patton, 1969b), but has recently been identified from the SB-1A Local Fauna, as well (G. Hayes, personal communication, 1997).

Toledo Bend produces at least six different species of rodents, including a beaver similar to "*Monosaulax*" *hesperus* (i.e., *Neatocastor hesperus*, see Korth, 1996; also see review by Xu, 1994), two new species of the heteromyid *Proheteromys*, the geomyoid *Texomys*, the squirrel *Protospermophilus*, and a putative new genus and species of eomyid or cricetid that resembles the European middle Oligocene cricetid *Pseudocricetodon montalbanensis* (Albright, 1996).

The seven perissodactyls and nine artiodactyls are all low-crowned browsing forms. The horses, rhinos, tapir, and chalicothere were mentioned earlier, as were most of the artiodactyls. The only camelid is a small form represented by a jaw fragment with two teeth that closely resembles the coastal plain endemic, *Floridatragulus*. The hypertragulid is placed in the genus *Nanotragulus*, but differs from *Nanotragulus loomisi*, the common Buda species, in its slightly larger size, and from *N. ordinatus* of the Castolon Local Fauna in its lower crowned teeth. There are at least two, but possibly three, peccaries, the smallest of which is referred to "*Cynorca*" *sociale*, and the entelodont is the size of *Dinohyus hollandi* but has peculiarities that suggest a possible new species (Albright, 1991).

The apparent age difference between the Toledo Bend and Garvin Gully faunas is not an artifact of different paleoecological settings. The tapir at Toledo Bend is the same species as that from the early Arikareean Monroe Creek Formation and the amphicyonid and dwarf chalicothere are the same species as those in the Buda Local Fauna. In the Garvin Gully Fauna, the Hemingfordian amphicyonid *Amphicyon longiramus* is present and there is no record of a chalicothere or tapir. The ubiquitous early Hemingfordian horse, *Parahippus leonensis*, so common in the Garvin Gully and Thomas Farm faunas, is absent at Toledo Bend, being replaced by the more primitive *Anchippus texanus*. A second, smaller horse at Toledo Bend resembles *Miohippus* sp. from the Cow House Slough Local Fauna more closely than the diminutive *Archaeohippus blackbergi* often found with *P. leonensis* in Gulf Coastal Plain early Hemingfordian faunas. Camels, too, are abundant and diverse in the coastal plain during the early Hemingfordian, yet they are extremely rare at Toledo Bend. The two larger rhinoceroses at Toledo Bend belong to the Arikareean genus *Diceratherium* rather than the early Hemingfordian *Menoceras barbouri*, and the dwarf rhino from Toledo Bend is similar to that from Brooksville. Finally, hypertragulids are present in the Toledo Bend and Buda faunas, but are absent in the Garvin Gully and Thomas Farm faunas. However, both the Toledo Bend and Garvin Gully faunas have a large entelodont and the common early Hemingfordian protoceratid, *Prosynthetoceras texanus*. It has not yet been determined if the entelodont from both faunas is the same species, and *P. texanus*, quite common at Garvin Gully and Thomas Farm, is very rare at Toledo Bend. At Toledo Bend, a small protoceratid that resembles most closely the Whitneyan to Arikareean genus *Protoceras* is the most abundantly occurring mammal. For these reasons the

Toledo Bend Local Fauna and the Garvin Gully Fauna are not considered temporal equivalents.

In such a well-sampled fauna (approximately 700 elements referred to mammals), it is intriguing that no oreodont remains were found. This is unusual considering that: (1) the early Miocene was their time of maximum diversity, (2) other mammals from the Great Plains (even rare ones) are abundant at Toledo Bend, and (3) oreodonts occur in the Whitneyan and Arikareean of Florida. On the other hand, oreodonts become rare by the Hemingfordian and are unknown thereafter in Florida.

Cedar Run Local Fauna. The other early Miocene assemblage from the Texas Coastal Plain listed by Wood et al. (1941) is the Cedar Run Local Fauna, from Washington County. In their original description, Wood and Wood (1937, p. 140) noted that most of the terrestrial vertebrates were "close to, if not identical with, forms from the Harrison [Formation] of Nebraska." Tedford et al. (1987) also noted the similarity with late Arikareean faunas of the Great Plains. Original taxonomic designations included *Blastomeryx texanus*, *Blastomeryx* cf. *primus*, an indeterminate genus of peccary, "Dog aff. *Tephrocyon*," ?*Palaeolagus* sp., *Palaeocastor* cf. *simplicidens* and *Palaeocastor* cf. *barbouri*, *Archaeohippus blackbergi*, a second, larger horse, and a rhinoceros. Wilson (1960) added "*Daphoenodon* cf. *superbus*," thus strengthening his view that the Garvin Gully Fauna, within which he included the Cedar Run Local Fauna, was late Arikareean in age. But the well-sampled assemblages that typify the Garvin Gully Fauna (e.g., Garvin Farm, Hidalgo Bluff), are now considered early Hemingfordian in age and few taxa from Cedar Run are found in them. It is here proposed that the Cedar Run Local Fauna be excluded from the Garvin Gully Fauna.

The Cedar Run Local Fauna is most similar to the Toledo Bend Local Fauna. They share at least three distinctive taxa: *Daphoenodon*, the tapir, and the small protoceratid. The rhino and horse from Cedar Run might also match those from Toledo Bend. Based on the abundance of material from Toledo Bend of species never before recorded, or only poorly known previously in the Texas Coastal Plain, taxonomic assessment of the Cedar Run Local Fauna can now be better evaluated. The following is a revision of that assemblage.

The report by Wilson (1960) of *Daphoenodon* cf. *superbus* from Cedar Run is, as noted above, likely one of the reasons he considered the Garvin Gully Fauna late Arikareean in age; the genus appears to be restricted to Ar2-4. But *Daphoenodon* is not present in any of the other well-sampled assemblages that comprise the Garvin Gully Fauna (nor is it recorded from the Thomas Farm Local Fauna in Florida), although another amphicyonid, the early Hemingfordian *Amphicyon longiramus*, is present (Wilson, 1960). Thus, to indicate that the age of the Garvin Gully Fauna is late Arikareean based on the presence of *Daphoenodon*, and that *Daphoenodon* and *Amphicyon* were "found associated" (Wilson, 1960, p. 998) is somewhat misleading. In the Gulf Coastal Plain, *Daphoenodon* is a component of medial to late Arikareean faunas and *Amphicyon* occurs in the early Hemingfordian. Because *Daphoenodon notionastes* occurs in the Toledo Bend Local Fauna, and because other taxa are shared as well, the single bear-dog upper molar from Cedar Run might better be referred to *D. notionastes*. Future studies are needed to confirm this.

The teeth from Cedar Run referred to *Blastomeryx texanus* were later referred to *Prosynthetoceras texanus* by Patton and Taylor (1971, p. 140). Teeth referred to *Blastomeryx* cf. *primus*, however (AMNH 30084, 30085), are smaller and less hypsodont than those referred to *P. texanus* and most closely resemble teeth of the small protoceratid from Toledo Bend tentatively referred to *Protoceras*.

The partial tooth originally referred by Wood and Wood (1937) to Peccary, gen. indet. (AMNH 30086) matches the talonid of lower molars from the tapir at Toledo Bend. Based on two teeth found in Nebraska in 1917, this tapir was originally referred to *Miotapirus marslandensis* by Schoch (1984). Study of the nearly complete dentition of this species from Toledo Bend, however, indicates that a new genus is represented; a detailed description is in press.

The rhinoceros maxilla, originally referred to *Caenopus*, was not found at the same locality as the other fossils, but about 400 m west of the Cedar Run locality in another small drainage called Water Run (Wood and Wood, 1937, p. 137). No other mammal fossils were found associated with the rhino specimen. Prothero and Manning (1987), in their review of rhinos from the Texas Coastal Plain, considered this specimen a representative of the common latest Arikareean Great Plains form, *Menoceras arikarense*. Although the specimen is highly worn, thus making comparisons and accurate taxonomic identification difficult, personal examination found its morphology more similar to *Diceratherium annectens* (Albright, 1991, p. 177).

At least two horses occur in the Toledo Bend and Cedar Run local faunas. The small Cedar Run horse appears to be referable to the diminutive *Archaeohippus blackbergi*; it is smaller than the horse from Toledo Bend referred to *Miohippus*. It has not yet been determined if the larger horse from Cedar Run, which is represented by a single fragmented tooth, might be the larger Toledo Bend horse *Anchippus texanus*.

The Cedar Run and Toledo Bend beavers are separate species. Whereas the Toledo Bend species resembles *Neatocastor hesperus*, referral of the smaller Cedar Run beaver to the common Arikareean genus *Palaeocastor* is maintained pending further study. Table 2 includes the revised list of mammals from the Cedar Run Local Fauna.

Aiken Hill Local Fauna. Another small assemblage previously considered late Arikareean in age is that from Aiken Hill in Walker County. The earliest reference to this locality is by Hesse (1943) who spelled it "Akin Hill" (as did Patton and Taylor, 1971). It is only briefly mentioned again by Quinn (1955) and Wilson (1957, 1962), who both referred to it as Aiken Hill and included it within the Garvin Gully Fauna. Hesse's (1943) faunal list included the mammals *Blastomeryx* sp., a medium-size camel, an indeterminate rhino, and *Prosynthetoceras*. Most of the material is of poor quality, although a skull and associated postcrania of *Prosynthetoceras texanus* found there were described by Patton and Taylor (1971).

TABLE 2. MAMMALIAN TAXA FROM THE CEDAR RUN AND TOLEDO BEND LOCAL FAUNAS*

	Cedar Run Local Fauna	Toledo Bend Local Fauna
Insectivora		
Fam. gen., sp. indet.		X
Lagomorpha		
Palaeolagus sp.	X	
Rodentia		
Protospermophilus sp.		X
cf. *Neatocastor hesperus*		X
Palaeocastor cf. *P. simplicidens*	X	
Palaeocastor cf. *P. barbouri*	X	
Eomyidae or Cricetidae, gen. and sp. undet		X
Proheteromys toledoensis§		X
Proheteromys sabinensis§		X
Texomys sp. §		X
Carnivora		
Palaeogale?		X
Daphoenodon notionastes§	X	X
Perissodactyla		
Anchippus texanus	?	X
Miohippus sp.		X
Archaeohippus blackbergi	X	?
Moropus sp.§		X
Motapirus marslandensis	X	X
Diceratherium annectens	X	X
Diceratherium armatum		X
Rhinocerotidae, new gen. and sp. §		X
Artiodactyla		
Dynohyus sp.		X
Cynorca sociale		X
?*Hesperhys* sp. cf. *H. pinensis*		X
?*Hesperhys* sp.		X
Arretotherium acridens		X
Floridatragulus sp. §		X
Prosynthetoceras texanus§	X	X
?*Protoceras* sp.	X	X
Nanotragulus sp.		X

*Taxa totals: 25 genera, 28 species. Two species of *Palaeocastor* counted as one.
Genera endemic to the Gulf Coastal Plain (3).
§Species endemic to the Gulf Coastal Plain (8).

Patton and Taylor (1971) documented a relatively unbroken evolutionary sequence of Gulf Coastal Plain protoceratids that increased in size through sequentially younger stratigraphic levels. This sequence started with the relatively diminutive Garvin Gully specimens of *P. texanus*, to the somewhat larger Burkeville specimens of the same species, through *P. francisi* in the Barstovian Cold Spring Local Fauna, to the still larger *Synthetoceras tricornatus* in the Clarendonian Lapara Creek Fauna. Some of the Aiken Hill specimens of *P. texanus* (e.g., F:AM 34181, TMM-TAMC 40422-4), but not all (e.g., F:AM 53500), are smaller than other specimens of the same taxon in the Garvin Gully and Thomas Farm Local Faunas suggesting a slightly older age for Aiken Hill, but these specimens do not differ in morphology. The teeth are relatively just as hypsodont and derived in their morphology as those from the early Hemingfordian faunas. Teeth of the small protoceratid from Toledo Bend are even smaller than those from Aiken Hill and they retain primitive features absent in the Aiken Hill specimens. Tedford et al. (1987) considered the fauna from Aiken Hill correlative with that from Cedar Run in part because of its presumed stratigraphic occurrence near the base of the Fleming Formation. But the base of the Fleming Formation is poorly defined, particularly in east Texas (see below), resulting in uncertain stratigraphic placement of many localities. The Aiken Hill Local Fauna is here considered older than the major assemblages that typify the Garvin Gully Fauna, such as those from Garvin Farm or Hidalgo Bluff, but younger than the Cedar Run Local Fauna, perhaps latest Arikareean to earliest Hemingfordian, based on the stage of evolutionary development of *Prosynthetoceras*.

In summary, the Cedar Run Local Fauna, from immediately above what Wood and Wood (1937) considered the contact between the Oakville Formation and underlying Catahoula Formation in Washington County, is here considered nearly correlative with the medial Arikareean Toledo Bend Local Fauna and is not considered a part of the early Hemingfordian Garvin Gully Fauna (Fig. 2). The Aiken Hill locality, and perhaps a few other single taxon sites in the lower Fleming Formation (Wilson, 1957), may be of latest Arikareean to earliest Hemingfordian age, i.e., intermediate in age between the older Cedar Run Local Fauna and those younger local faunas of early Hemingfordian age that typify the Garvin Gully Fauna. The Cedar Run Local Fauna should not be equated with the Aiken Hill and other "base of the Oakville" Garvin Gully local faunas because the latter have taxa more characteristic of the early Hemingfordian. Table 2 lists the mammals recorded from localities in the Texas Coastal Plain considered Arikareean in age.

Problems in correlation

The Toledo Bend site is located within about 400 m of the Texas/Louisiana border and is stratigraphically placed in the context of western Louisiana geology (Albright, 1991). Across this state boundary interpretations regarding the contact between the Fleming Formation and the underlying Catahoula Formation differ. Consequently, investigation into the stratigraphic placement of the site raises further questions about a historically confusing early Miocene Texas Coastal Plain stratigraphy.

Fisk (1940), Welch (1942), and Rogers and Calandro (1965), determined that the Fleming Formation in Vernon Parish, Louisiana (immediately east of Newton County, Texas), is divisible into six members. These members are recognizable in electric resistivity logs (see Rogers and Calandro, 1965, plate 2), and the Toledo Bend locality occurs in the region where one of the lower members, the Carnahan Bayou Member, crops out (Fig. 3).

Welch (1942, p. 51–52) discussed the depositional environment of the Carnahan Bayou Member and noted the following:

The Carnahan Bayou Member is homogeneous only in its heterogeneity and is mapped separately chiefly because it lies between two mappable

calcareous members. It consists dominantly of gray clay, silt and sand, which are sporadically consolidated into shale, siltstone and sandstone. A few outcrops of brown shale, quartzitic sandstone, black chert gravel, and lignite were found. Bentonite, reworked and redeposited with clay and silt, is common in the central portion of the member but also occurs in the lower part. . . . The Carnahan Bayou member is regarded as fluviatile in origin and the sands and gravels as channel deposits.

In east Texas, on the other hand, the Fleming Formation is not separated into formally recognized members (Galloway et al., 1986; Galloway et al., 1991). Furthermore, outcrop patterns of the Catahoula and Fleming formations, as shown in the *Geologic Atlas of Texas* (Barnes, 1967), the *Geologic Map of Louisiana* (Snead and McCulloh, 1984), and Welch's (1942) geologic map of Vernon Parish, are not congruent across the Texas/Louisiana border. As Figure 3 shows, the upper Catahoula Formation in eastern Texas is mapped as the Carnahan Bayou and Lena members of the Fleming Formation in western Louisiana. As a result, the Toledo Bend locality would be placed in the Catahoula Formation, rather than the lower Fleming Formation, in the context of east Texas geology. This may explain why Texas Coastal Plain faunas located farther to the west and placed near the base of the Fleming Formation, as mapped in Texas, are paleontologically more recent than the Toledo Bend Local Fauna, which is more than 198 m above the base of the Fleming Formation, as mapped in Louisiana (Albright, 1991). The concept of a "base of the Fleming Formation" has a different meaning depending on whose interpretation, or which geologic map, one chooses. What is considered the base of the Oakville and Fleming Formations in central to eastern Texas, where early Hemingfordian assemblages included in the Garvin

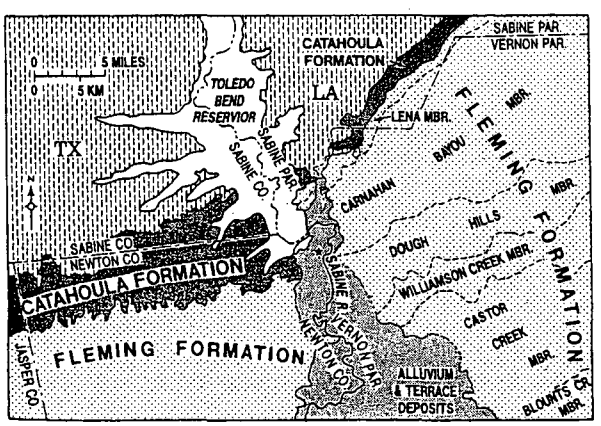


Figure 3. Diagram showing mapping discrepancy of the Fleming Formation across Texas-Louisiana border. Star denotes Toledo Bend locality. The site is considered to be located in the Carnahan Bayou Member of the Fleming Formation according to Louisiana geologic interpretations, whereas it would be placed in the upper part of the Catahoula Formation by Texas interpretations. Compiled from Barnes (1967) and Snead and McCulloh (1984). Reproduced from Albright (1996) with permission from Journal of Vertebrate Paleontology.

Gully Fauna are found, does not represent the same chronozone as that represented by the base of the Fleming Formation in western Louisiana near the Toledo Bend locality (Fig. 2).

Correlation problems in this region date back to Veatch's work in 1906. Welch (1942, p. 45–46) summarized the problem and noted that the Catahoula/Fleming contact was never clearly defined in the first place and that later criteria used to locate it were misleading. One explanation is the occurrence of at least three different fluvial facies within the Fleming (and Oakville) Formation: "channel sand, point bar sand, and back-swamp clay" (Wilson, 1968, p. 138). Consider the following quote by Galloway et al. (1991, p. 247) regarding Gulf Coast stratigraphic subdivision and terminology:

> The Cenozoic stratigraphic section over most of the northern and northwestern flanks of the [Gulf of Mexico] basin is dominantly composed of very thick, laterally variable, and monotonous sequences of interbedded sandstones, siltstones, and shales difficult to break down into distinctive lithostratigraphic units (groups, formations, and members) that can be recognized and mapped over appreciable areas. The boundaries, both vertical and lateral, between recognized units are generally transitional, and in most cases vague, subjective, and the topic of durable controversies, the result of the recurrence of closely similar and related environments that persisted in the Gulf of Mexico basin during the entire Cenozoic.

Although the Miocene stratigraphy of the Texas Coastal Plain is reasonably well understood due to intensive exploration efforts by the petroleum industry, lithostratigraphy is still of limited and tenuous utility when applied to correlating the widely distributed vertebrate faunas found in the updip, nonmarine outcrops of the lower Fleming Formation in eastern Texas and western Louisiana. This is due to the lateral variability of nonmarine sediments deposited in a low-relief fluvial and deltaic environment. Referring to the updip, nonmarine, facies of the Catahoula and Fleming Formations, Wilson (1968, p. 136) pointed out that they cannot be thought of as single lithologic units that are "carried along strike like time-stratigraphic units." These fluvial and delta-plain sediments were deposited by a dynamic feeder system that saw a shift from Texas in the Oligocene to eastern Texas-western Louisiana in the early Miocene (Galloway et al., 1986, 1991). Glowacz and Horne (1971, p. 385), in studying the Catahoula/Fleming contact in Rapides Parish, Louisiana, found that, because the boundary represents a depositional facies change from channel fill, natural levee, and swamp facies in the Catahoula Formation to lacustrine, swamp, and abandoned channel facies in the Fleming, it is "not the same age anywhere."

Biostratigraphic implications. The Miocene deposits of the Texas Coastal Plain have typically been separated into four biochronologic divisions (Hesse, 1943; Quinn, 1952, 1955; Wilson, 1956, 1957, 1959, 1960; Patton, 1969a; Forsten, 1975b; Tedford et al., 1987). The oldest, prior to the discovery of the Toledo Bend Local Fauna, was characterized by the well-known, early Hemingfordian Garvin Gully Fauna, as discussed previ-

ously, which is found in the Oakville Formation, but also includes local faunas from the lower part of the laterally equivalent Fleming Formation as far east as San Jacinto County (Wilson, 1957). Spradlin (1980) noted that typical Oakville Formation lithology was traceable in surface outcrop as far east as Huntsville, Walker County, Texas, although the *Geologic Atlas of Texas* (Barnes, 1974) shows the Oakville Formation extending no farther east than the Brazos River (the Brazos County/Grimes County boundary). Stratigraphically higher in the Fleming Formation is the medial Barstovian Burkeville Fauna, followed still higher in the section by the late Barstovian Cold Spring Fauna. The fourth biochronologic division is characterized by the early Clarendonian Lapara Creek Fauna from the overlying Goliad Formation (Fig. 2). It is here proposed that a fifth biochronologic division be recognized, that which includes the "medial" to late Arikareean Toledo Bend Fauna, within which are included the Cedar Run and Toledo Bend local faunas. The Toledo Bend Fauna, therefore, would characterize the oldest of five biochronologic divisions of the Texas Coastal Plain Miocene. Problems with the lithostratigraphic placement of these assemblages were discussed above and are depicted through the use of question marks and dashed line "formation" boundaries in Figure 2. In summary, the Toledo Bend Local Fauna occurs in the Catahoula Formation according to geologic mapping of east Texas, which conforms stratigraphically with the occurrence of early Hemingfordian faunas in the overlying Oakville and lower Fleming Formations, but in the Carnahan Bayou Member of the Fleming Formation according to the geologic map of Louisiana.

Implications regarding midcontinent and Gulf Coastal Plain biogeographic provinces

Ten vertebrate assemblages in the Gulf Coastal Plain considered Arikareean in age have produced approximately 65 mammalian species, including those of indeterminate or as yet undetermined affinities (Tables 1, 2). This count is also based on a number of assumptions that may be reevaluated as future studies are made and more material recovered. The assumptions are:

1. That the two species of *Proheteromys* in Texas are different from those in Florida;
2. That the primitive parahippine from White Springs, Franklin Phosphate Pit, Toledo Bend, and Cedar Run, are the same taxon, i.e., *Anchippus texanus;*
3. That the indeterminate genus and species of equid from SB-1A and Buda may be equivalent, but material is too limited to assume referral to *Anchippus texanus*;
4. That the new genus and species of dwarf rhinoceros from Toledo Bend is the same taxon as at Brooksville and Franklin Phosphate Pit;
5. That the tapir from Toledo Bend is not the same species as that from Brooksville;
6. That *Dinohyus* sp. from Franklin Phosphate Pit and "*Dinohyus*" sp. from Toledo Bend are tentatively considered the same taxon;
7. That *Cynorca* sp. from Cow House Slough and Buda and "*Cynorca*" *sociale* from Toledo Bend are the same taxon;
8. That the oreodonts from Buda and Brooksville are the same species—they may belong to the same genus as the Martin-Anthony species, and that the White Springs and Cow House Slough oreodonts represent two additional genera (three genera, four species);
9. That the small new genus and species of camelid from White Springs, Cow House Slough, Franklin Phosphate Pit, Buda, and Martin-Anthony is considered the same taxon, but there is also a second, indeterminate taxon at Buda;
10. That *Nanotragulus loomisi* from Buda and *Nanotragulus* sp. from Toledo Bend are different species; and
11. That the two Cedar Run beavers are considered one species of *Palaeocastor*, that those from Toledo Bend and White Springs are considered the same taxon (?*Neatocastor*), and that the beaver from Cow House Slough is a third taxon (?*Agnotocastor*).

It has long been recognized that the Gulf Coastal Plain was a distinct biogeographic province during the early Miocene. Quinn (1955) and Wilson (1956) noted this in their early studies and later, more detailed studies on specific groups of mammals, such as the artiodactyls (Patton, 1969a; Patton and Taylor, 1971), horses (Forsten, 1975a,b), and rodents (Slaughter, 1981), provided corroboration. But when did this distinction begin? When the above mentioned authors were actively studying the Miocene of the Gulf Coastal Plain, the oldest well-sampled faunas were those of early Hemingfordian age found at Garvin Gully and Thomas Farm. Because a faunal distinction between the midcontinent and the Gulf Coastal Plain was already apparent in these important assemblages, some kind of zoogeographic/climatic filter must have been emplaced prior to the early Hemingfordian.

When Arikareean mammals were discovered in Florida, similar conclusions regarding separate paleobiogeographic provinces were reached on the basis of such groups as the carnivores (Frailey, 1978, 1979) and oreodonts (MacFadden, 1980). Although the Florida faunas, including the Whitneyan-age I-75 Local Fauna (Patton, 1969b), maintained a basic resemblance to Arikareean faunas of the Great Plains, they also included, as Frailey (1978, p. 19) noted, the "added dimension of regional differentiation." The indigenous fauna that characterized this regional differentiation is commonly termed the early Miocene Gulf Coast Chronofauna (Webb, 1977; Prothero and Sereno, 1982; Tedford et al., 1987). But the underlying resemblance to Great Plains faunas should not be understated; a faunal link between Gulf Coast and Great Plains faunas (Webb, 1977) was never severed.

Approximately 54 genera representing 65 species are recorded from Arikareean sites in the Gulf Coastal Plain: 43 genera from the eight sites in Florida and 25 from the two Texas localities (Fig. 4A,B). Thirty-eight of the 41 identified genera (92%) are also recorded in the Great Plains; 29 from Florida and 20 from Texas. Only six genera (14.6%), five of which occur in Florida and three of which occur in Texas, are endemic to the

Gulf Coastal Plain. These endemics are *Texomys*, the new genus of eomyid, the new genus of dwarf rhino, the new genus of small camelid, *Nothokemas*, and *Floridatragulus*. *Texomys* and the dwarf rhino are found in Texas and Florida. Of the 46 Florida species, at least 11 (24%) are endemic to the Gulf Coastal Plain (Fig. 4B). Many others may be, but limited material precludes species level identifications. Nevertheless, of the 11 endemic species, 6 (54%) share generic level affinities with Great Plains taxa: *Proheteromys* cf. *P. floridanus* and *Proheteromys* sp. undet., *Daphoenodon notionastes*, "*Bassariscops*" *achoros*, *Paroligobunis frazieri*, and *Moropus* sp. This number will likely increase as additional genera for which there is ample material, such as *Centetodon*, *Parvericius*, *Archaeolagus*, *Megalagus*, *Protosciurus*, *Leidymys*, *Phlaocyon*, *Palaeogale*, *Miohippus*, *Oxydactylus*, and the oreodonts, are studied in detail. Numerous small mammals from the new Brooksville 2 Local Fauna currently under study may also substantially increase the number of endemic taxa on record. The 25 Texas genera are represented by 28 species, 8 of which are endemic to the region (Fig. 4B). Of these eight, five (63%) share generic level affinities with Great Plains taxa: *Proheteromys toledoensis*, *P. sabinensis*, *Daphoenodon notionastes*, *Moropus sp.*, and *Prosynthetoceras texanus*. This level of generic similarity between the Gulf Coastal Plain and the Great Plains implies that, at the onset of the Miocene, the presumed midcontinent/coastal plain filter cannot have been emplaced for very long.

Different species of the same genera in both the Gulf Coastal Plain and the Great Plains confirm that the two provinces were ecologically distinct, but this distinction may have begun only slightly earlier than the medial Arikareean when the Gulf Coast forms first appear. Noting the dominance of High Plains immigrant taxa in the medial Barstovian Trinity River Local Fauna, Texas (Fig. 2), Webb (1977, p. 370) showed that faunal continuity with the Great Plains Fauna had been

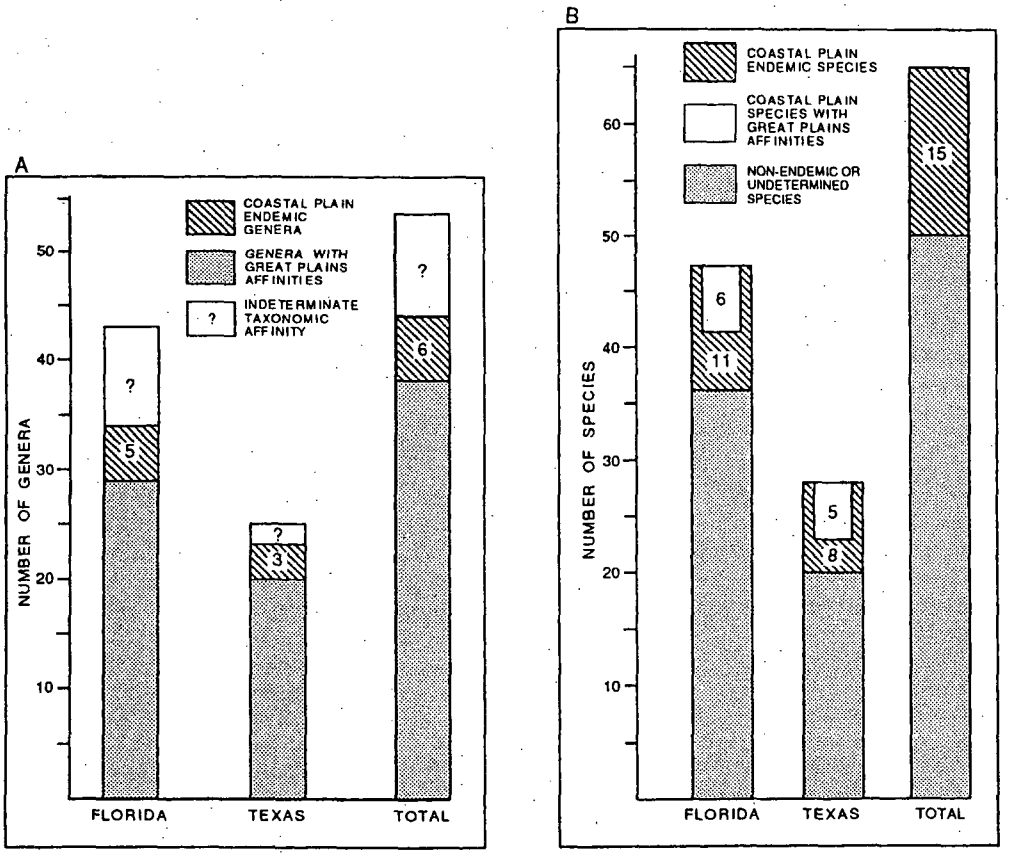


Figure 4. A, Bar chart showing the number of genera from Arikareean localities in Florida and Texas including endemic taxa and those sharing affinities with northern Great Plains forms. B, Bar chart showing the number of species from Arikareean localities in Florida and Texas. Endemic species sharing generic level affinities with Great Plains genera are distinguished from species of Coastal Plain endemic genera.

reestablished by that time and that the "peculiarities of the Gulf Coast Fauna [had] nearly disappeared. . . ." Thus the isolation of the Gulf Coast fauna appears to have peaked during the Arikareean through Hemingfordian. The Toledo Bend Local Fauna is the best view yet into the timing of midcontinent/Gulf Coastal Plain faunal disparity and may, together with those Arikareean faunas in Florida, represent the acme of endemism in the early Miocene Gulf Coast. Unfortunately, resolution beyond that which can be obtained through the "Toledo Bend window" is doubtful because of the near absence of exposed outcrops of pre-Arikareean nonmarine sedimentary sequences that might harbor the fossils necessary for such an investigation.

Prothero and Sereno (1982, p. 25–26) discussed the high diversity of the early Miocene Gulf Coast fauna as resulting from ecotone conditions, i.e., overlapping geographic ranges of Gulf Coastal Plain endemic taxa with those of immigrant High Plains forms. The Toledo Bend Local Fauna supports this hypothesis, and provides perhaps the best example of this mixing of faunas. Although the Toledo Bend Local Fauna includes numerous Gulf Coast endemic species, there are also species previously known only from Arikareean faunas of the northern Great Plains. "*Miotapirus*" *marslandensis* occurs in the medial Arikareean Monroe Creek Formation, South Dakota, and the type specimen is presumably from the early Hemingfordian Runningwater Formation, Nebraska; the anthracothere *Arretotherium acridens* was originally described from the Arikareean of Montana; the rhino *Diceratherium* is known primarily from the Arikareean of Oregon and the Great Plains; and Arikareean species of *Protoceras* are recorded only from South Dakota, Nebraska, and Wyoming (interestingly, Whitmore and Stewart [1965] reported *Diceratherium* and a primitive protoceratid they referred to *Protoceras* from Panama. Patton and Taylor [1973], however, later referred the protoceratid to *Paratoceras*, a genus known only from the Texas Coastal Plain). Important here is that these typically Great Plains species are the most abundantly represented mammals at Toledo Bend. Thus, while some groups from the Great Plains were evolving into new species more suited to the subtropical environments of the Gulf Coastal Plain, certain other groups indicate that barriers to interchange between the two regions were evidently only weakly imposed. The only late Arikareean Great Plains groups that appear to have been filtered out of entry into the Gulf Coastal Plain during that time are grazing ungulates such as the rhino *Menoceras*, large species of chalicotheres like *Moropus elatus*, grazing oreodonts, and stenomyline camelids. Large ungulates that were not dependent on grazing, such as entelodonts and the browsing species listed above, passed through the filter.

It may be that the Gulf Coastal Plain records of these Great Plains species represent the terminal occurrences of these lineages. As common as these species are at Toledo Bend, they are absent in the well-sampled early Hemingfordian faunas of Texas and they are not known from any of the Arikareean or early Hemingfordian sites in Florida. The only exception is a fragmented upper molar and lower molar fragment of an anthracothere in Garvin Gully collections at the Texas Memorial Museum (see Albright, 1991, p. 220). In the early Hemingfordian, the mixed feeder *Parahippus leonensis*, the subhypsodont rhino *Menoceras barbouri*, and higher crowned camelids are common, thus confirming a weakening in the Great Plains/Coastal Plain faunal disparity that Webb (1977) noted had nearly disappeared by the medial Barstovian.

In addition to a Gulf Coastal Plain–Great Plains faunal link during the earliest Miocene, certain lower vertebrate members of the Toledo Bend Local Fauna, such as the large river turtle *Dermatemys* and the coastal fish *Centropomus* (Albright, 1994), indicate that there was also a link to the Neotropical realm. Today the Central American river turtle, *Dermatemys mawei*, is wholly restricted to southern Mexico and Central America, and the common snook, *Centropomus undecimalis*, is found only in coastal waters of southern Texas, southern Florida, the southern Gulf of Mexico, and the Caribbean Sea (Albright, 1994, Fig. 5.3, and references within). Gulf Coastal Plain genera not previously recorded in Great Plains faunas likely originated in regions to the south (Slaughter, 1981; Whitmore and Stewart, 1965) and thus mark the early phase of expansion of the Neotropical biogeographic province into the southeastern United States (Webb, 1977; MacFadden and Webb, 1982).

Obviously, the Great Plains and the Gulf Coastal Plain represented different environments by the Arikareean. The trend that began in the mid- to late Eocene toward a cooler, drier world climate, culminating in the mid-Oligocene "Big Chill" (see Prothero, 1994), was effectively transforming the midcontinent, in the early Miocene, from a woodland savanna community with "a complex mosaic of different local habitats with corridors of gallery forests persisting along fluvial systems" (Webb, 1989, p. 189) into a region of open grasslands with forests essentially confined to river courses (Retallack, 1992). Vast interchannel plains mantled with fine-grained, eolian transported volcaniclastic loess and bone-packed desiccated waterholes in northwestern Nebraska (Hunt, 1990) paint a significantly different picture than that inferred by a fauna less than 100 km from an early Miocene Gulf of Mexico whose major constituents were alligators, large soft-shelled turtles, tapirs, and anthracotheres (Albright, 1994). That the climate between the changing midcontinent and more equable Gulf Coast was different is obvious, but the Toledo Bend Local Fauna confirms that a true barrier to free dispersal was never securely emplaced. Browsing ungulates that inhabited forested environments of the midcontinent were able to expand southward into the lush subtropical regime along the Gulf Coast, while grazing forms adapted to the more open environments of the woodland savanna community remained confined to their northern range. As the Miocene progressed, an eastward expansion into the gulf region of a woodland savanna corridor "that earlier had been largely confined to the Rio Grande Trench" (Webb, 1977, p. 371) resulted in a more open environment into which the more grazing adapted forms earlier confined to the northern Great Plains could finally expand. It was this event that resulted in the dwindling and eventual extinction of the early Miocene Gulf Coast browsing fauna

and the subsequent continuity of northern and southern faunas by the medial Barstovian. Although these ideas are not new, they are now much better supported and temporally constrained by the Toledo Bend Local Fauna.

CONCLUSIONS

The Arikareean North American Land Mammal Age in the Gulf Coastal Plain is represented primarily by 10 different localities in two states, Texas and Florida. Most of the approximately 65 species, representing 54 genera, share close affinities with taxa of similar age from the northern Great Plains. More than 90% of all identified Gulf Coastal Plain Arikareean genera are also known in the Great Plains, whereas only six genera are endemic to the gulf region. Thus, the mammalian disparity between the midcontinent and the Gulf Coastal Plain appears to have been a primarily Arikareean and early Hemingfordian event that progressively collapsed as a northwardly expanding Neotropical woodland savanna environment allowed access to the gulf region by the rapidly developing midcontinent grazing fauna. In the early Arikareean, a general drying trend and change of habitus in the midcontinent sent many mammals, particularly riparian adapted browsers, to more suitable environments throughout the Gulf Coastal Plain. But a northward and eastward expansion of the Neotropical realm into the gulf region destabilized the Gulf Coast endemic fauna. This eventually allowed occupation by immigrant Great Plains forms, thereby reestablishing midcontinent–coastal plain faunal continuity by the Barstovian.

The Toledo Bend Local Fauna from easternmost Texas provides one of the best opportunities to investigate the nature and timing of midcontinent–Gulf Coastal Plain biotic disparity because it includes: (1) numerous species found previously only in the northern Great Plains; (2) Gulf Coastal Plain endemic species shared with Florida whose generic affinities implicate a previous link with northern taxa; and (3) taxa indicative of an early phase of the northwardly expanding Neotropical component.

The Arikareean faunas from the Gulf Coastal Plain are dated on the basis of their similarity to genera and species from geochronologically more accurately dated localities of the continental interior, or from their stage of evolutionary development. Lithostratigraphic criteria are of little benefit with respect to correlation of Arikareean faunas in Florida because of the nature of the context in which the faunas are found, that is, widely distributed karst infillings. The stratigraphy of the Texas Coastal Plain is more straightforward and understood to a greater extent than in Florida, but pervasive facies changes along the strike of the fluvial and deltaic sedimentary sequences that harbor Arikareean faunas complicate attempts to accurately correlate the localities. Future work in both Texas and Florida, together with additional finds in the states between, will no doubt aid in refining the correlation of those assemblages already known, as well as provide new information useful for further understanding the Gulf Coastal Plain as a separate biogeographic province during the early Miocene. Basking in a warm, humid, equable climate, the Gulf Coastal Plain throughout the latest Oligocene–earliest Miocene was ideally situated for the evolution of a separate biogeographic province as northern forms sought refuge from a changing climatic regime and southern forms found new territory in which to expand.

ACKNOWLEDGMENTS

I express my appreciation to Robert Hunt for his invitation to contribute a paper to the Symposium on Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups and for his interest in the early Miocene of the Gulf Coastal Plain; and to Gary Morgan for many helpful discussions regarding Arikareean sites in Florida. M. O. Woodburne reviewed an early draft and provided helpful suggestions. A. Barnosky, R. Emry, R. Hunt, and D. Webb reviewed the manuscript and provided helpful comments that improved the chapter. Discussions with G. Hayes were also helpful and much appreciated. This work is University of Florida Contribution to Vertebrate Paleontology No. 491.

REFERENCES CITED

Albright, L. B., 1990, A submerged early Miocene site in the Fleming Formation near Toledo Bend Dam, Texas: Journal of Vertebrate Paleontology Abstracts of Programs, v. 10, p. 12A.

Albright, L. B., 1991, The vertebrate paleontology, taphonomy, and paleoecology of a new early Miocene site in the Fleming Formation near Toledo Bend Dam, Newton County, Texas, and its importance to Gulf Coastal Plain biostratigraphy [M.S. thesis]: Baton Rouge, Louisiana State University, 319 p.

Albright, L. B., 1992, New remains of the rare early Miocene tapir, *Miotapirus marslandensis*, from the Texas Coastal Plain: Journal of Vertebrate Paleontology Abstracts of Programs, v. 12, p. 15A.

Albright, L. B., 1994, Lower vertebrates from an Arikareean (earliest Miocene) fauna near the Toledo Bend Dam, Newton County, Texas: Journal of Paleontology, v. 68, p. 1131–1145.

Albright, L. B., 1996, Rodents and carnivores of the Toledo Bend Local Fauna: An Arikareean (earliest Miocene) assemblage from the Texas Coastal Plain: Journal of Vertebrate Paleontology, v. 16, p. 458–473.

Barnes, V. E., 1967, Palestine sheet: University of Texas, Bureau of Economic Geology, Geologic Atlas of Texas, scale 1:250,000.

Barnes, V. E., 1974, Austin sheet: University of Texas, Bureau of Economic Geology, Geologic Atlas of Texas, scale 1:250,000.

Berggren, W. A., Kent, D. V., Swisher, C. C., and Aubry, M. P., 1995, A revised Cenozoic geochronology and chronostratigraphy, *in* Berggren, W. A., Kent, D. V., Aubry, M. P., and Hardenbol, J., eds., Geochronology, time scales, and global stratigraphic correlation: Tulsa, Society of Economic Mineralogists and Paleontologists Special Publication 54, p. 129–212.

Fisk, H. N., 1940, Geology of Avoyelles and Rapides Parishes: Geological Bulletin of the Louisiana Geological Survey, v. 18, 240 p.

Forsten, A., 1975a, The fossil horses from the Texas Gulf Coastal Plain: Journal of Paleontology, v. 49, p. 395–399.

Forsten, A., 1975b, The fossil horses of the Texas Gulf Coastal Plain: A revision. Pearce-Sellards Series, no. 22, 86 p.

Frailey, C. D., 1978, An early Miocene (Arikareean) fauna from northcentral Florida (the SB-1A local fauna): Occasional Papers of the Museum of Natural History, University of Kansas, no. 75, p. 1–20.

Frailey, C. D., 1979, The large mammals of the Buda local fauna (Arikareean:

Alachua County, Florida): Bulletin of the Florida State Museum, Biological Sciences, v. 24, p. 123–173.

Galloway, W. E., Jirik, L. A., Morton, R. A., and DuBar, J. R., 1986, Lower Miocene (Fleming) depositional episode of the Texas Coastal Plain and Continental Shelf: Structural framework, facies, and hydrocarbon resources: University of Texas, Bureau of Economic Geology, Report of Investigations 150, 44 p.

Galloway, W. E., Bebout, D. G., Fisher, W. L., Dunlap, J. B., Cabrera-Castro, R., Lugo-Rivera, J. E., and Scott, T. M., 1991, Cenozoic, *in* Salvador, A., ed., The Gulf of Mexico Basin: Boulder, Colorado, Geological Society of America, Geology of North America, v. J, p. 245–324.

Glowacz, M. E., and Horne, J. C., 1971, Depositional environments of the early Miocene as exposed in the Cane River Diversion Canal, Rapides Parish, Louisiana: Transactions of the Gulf Coast Association of Geological Societies, v. 21, p. 379–386.

Hay, O. P., 1924, Description of some fossil vertebrates from the upper Miocene of Texas: Proceedings of the Biological Society of Washington, v. 37, 20 p.

Hesse, C. J., 1943, A preliminary report on the Miocene vertebrate faunas of southeast Texas: Proceedings and Transactions of the Texas Academy of Science, v. 26, p. 157–179.

Hunt, R. M., 1990, Taphonomy and sedimentology of Arikaree (lower Miocene) fluvial, eolian, and lacustrine paleoenvironments, Nebraska and Wyoming; a paleobiota entombed in fine-grained volcaniclastic rocks, *in* Lockley, M. G., and Rice, A., eds., Volcanism and fossil biotas: Geological Society of America Special Paper 244, p. 69–112.

Jones, D. S., Mueller, P. A., Hodell, D. A., and Stanley, L. A., 1993, $^{87}Sr/^{86}Sr$ geochronology of Oligocene and Miocene marine strata in Florida, *in* Zullo, V. A., Harris, W. B., Scott, T. M., and Portell, R. W., eds., The Neogene of Florida and adjacent regions; Proceedings. 3rd Bald Head Island Conference on Coastal Plains Geology: Florida Geological Survey Special Publication 37, p. 15–26.

Korth, W. W., 1996, A new genus of beaver (Mammalia: Castoridae: Rodentia) from the Arikareean (Oligocene) of Montana and its bearing on castorid phylogeny: Carnegie Museum Annals, v. 65, p. 167–179.

Lillegraven, J. A., 1981, Evolutionary relationships of middle Eocene and younger species of *Centetodon* (Mammalia, Insectivora, Geolabididae) with a description of *Ankylodon* (Adapisoricidae): University of Wyoming Publications, v. 45, 115 p.

MacFadden, B. J., 1980, An early Miocene land mammal (Oreodonta) from a marine limestone in northern Florida: Journal of Paleontology, v. 54, p. 93–101.

MacFadden, B. J., and Webb, S. D., 1982, The succession of Miocene (Arikareean through Hemphillian) terrestrial mammalian localities and faunas in Florida, *in* Scott, T. M., and Upchurch, S. B., eds., Miocene of the southeastern United States: Florida Department of Natural Resources, Bureau of Geology Special Publication 25, p. 186–199.

Manning, E., 1990, The late early Miocene Sabine River: Transactions of the Gulf Coast Association of Geological Societies, v. 40, p. 531–549.

Morgan, G. S., 1989, Miocene vertebrate faunas from the Suwannee River Basin of north Florida and south Georgia, *in* Morgan, G. S., ed., Miocene paleontology and stratigraphy of the Suwannee River Basin of north Florida and south Georgia; Guidebook 30: Tallahassee, Southeastern Geological Society, p. 26–53.

Morgan, G. S., 1993, Mammalian biochronology and marine-nonmarine correlations in the Neogene of Florida: Florida Geological Survey Special Publication 37, p. 55–66.

Morgan, G. S., 1994, Miocene and Pliocene marine mammal faunas from the Bone Valley Formation of central Florida, *in* Berta, A., and Deméré, T. A., eds., Contributions in Marine Mammal Paleontology Honoring Frank C. Whitmore, Jr.: Proceedings of the San Diego Society of Natural History, no. 29, p. 239–268.

Patton, T. H., 1967, Oligocene and Miocene vertebrates from central Florida, *in* Brooks, H. K., and Underwood, J. R., eds., Miocene-Pliocene problems of peninsular Florida; 13th Field Trip: Tallahassee, Southeastern Geological Society, p. 3–10.

Patton, T. H., 1969a, Miocene and Pliocene artiodactyls, Texas Gulf Coastal Plain: Bulletin of the Florida State Museum, v. 14, p. 115–226.

Patton, T. H., 1969b, An Oligocene land vertebrate fauna from Florida: Journal of Paleontology, v. 43, p. 543–546.

Patton, T. H., and Taylor, B. E., 1971, The Synthetoceratinae (Mammalia, Tylopoda, Protoceratidae): Bulletin of the American Museum of Natural History, v. 145, p. 119–218.

Patton, T. H., and Taylor, B. E., 1973, The Protoceratinae (Mammalia, Tylopoda, Protoceratidae) and the systematics of the Protoceratidae: Bulletin of the American Museum of Natural History, v. 150, p. 347–414.

Prothero, D. R., 1994, The Eocene-Oligocene transition: paradise lost: New York, Columbia University Press, 291 p.

Prothero, D. R. , and Emry, R. J., 1996, Summary, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: New York, Cambridge University Press, p. 664–683.

Prothero, D. R., and Manning, E., 1987, Miocene rhinoceroses from the Texas Gulf Coastal Plain: Journal of Paleontology, v. 61, p. 388–423.

Prothero, D. R., and Sereno, P. C., 1982, Allometry and paleoecology of medial Miocene dwarf rhinoceroses from the Texas Gulf Coastal Plain: Paleobiology, v. 8, p. 16–30.

Quinn, J. H., 1952, Recognition of hipparions and other horses in the middle Miocene mammalian faunas of the Texas Gulf region: University of Texas, Bureau of Economic Geology Report of Investigations 14, p. 5–6.

Quinn, J. H., 1955, Miocene Equidae of the Texas Gulf Coastal Plain. University of Texas Bureau of Economic Geology Publications 5516, p. 1–90.

Retallack, G. J., 1992, Paleosols and changes in climate and vegetation across the Eocene/Oligocene boundary, *in* Prothero, D. R., and Berggren, W. A., eds., Eocene-Oligocene climatic and biotic evolution: Princeton, New Jersey, Princeton University Press, p. 382–398.

Rich, T. H. J., and Patton, T. H., 1975, First record of a fossil hedgehog from Florida (Erinaceidae, Mammalia): Journal of Mammalogy, v. 56, p. 692–692.

Rogers, J. E., and Calandro, A. J., 1965, Water resources of Vernon Parish, Louisiana: Louisiana Department of Conservation Geological Survey and Department of Public Works Water Resources Bulletin 6, 104 p.

Schoch, R. M., 1984, The type specimens of *Tapirus validus* and ?*Tapiravus rarus* (Mammalia, Perissodactyla), with a review of the genus, and a new report of *Miotapirus* (*Miotapirus marslandensis* Schoch and Prins new species) from Nebraska: Postilla, no. 195, p. 1–11.

Simpson, G. G., 1930, Tertiary land mammals of Florida: Bulletin of the American Museum of Natural History, v. 59, p. 149–211.

Slaughter, B. H., 1981, A new genus of geomyoid rodent from the Miocene of Texas and Panama: Journal of Vertebrate Paleontology, v. 1, p. 111–115.

Snead, J. I., and McCulloh, R. P., compilers, 1984, Geologic map of Louisiana, 1984: Baton Rouge, Louisiana Geological Survey, scale 1:500,000.

Spradlin, S. D., 1980, Miocene fluvial systems: southeast Texas [M.S. thesis]: Austin, University of Texas 139 p.

Stevens, M. S., 1977, Further study of Castolon Local Fauna (early Miocene) Big Bend National Park, Texas: Pearce-Sellards Series, no. 28, p. 1–69.

Stevens, M. S., Stevens, J. B., and Dawson, M. R., 1969, New early Miocene formation and vertebrate local fauna, Big Bend National Park, Brewster County, Texas: Pearce-Sellards Series 15, p. 1–53.

Tedford, R. H., and Hunter, M. E., 1984, Miocene marine-nonmarine correlations, Atlantic and Gulf Coastal Plains, North America: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 47, p. 129–151.

Tedford, R. H., Skinner, M. F., Fields, R. W., Rensberger, J. M., Whistler, D. P., Galusha, T., Taylor, B. E., Macdonald, J. R., and Webb, S. D., 1987, Faunal succession and biochronology of the Arikareean through Hemphillian interval (late Oligocene through earliest Pliocene epochs), in North America, *in* Woodburne, M. O., ed., Cenozoic mammals of North America, geochronology and biostratigraphy: Berkeley, University of California Press, p. 153–210.

Tedford, R. H., Swinehart, J. B., Swisher, C. C., Prothero, D. R., King, S. A., and Tierney, T. E., 1996, The Whitneyan-Arikareean transition in the High Plains, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: New York, Cambridge University

Press, p. 312–334.

Veatch, A. C., 1906, Geology and underground water resources of northern Louisiana and southern Arkansas: U.S. Geological Survey Professional Paper 46, 422 p.

Webb, S. D., 1977, A history of savanna vertebrates in the New World. Part 1: North America: Annual Review of Ecology and Systematics, v. 8, p. 355–380.

Webb, S. D., 1989, The fourth dimension in North American terrestrial mammal communities, *in* Morris, D. W., Abramsky, Z., Rox, B. J., and Willig, M. R., eds., Patterns in the structure of mammalian communities: Texas Technical University Special Publication 28, p. 181–203.

Webb, S. D., Hulbert, R. C., and Lambert, W. D., 1995, Climatic implications of large-herbivore distributions in the Miocene of North America, *in* Vrba, E. S., Denton, G. H., Partridge, T. C., and Burckle, L. H., eds., Paleoclimate and evolution: with emphasis on human origins: New Haven, Yale University Press, p. 91–108.

Welch, R. N., 1942, Geology of Vernon Parish: Louisiana Department of Conservation Geological Bulletin 22, 90 p.

Westgate, J. W., 1992, *Dinohyus* aff. *D. hollandi* (Mammalia, Entelodontidae) in Alabama: Journal of Paleontology, v. 66, p. 685–687.

Westgate, J. W., 1993, A giant entelodont from Oligocene deposits of the Texas Coastal Plain: Journal of Vertebrate Paleontology Abstracts of Papers, v. 13, p. 62A.

Whitmore, F. C., and Stewart, R. H., 1965, Miocene mammals and Central American seaways: Science, v. 148, p. 180–185.

Wilson, J. A., 1956, Miocene formations and vertebrate biostratigraphic units, Texas Coastal Plain: Bulletin of the American Association of Petroleum Geologists, v. 40, p. 2233–2246.

Wilson, J. A., 1957, Early Miocene entelodonts, Texas Coastal Plain: American Journal of Science, v. 255, p. 641–649.

Wilson, J. A., 1959, Stratigraphic concepts in vertebrate paleontology: American Journal of Science, v. 257, p. 770–778.

Wilson, J. A., 1960, Miocene carnivores, Texas Coastal Plain: Journal of Paleontology, v. 34, p. 983–1000.

Wilson, J. A., 1968, Tertiary shorelines, Texas Coastal Plain: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 5, p. 135–140.

Wood, H. E., and Wood, A. E., 1937, Mid-Tertiary vertebrates from the Texas Coastal Plain; fact and fable: American Midland Naturalist, v. 18, p. 129–146.

Wood, H. E., Chaney, R. W., Clark, J., Colbert, E. H., Jepsen, G. L., Reeside, J. B., and Stock, C., 1941, Nomenclature and correlation of the North American continental Tertiary: Geological Society of America Bulletin, v. 52, p. 1–48.

Woodburne, M. O., 1987, Systematic index, *in* Woodburne, M. O., ed., Cenozoic mammals of North America, geochronology and biostratigraphy: Berkeley, University of California Press, p. 293–308.

Woodburne, M. O., and Swisher, C. C., 1995, Land mammal high-resolution geochronology, intercontinental overland dispersals, sea level, climate, and vicariance, *in* Berggren, W. A., Kent, D. V., Aubry, M. P., and Hardenbol, J., eds., Geochronology, time scales, and global stratigraphic correlation: Tulsa, Society of Economic Mineralogists and Paleontologists Special Publication 54, p. 335–364.

Xu, Xiaofeng, 1994, Evolution of Chinese Castoridae, *in* Tomida, Y., Chuankuei, L., and Setoguchi, T., eds., Rodent and Lagomorph families of Asian origins and diversification, Proceedings of Workshop WC-2, 29th International Geological Congress, Kyoto, Japan: Tokyo, National Science Museum, p. 77–98.

Xu, Xiaofeng, 1996, Castoridae, *in* Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: New York, Cambridge University Press, p. 417–432.

MANUSCRIPT ACCEPTED BY THE SOCIETY DECEMBER 5, 1997

Printed in U.S.A.

Geological Society of America
Special Paper 325
1998

Lithostratigraphy, paleontology, and biochronology of the Chadron, Brule, and Arikaree Formations in North Dakota

John W. Hoganson and Edward C. Murphy
North Dakota Geological Survey, 600 East Boulevard Avenue, Bismarck, North Dakota 58505-0840
Nels F. Forsman
Department of Geology and Geological Engineering, University of North Dakota, Grand Forks, North Dakota 58202-8358

ABSTRACT

Exposed strata of the White River Group (Chadron and Brule Formations) and Arikaree Formation in North Dakota occur as sparse and isolated erosional remnants only in southwestern parts of the state, primarily in the Chalky Buttes, Little Badlands, and Killdeer Mountains areas.

The Chadron Formation in North Dakota has been divided into two members. The basal Chalky Buttes Member is lithologically variable and consists of yellow-green to white, gravel-bearing, cross-bedded sandstones and sandy mudstones. This member unconformably overlies strata of the Fort Union Group and conformably underlies the South Heart Member of the Chadron Formation. The South Heart Member consists of green to gray smectitic claystones containing lenticular beds of freshwater limestones and conformably underlies the Brule Formation. Mammalian fossils of Chadronian age have been recovered from both members.

The Brule Formation is composed of pinkish-brown to gray-green complexly interbedded claystones, calcareous mudstones, siltstones, freshwater limestones, tuffaceous beds, and cross-bedded sandstones. The Antelope Creek tuff in the lower part of the formation is a useful marker bed for correlation between isolated outcrops. The Brule Formation is unconformably overlain by the Arikaree Formation, and the Brule is easily distinguishable lithologically, and by color, from the Chadron and Arikaree Formations. Mammalian remains indicate an Orellan age for the lower part of the formation and a possible Whitneyan age for the upper part.

The Arikaree Formation consists of concretionary, cross-bedded, calcareous sandstones, siltstones, silty claystones, tuffaceous beds, and carbonates, and caps several isolated buttes in southwestern North Dakota. A basal conglomerate commonly marks the disconformity that separates the Arikaree rocks from the White River Group. The Arikaree Formation is overlain by a thin veneer of unconsolidated lag sediments, mostly gravel, of unknown age. Vertebrate fossils are extremely sparse in the Arikaree Formation in North Dakota but tend to suggest a late Whitneyan to late Arikareean age. This is consistent with an age of 25.1 ± 2.2 Ma from fission track dating of volcanic glass from the middle part of the Arikaree Formation in the Killdeer Mountains.

Hoganson, J. W., Murphy, E. C., and Forsman, N. F., 1998, Lithostratigraphy, paleontology, and biochronology of the Chadron, Brule, and Arikaree Formations in North Dakota, *in* Terry, D. O., Jr., LaGarry, H. E., and Hunt, R. M., Jr., eds., Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America): Boulder, Colorado, Geological Society of America Special Paper 325.

INTRODUCTION

This investigation was initiated to determine the lithostratigraphic and biostratigraphic relationships of the Chadron, Brule, and Arikaree Formations in North Dakota to identify paleofaunas restricted to these formations, to determine the age of the strata using radiometric and biochronologic indicators, and to attempt to identify similarities and differences of these formations in North Dakota compared to those in other areas of the midcontinent. This 5-yr study began as a Cooperative Geological Mapping Program (COGEOMAP) between the U.S. Geological Survey and the North Dakota Geological Survey. Results, published in detail as a North Dakota Geological Survey Report of Investigation (Murphy et al., 1993), are summarized and updated here.

Exposures of the Chadron, Brule, and Arikaree Formations are sparse and scattered in North Dakota in isolated buttes and butte complexes in the southwestern part of the state. Principal outcrops exist in the areas of Chalky Buttes (Slope County), Little Badlands (Stark County), Rainy Buttes (Slope County), and the Killdeer Mountains (Dunn County) (Fig. 1). These units have been mapped at only 27 butte localities in southwestern North Dakota, a very small area compared to exposures of correlatives in other areas of the midcontinent (Fig. 1). Commonly, only the caprock is exposed on these buttes and 11 of the buttes have an areal extent of 10 acres or less. Even though exposures of these units are restricted, they are important because they are the most northeasterly occurrences of these strata in North America.

PREVIOUS INVESTIGATIONS

Cope (1883a) was the first to introduce White River terminology to North Dakota for outcrops in the Chalky Buttes and presented the first list of vertebrate taxa from White River deposits in North Dakota (Fig. 2). He also described two new species of fish collected by Charles White from Chadron Formation lacustrine deposits capping Sentinel Butte (Cope, 1883b). The Carnegie Museum sent a geologic reconnaissance expedition led by Earl Douglass to North Dakota and other western states in 1905. In his interesting account of this expedition, Douglass (1909) identified rock exposures in the Little Badlands as White River strata and reported the occurrence of many vertebrate taxa from White River beds in the Little Badlands and Chalky Buttes. He described, in detail, the rhinoceros (Douglass, 1908a) and horse (Douglass, 1908b) fossils collected during the expedition. Outcrops in the Killdeer Mountains were first assigned to the "White River formation" by Quirke (1913, 1918).

The first systematic investigation of White River rocks in southwestern North Dakota was by Leonard (1922). He divided the White River formation into three informal units: "Lower White River beds," "Middle White River (*Oreodon*) beds," and "Upper White River beds." Several fossils of "typical White River Oligocene mammals" were collected from the Little Badlands during his study (Leonard, 1922, p. 220). Powers (1945) also recognized a three-fold division of White River strata in North Dakota but termed these units, in ascending order, "*Titanotherium* beds," "*Oreodon* beds," and "*Protoceras* beds." Denson et al. (1959) elevated the White River Formation in North Dakota to group status, and Moore et al. (1959) identified the rocks as the Chadron and Brule Formations (Fig. 2). Moore et al. (1959) were also first to apply the term Arikaree Formation to the 21 m of capping sandstone at Chalky Buttes, although Seager et al. (1942) had suggested that Miocene rocks may overlie the White River beds in North Dakota.

During the 1940s to 1960s several investigations were carried out in southwestern North Dakota to evaluate the economic potential (uranium, claystone, and limestone) of the White River beds. The most comprehensive of these studies was by Denson and Gill (1965), who examined the uranium potential of White River strata. They measured several sections, produced a cross section that for the first time correlated these units among several buttes, described the lithology and mineralogy of the formations, and noted the biostratigraphy of important vertebrate fossils. In addition, they were first to map the tuffaceous-marlstone cap rocks of the Killdeer Mountains as the Arikaree Formation, and suggested that these strata are Miocene in age.

Two important studies of the White River Group and Arikaree Formation were conducted by doctoral students from the Geology Department at the University of North Dakota. Stone (1973) measured sections in the Chalky Buttes, Little Badlands, and Killdeer Mountains, and correlated the units in a geologic cross section. He split the Chadron Formation into three members, and the Brule Formation into two members. Even though these member names were used by subsequent workers, Hoganson (1986) noted that Stone's members were not formally proposed in accordance with the rules established by the North American Stratigraphic Code, and thus the terms should be considered informal stratigraphic names. Hoganson and Lammers (1992) suggested that Stone's informal Brule members were of questionable utility. Stone (1973) also informally replaced the term Arikaree Formation with the name "Killdeer formation," because he believed that the cap rocks on many of the North Dakota buttes were lithologically different than typical Arikaree lithologies in South Dakota and Nebraska. Stone found fossils weathering out of the "Killdeer formation" at White Butte and tentatively identified them as the hypertragulid *Hypertragulus minor* and the rhinocerotid *Amphicaenopus*. He cited those taxa and the 1958 collection of the skull of *Palaeocastor* by Jean Hough, University of Chicago, from the "Killdeer formation" at Golden Butte (Obritsch Ranch) as evidence to suggest similarity to the Arikaree Group of Nebraska and an Arikareean age for the "Killdeer formation." Stone's fossils could not be located for restudy.

Delimata (1975) conducted a detailed stratigraphic and petrographic study of the Arikaree Formation in the Killdeer Mountains for his dissertation. One of Delimata's primary objectives was to determine the age of the cap rocks on the Killdeer Mountains by finding diagnostic fossils. He was unsuccessful in that attempt but determined that the formation consists mostly of highly altered, tuffaceous carbonates in the Killdeer Mountains.

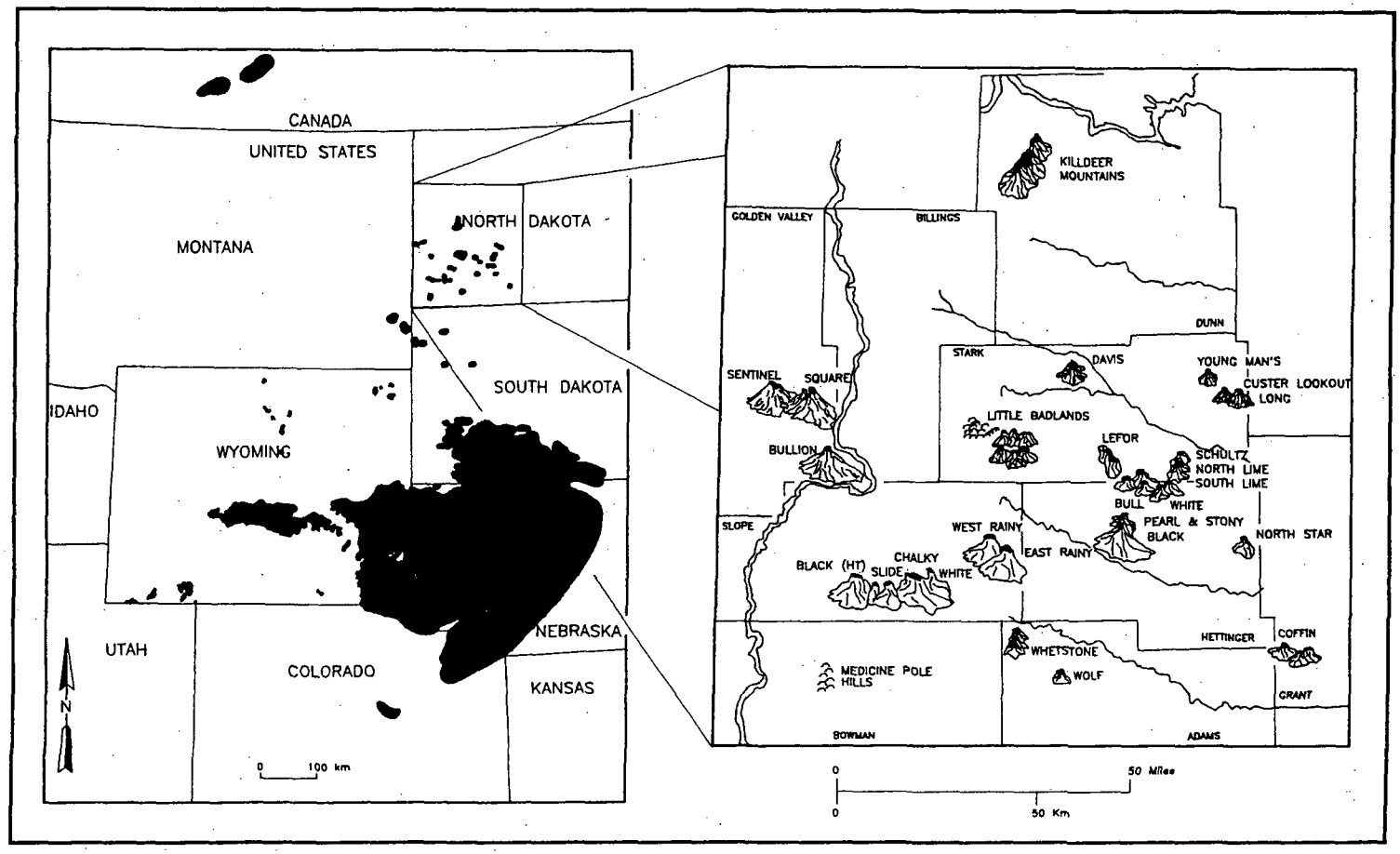


Figure 1. Areas of White River Group and Arikaree Formation exposures in the northern mid-continent (modified from Seeland, 1985; Lillegraven, 1993). Detail map shows buttes and butte complexes where White River and Arikaree strata are exposed in southwestern North Dakota.

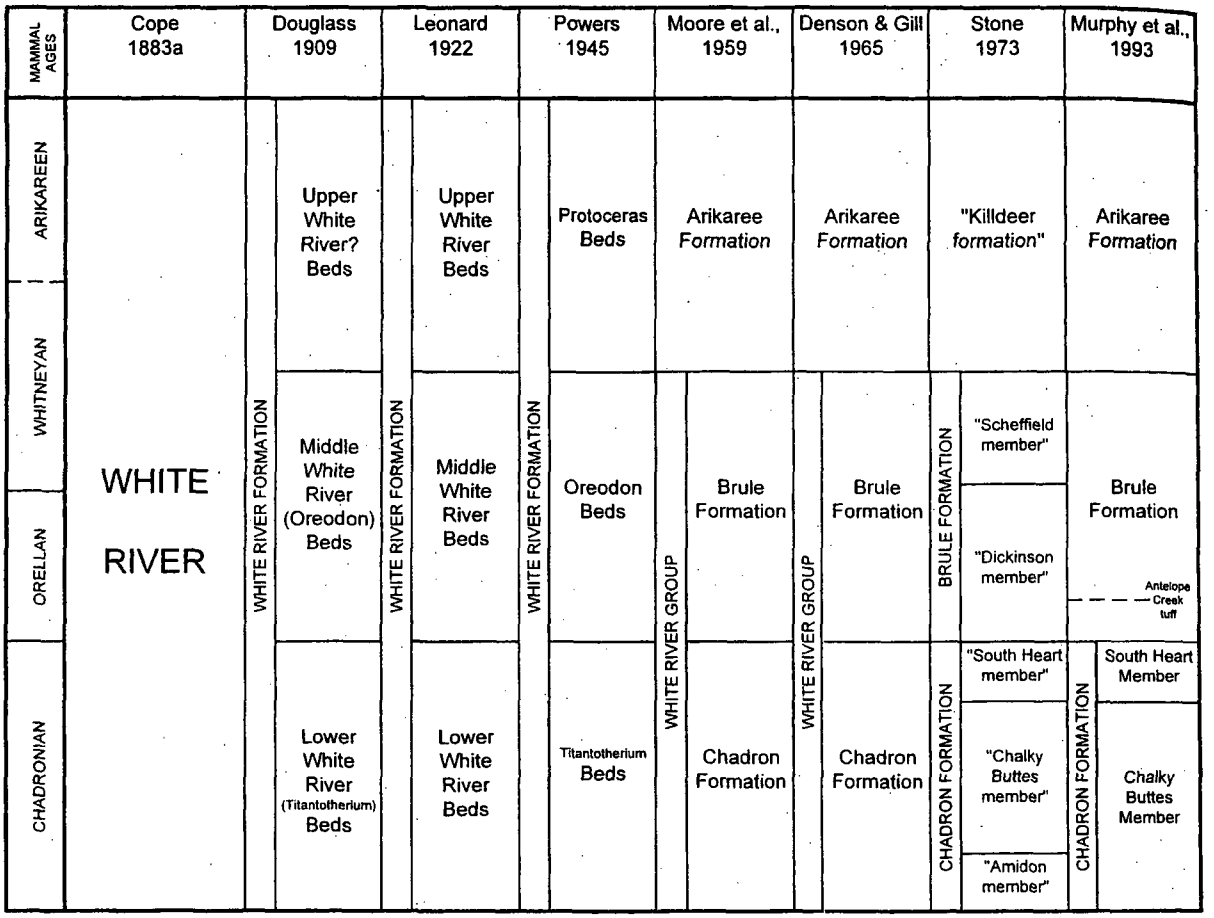


Figure 2. History of White River Group and Arikaree Formation stratigraphic nomenclature in North Dakota.

Delimata's numerous, detailed measured sections have remained a useful source for stratigraphic information.

Most previous investigations of White River and Arikaree strata in North Dakota have included reports of isolated fossil occurrences. The vertebrate fossil record of the Brule Formation in North Dakota was outlined by Hoganson and Lammers (1992). All known vertebrate, invertebrate, and plant taxa from the North Dakota Brule Formation were listed in their report. They established an Orellan age for the lower Brule Formation in North Dakota but could not determine the age of the upper Brule because of the sparsity of fossils in that part of the section. The most comprehensive discussion of the paleontology of the North Dakota Chadron, Brule, and Arikaree Formations was in a report by us (Murphy et al., 1993); a complete listing of fossils from the formations was given. Also included in that study were detailed descriptions of 31 measured sections and discussions of the petrology, lithostratigraphy, biochronology, and radiometric ages of the strata. A possible Whitneyan age was proposed for the upper Brule Formation in North Dakota and a late Whitneyan to late Arikareean age was confirmed for the Arikaree Formation in North Dakota in that study. Two members of the Chadron Formation, the Chalky Buttes Member and the overlying South Heart Member, were formally named in that report.

METHODS

Field activities included measuring geologic sections, coring in areas of poor or nonexistent outcrops, collection of lithologic samples for petrographic characterization and potential radiometric dating, collection of paleontologic specimens for biochronologic and paleobiogeographic analyses, and field reconnaissance of the geologic type-sections of the Chadron, Brule, and Arikaree Formations in South Dakota and Nebraska for comparison to the presumed equivalent North Dakota formations.

Laboratory procedures included hand specimen and thin section petrographic analysis, x-ray analysis of claystones, fission-track analysis of glass shards, trace-element analysis of glass separates, and fossil preparation and identification.

CHADRON FORMATION

Lithostratigraphy

Chalky Buttes Member. As in most places in the upper Midwest, the Chadron Formation in North Dakota is the basal formation in the White River Group. The Chadron Formation is divided into two easily recognizable members in North Dakota, the Chalky Buttes Member and the overlying South Heart Member (Fig. 3). The Chalky Buttes Member consists of light colored, generally white to yellow-green, gravel-bearing, cross-bedded sandstone and sandy mudstone. Pebbles and cobbles of volcanic porphyries, apparently unique to this member, and other exotic rocks are oriented along bedding planes in some of the better exposures (Fig. 4). At a few localities, a 4.5-m-thick white to purple mudstone is found at the base of the member. The Chalky Buttes Member is, on the average, 6 m thick in North Dakota, and reaches a

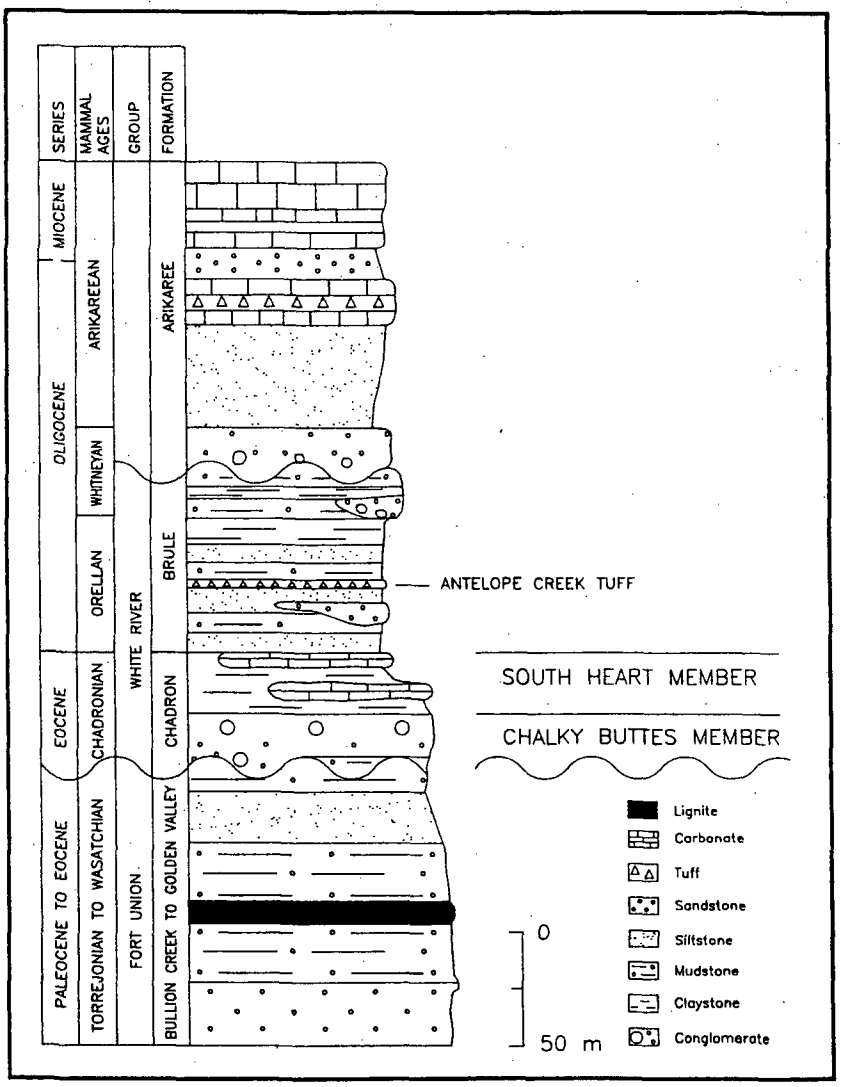


Figure 3. Generalized stratigraphic column for the Chadron, Brule, and Arikaree Formations in North Dakota.

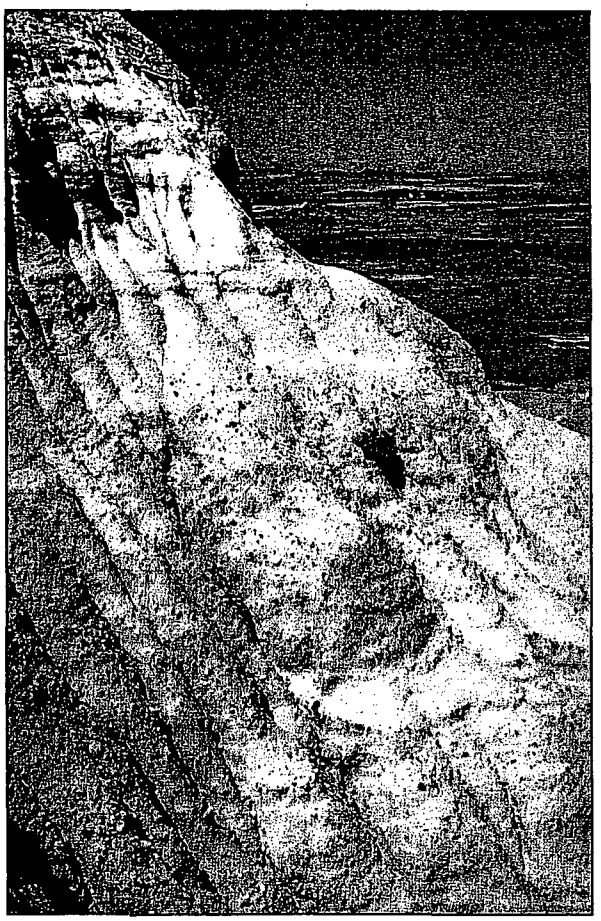


Figure 4. White to light gray conglomerate and sandy mudstone (22 m thick) of the Chalky Buttes Member (Chadron Formation) at the member's type section in the Chalky Buttes (NW1/4, NE1/4, NE1/4, sec. 25, T. 134 N., R. 101 W.), Slope County. View is to the north.

maximum thickness of 24 m in the Chalky Buttes. Measured sections of this member were given by Murphy et al. (1993).

The Chalky Buttes Member rests unconformably on the Bullion Creek (Paleocene), Sentinel Butte (Paleocene), or Golden Valley Formations (Paleocene and Eocene) of the Fort Union Group. The contact is sharp between the overlying brightly colored white to yellow-green gravel-bearing, cross-bedded sandstone or sandy mudstone of the Chalky Buttes Member and usually somber gray to brown beds of alternating sandstone, siltstone, mudstone, claystone, and lignite of the Fort Union Group. In Slope County, North Dakota, at the type locality in the Chalky Buttes, the basal contact of this member is placed above a brightly colored 0.6- to 0.9-m-thick analcime bed (paleosol) in the Sentinel Butte Formation. The Chalky Buttes Member is conformably overlain by the South Heart Member of the Chadron Formation.

The Chalky Buttes Member is lithologically similar to the Chamberlain Pass Formation in the Big Badlands of South Dakota and Nebraska (Evans and Terry, 1994; Terry et al., 1995), and Lillegraven's (1970) "dazzling white" unit of the Chadron Formation at Slim Buttes, South Dakota (Fig. 5).

South Heart Member. The South Heart Member, the most distinctive and continuous lithostratigraphic unit in the White River Group in North Dakota, is the upper member of the Chadron Formation (Fig. 3). The rocks consist of brown to gray-green smectitic claystone that commonly contains thin (<1.5 m thick) lenticular beds of freshwater limestone, particularly in the upper part of the member (Fig. 6). These limestone beds are often ledge-formers and cap some of the small buttes in the study area (Fig. 7). Radially fibrous calcite concretions are often found on South Heart outcrops. South Heart claystones have "popcorn weathering" surfaces and drape over underlying strata, thus making it difficult to locate contacts. The rocks weather to distinctive grayish domed "haystack" hills and butte caps in areas of the Little Badlands and Chalky Buttes. The South Heart Member attains a maximum thickness of 29 m at White Butte (Slope County), and ranges from about 3–18 m elsewhere in the study area. Measured sections of the South Heart Member in southwestern North Dakota were given by Murphy et al. (1993).

Even though the South Heart Member is conformably underlain by the Chalky Buttes Member of the Chadron Formation, and is conformably overlain by the Brule Formation, it is easily distinguishable from those units lithologically, by color, and by weathering characteristics. The rounded exposures and popcorn weathered appearing, somber grayish-colored claystone of the South Heart Member stands in sharp contrast to the dazzling white of the underlying gravelly sandstones of the Chalky Buttes Member and the pink and brown siltstones and mudstones of the overlying Brule Formation, both of which weather to vertical cliffs. The South Heart Member is an excellent "marker" horizon in southwestern North Dakota.

The South Heart Member is lithologically similar to the Peanut Peak Member of the Chadron Formation in the Big Badlands of South Dakota and Nebraska (Evans and Terry, 1994; Terry et al., 1995) and to Lillegraven's (1970) "typical Chadron" at Slim Buttes, South Dakota (Fig.5).

Paleontology

Vertebrate fossils have been recovered from six localities in the Chadron Formation of North Dakota. Brontothere remains have been found in both members of the Chadron Formation (Murphy et al., 1993). They are the only mammalian fossils recorded from the South Heart Member. The only known diverse vertebrate fossil assemblage in the Chadron Formation is from the Medicine Pole Hills area of Bowman County. At that locality, erosional remnants of the Chalky Buttes Member are exposed in isolated outcrops and blowouts. Brontothere remains were first recovered from this locality by Leonard (1922), and then by other early workers (Hares, 1928; Benson,

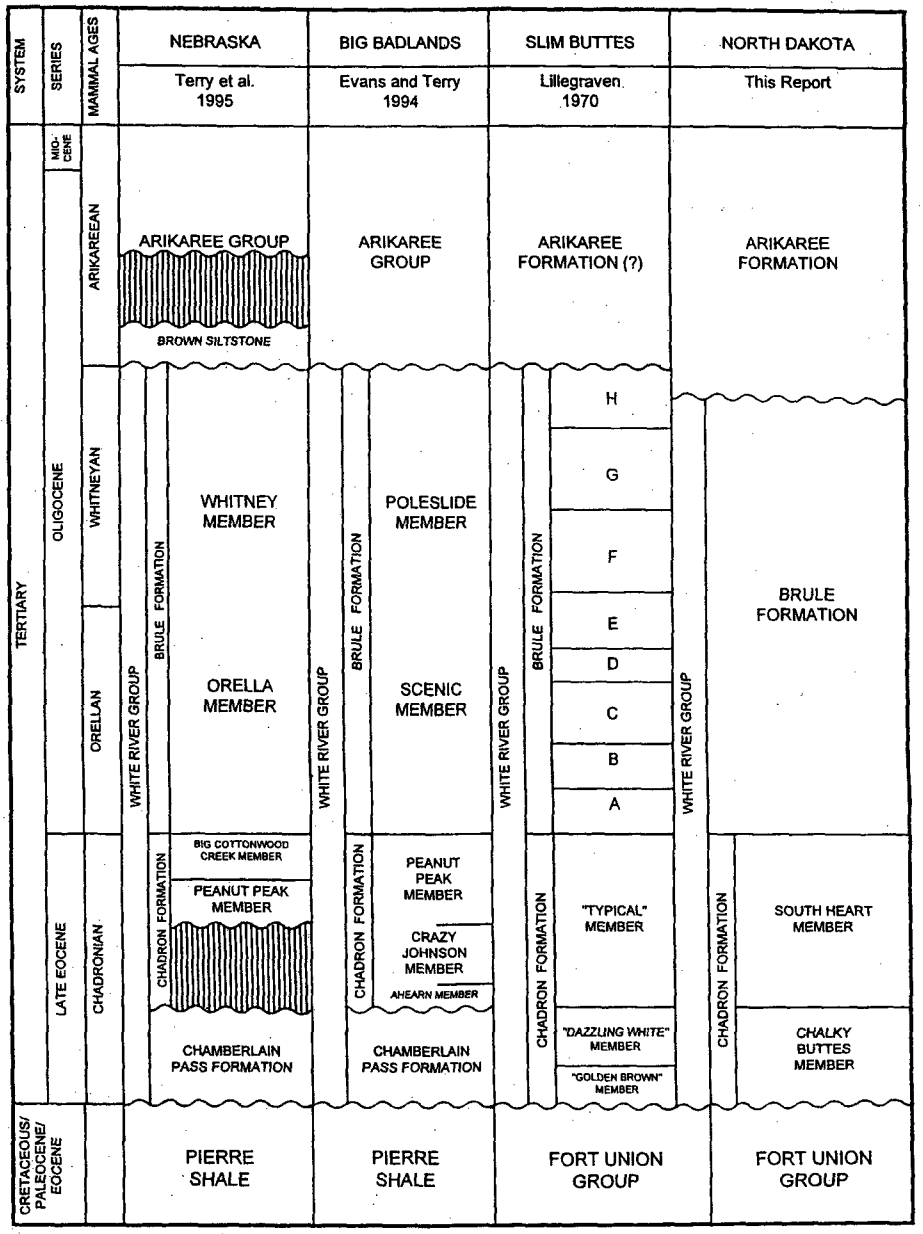


Figure 5. Proposed correlation of the White River Group and Arikaree Formation in North Dakota with lithostratigraphic units in South Dakota and Nebraska.

1952; Denson et al., 1959; Kepferle and Culbertson, 1955). Pearson (1993) determined that a diverse vertebrate fossil assemblage is present at this site, and called the fauna the Medicine Pole Hills local fauna. Remains of 31 mammalian (within 11 orders and 22 families), 5 fish, 2 amphibian, and 10 reptilian taxa have been identified from this locality; it is anticipated that additional taxa will be found as work continues (Pearson and Hoganson, 1995a,b).

Ostracodes, algal crusts, gastropods, and fish remains have been found in freshwater limestones in the South Heart Member of the Chadron Formation at several localities (Murphy et al., 1993). No leaf fossils have been reported from the Chadron Formation in North Dakota.

Figure 6. Brown to dark gray smectitic claystone (upper 9.5 m in photo) of the South Heart Member (Chadron Formation) at the member's type section, Fitterer Ranch (NW1/4, SW1/4, NE1/4, sec. 7, T. 137 N., R. 97 W.), Stark County. Note draping of the South Heart Member claystone. Arrow points to contact between the South Heart and underlying Chalky Buttes Members of the Chadron Formation. View is to the northeast.

Figure 7. White freshwater limestone (1 m thick) of the South Heart Member (Chadron Formation) capping Bull Butte (sec. 18, T. 136 N., R. 93 W.), Hettinger County. View is to the north.

Biochronology

Murphy et al. (1993) suggested that brontothere remains in both the Chalky Buttes Member and South Heart Member of the Chadron Formation imply a Chadronian age for the formation. The Chadronian equid, *Miohippus assiniboiensis*, reported by Prothero and Shubin (1989) from the Chadron Formation in the Little Badlands, Stark County, also suggests a Chadronian age for the Chadron Formation in North Dakota. The occurrence of *Herpetotherium valens, Sinclairella, Centetodon chadronensis, Stibarus montanus, Trigonias*, brontotheres, and possibly other taxa listed by Pearson and Hoganson (1995a) from the Medicine Pole Hills local fauna also imply a Chadronian age. Pearson and Hoganson (1995a) suggested that the Medicine Pole Hills local fauna may have affinities with the Calf Creek local fauna of the Cypress Hills Formation in Saskatchewan, believed to be early Chadronian in age (Storer, 1978). The recognition by Heaton and Emry (1996) that leptomerycids from the Medicine Pole Hills local fauna may represent an early evolutionary grade of *Leptomeryx yoderi*, or perhaps a new species, also suggests an early Chadronian age for the Chadron Formation at the Medicine Pole Hills locality.

BRULE FORMATION

Lithostratigraphy

The lower part of the Brule Formation in North Dakota consists of pinkish-brown to gray-green complexly interbedded claystones, calcareous mudstones, siltstones, freshwater limestones, tuffaceous beds, and cross-bedded sandstones (Figs. 3, 8). The channel sandstones, such as the "Fitterer bed," are generally less than 3 m thick, are difficult to trace for long distances, and usually bear mammalian fossils. A useful marker bed in the lower part of the formation was called the Antelope Creek tuff by Murphy et al. (1993) (Fig. 8). The upper part of the formation is lithologically less variable, and consists primarily of alternating beds of mudstones and thin beds of swelling claystones. In parts of the Little Badlands and at Rainy Buttes, cross-bedded conglomeratic sandstones are in the upper part of the formation. The Brule Formation ranges in thickness from 6 m in the Chalky Buttes (Slope County) to a maximum of 65 m at the Fitterer Ranch locality, Little Badlands (Stark County). Measured sections of Brule exposures in southwestern North Dakota were presented by Murphy et al. (1993).

The Brule Formation is unconformably overlain by the Arikaree Formation, and is conformably underlain by the South Heart Member of the Chadron Formation. Brule rocks are easily distinguishable by lithology and color from the underlying South Heart Member (i.e., the pinkish-brown siltstone and mudstone of the

Figure 8. Pinkish-brown to light brown mudstones, siltstones, and fine-grained sandstones (36 m thick) of the Brule Formation at Fitterer Ranch (SW1/4, sec. 7, T. 137 N., R. 97 W.), Stark County. Arrows point to Antelope Creek tuff (1 m thick). View is to the north.

Brule compared to the gray-green claystone of the South Heart Member). A conglomerate is usually found at the base of the Arikaree Formation where it overlies the Brule. It is at times difficult to distinguish these Arikaree conglomerates from conglomeratic sandstones in upper parts of the Brule Formation. Murphy et al. (1993) placed these channel sandstones in either the Arikaree or Brule Formation, depending on occurrences of typical Brule pinkish-brown siltstones or mudstones above the conglomeratic unit.

Lithologically, the lower part of the Brule Formation in North Dakota appears similar to the Scenic Member of the Brule Formation in the Big Badlands of South Dakota, and the Orella Member of the Brule Formation in Nebraska (Fig. 5). The upper part of the Brule in North Dakota is lithologically similar to the Poleslide Member of the Brule Formation in the Big Badlands of South Dakota and the Whitney Member of the Brule Formation in Nebraska. Layering and color banding, attributed to paleosol development by Retallack (1983), however, is not as obvious in the Brule Formation in North Dakota as in the Big Badlands of South Dakota.

Paleontology

Several important vertebrate fossil sites are known in the Brule Formation in southwestern North Dakota. These include Fitterer Ranch, Obritsch Ranch, the Little Badlands (Stark County), and the Chalky Buttes, including the White Butte locality (Slope County). Most of the fossil occurrences, however, are from the lower half of the formation. The most comprehensive paleontologic treatments of the Brule Formation in North Dakota are by Hoganson and Lammers (1992) and Murphy et al. (1993). Both reports provide histories of Brule fossil research in North Dakota. Murphy et al. (1993) gave a list of all taxa identified from the North Dakota Brule. At least 90 mammalian (10 orders and 31 families), 3 fish, 2 amphibian, 8 reptilian, and 1 avian taxa have been reported from the Brule Formation in North Dakota. Freshwater and terrestrial gastropods, ostracodes, insect trace fossils, and fish and mammalian coprolites are also found in the North Dakota Brule Formation. No leaf fossils are found in the formation, but algal remains, charophyte oogonia, and hackberry endocarps are present.

Biochronology

Mammalian taxa reported from the lower Brule Formation by Hoganson and Lammers (1992) and Murphy et al. (1993) (including *Leptictis haydeni, Ischyromys typus, Eumys elegans, Mesohippus bairdi, Archaeotherium mortoni, Leptomeryx evansi, Hyaenodon horridus, H. crucians, Palaeolagus, Merycoidodon, Hesperocyon, Daphoenus,* and *Dinictis*) are generally considered characteristic of the Orellan (Emry et al., 1987), and strongly suggest at least the lower part of the Brule Formation in North Dakota was deposited during the Orellan. Skinner (1951) was the first to suggest that most of the Brule Formation in North Dakota is Orellan in age, which was confirmed by Hoganson and Lammers (1992). Magnetostratigraphic interpretation appears to corroborate an Orellan age for strata in the Little Badlands (Prothero et al., 1983).

The age of the upper Brule Formation in North Dakota has been less easy to establish because of the sparsity of mammalian fossils found in that part of the section. Wood et al. (1941) reported a fauna, possibly of Whitneyan age, from the Brule Formation at White Butte, Slope County. They noted, however, that this "White Butte local fauna" differed from other Whitneyan faunas. The stratigraphic position of the fauna has not been adequately established. Skinner (1951) also suggested that upper parts of the Brule Formation in North Dakota are Whitneyan in age, without providing much paleontologic evidence for his interpretation. Moore et al. (1959) interpreted the occurrence of *Miohippus, Leptauchenia,* and *Protoceras*? from the Brule Formation at Chalky Buttes to indicate a Whitneyan age. Although Denson and Gill (1965) also cited these fossils as evidence for a Whitneyan age, for at least part of the Brule Formation in the Chalky Buttes, they cautioned that the fossils probably were recovered from a slump block. The occurrence of the rodent *Eumys brachyodus* in the Little Badlands prompted Kihm (1990) to suggest that much of the Oligocene fauna in North Dakota is Whitneyan rather than Orellan in age. Hoganson and Lammers (1992) found no evidence of a Whitneyan age for the upper Brule Formation in their study citing the sparsity of fossils in the upper Brule as a deterrent to establishing an age for that part of the section. A possibly Whitneyan age for the upper Brule Formation at East and West Rainy Butte is suggested by meager fossil evidence (Murphy et al., 1993). Magnetostratigraphic results by Prothero et al. (1983) indicate that the upper part of the Brule Formation at Fitterer Ranch corresponds to Chron C12N, which implies a Whitneyan age. Very few fossils and no fossils of biochronologic significance have been found in the upper part of the Brule Formation at Fitterer Ranch.

ARIKAREE FORMATION

Lithostratigraphy

The Arikaree Formation in North Dakota is lithologically variable and consists of concretionary, cross-bedded, calcareous sandstones, siltstones, silty claystones, carbonates, and tuffaceous beds (Fig. 3). It unconformably overlies either the Chadron or Brule Formations in North Dakota and is overlain by a thin veneer of unconsolidated sediment, usually lag gravels of unknown provenance. The Arikaree is the butte-capping formation in several areas of southwestern North Dakota. A conglomerate is often found at the base of the formation, which allows easy differentiation except where conglomeratic channel sandstones occur at the top of the Brule Formation. The Arikaree Formation is best exposed in the Killdeer Mountains (Dunn County), where it attains a maximum thickness of 130 m (Fig. 9). Murphy et al. (1993) presented measured sections from most areas of exposure of the Arikaree Formation in southwestern North Dakota.

Figure 9. Approximately 44 m of light gray sandstone and carbonate of the Arikaree Formation exposed at Medicine Hole Plateau on the southeastern edge of South Killdeer Mountain (NE1/4, sec. 27, T. 146 N., R. 96 W.), Dunn County. View is to the northwest.

The Arikaree Formation consists of tuffaceous siltstones, sandstones, and carbonates in the Killdeer Mountains. Carbonates are also found in the Arikaree Formation at the Chalky Buttes, but in most other places siltstones and sandstones are the dominant Arikaree lithologies. The siltstones, sandstones, and carbonates are ledge-formers in the Killdeer Mountains. One of these ledges, 4.5–9 m thick, consists of interbedded highly tuffaceous sandstone, siltstone, and occasional calcite-indurated beds; it is unique because it contains abundant burrows (Fig. 10). Forsman (1986) termed this the "burrowed marker unit," terminology also adopted by Murphy et al. (1993). In addition, Forsman (1986) determined that nearly all of the lithologies in the Arikaree Formation in the Killdeer Mountains contain volcanic glass and could be characterized as slightly to highly tuffaceous.

The Arikaree Formation in North Dakota was deposited in a fluvial/lacustrine setting indicated by the cross-bedded conglomerates and sandstones generally observed in the lower part of the formation and by siltstones and carbonates that are the dominant lithologies in the upper part of the formation in the Killdeer Mountains. The siltstones are often finely laminated or exhibit oscillating ripples, and freshwater ostracodes are, at times, found entombed in the limestones. The dominance of these siltstone and carbonate lithologies in the upper part of the Arikaree Formation in the Killdeer Mountains indicates that lacustrine conditions prevailed in that part of North Dakota.

The Arikaree Formation in North Dakota is lithologically similar to the Gering and Harrison Formations of the Arikaree Group in northwestern Nebraska. Like the Arikaree Formation in North Dakota, those units are lithologically variable, but are predominantly fine-grained, tuffaceous sandstones and siltstones with interbedded carbonate lenses. The Gering and Harrison Formations in Sioux County, Nebraska, also contain small burrows at some localities similar to those in the burrowed marker unit in the Killdeer Mountains and in the Arikaree Formation at White Butte and West Rainy Butte, Slope County.

Paleontology

Fossils are extremely sparse in the Arikaree Formation of North Dakota. Remains of two genera of oreodonts (*Merychyus* and *Merycochoerus*) were collected from near the top of the Arikaree Formation in the Killdeer Mountains by the Frick Laboratory of the American Museum of Natural History in 1954. We recently collected another skull of an immature *Merychyus* from near the summit of the Killdeer Mountains. The only other fossils that have been found in the Arikaree Formation in the Killdeer Mountains are burrows and ostracodes. Elsewhere, vertebrate fossils reported from the Arikaree Formation in North Dakota include *Palaeocastor* from near the summit of the buttes at Obritsch Ranch in the Little Badlands area of Stark County (Murphy et al., 1993); *Miohippus obliquidens* from the top of East Rainy Butte, Slope County (Prothero and Shubin, 1989); *Elomeryx armatus* from the summit

Figure 10. Burrowed marker unit, a highly tuffaceous, burrowed sandstone containing calcite-indurated beds, in the Arikaree Formation, Killdeer Mountains (NW1/4,NE1/4,NE1/4 sec. 27, T. 46 N., R. 96 W.), Dunn County. Pick = 1 m.

of East Rainy Butte (Murphy et al., 1993); *Hypertragulus minor*? and *Amphicaenopus* from White Butte, Slope County (Stone, 1973); and *Leptauchenia decora* and *Nanotragulus* from White Butte, Slope County (Murphy et al. 1993).

Biochronology

The occurrence of *Elomeryx armatus, Miohippus obliquidens, Palaeocastor, Nanotragulus,* and *Leptauchenia decora* in the Arikaree formation in Slope (White Butte, East Rainy Butte) and Stark (Obritsch Ranch) Counties and *Merychyus* and *Merycochoerus* in the Arikaree Formation in the Killdeer Mountains, Dunn County, suggests two distinct ages for the Arikaree Formation in North Dakota. The Arikaree fossil assemblages from Slope and Stark Counties indicate a Whitneyan to early Arikareean age, whereas the fossil assemblage from the Killdeer Mountains indicates a late Arikareean age.

R. Tedford (personal communication., 1989, 1994) wrote that *Merychyus* and *Merycochoerus* have overlapping range zones in the late Arikareean. In addition, he noted that this overlap occurs in the uppermost part of the Arikaree Group (Upper Harrison beds; see MacFadden and Hunt, this volume) and the base of the Ogallala Group (Runningwater Formation) in Nebraska. R. Hunt (personal communication, 1996) also observed that the earliest occurrence of *Merycochoerus* in the Great Plains is found in the Upper Harrison beds that produced the latest Arikareean fauna of western Nebraska and southeastern Wyoming. Although *Merycochoerus* ranges into the early Hemingfordian Runningwater deposits in Nebraska, Hunt stated that the association of *Merycochoerus* with *Merychyus* in the Killdeer Mountains suggests a late Arikareean age. This interpretation of an Arikareean age for the Arikaree Formation in Dunn County is consistent with an age of 25.1 ± 2.2 Ma from fission track dating of volcanic glass from the middle part of the Arikaree Formation in the Killdeer Mountains (Murphy et al., 1993). This age was obtained from the base of the burrowed marker unit 27.4 m below the stratigraphic position of the *Merycochoerus-Merychyus* oreodont association. The appearance of *Merycochoerus* in this part of the Arikaree Formation compares favorably with this taxon's occurrence in the type Arikaree in northwest Nebraska.

The mammalian taxa present in the Arikaree Formation in North Dakota imply that the erosional remnants of the formation exposed in Stark and Slope Counties are older than those in Dunn County, although no fossils were found in the lower part of the formation in the Killdeer Mountains to establish a basal age for the formation in Dunn County.

SUMMARY

1. In southwestern North Dakota, exposures of the Chadron, Brule, and Arikaree Formations are scattered and isolated. They are best observed in the areas of the Little Badlands, Chalky Buttes, Rainy Buttes, and Killdeer Mountains.

2. The Chadron Formation is divisible into two members. The basal Chalky Buttes Member consists primarily of yellow-green to white, gravel-bearing, cross-bedded sandstones and sandy mudstones. The overlying South Heart Member is composed of green to gray smectitic claystones containing lenticular beds of freshwater limestones.

3. The Chadron Formation unconformably overlies formations within the Fort Union Group and conformably underlies the Brule Formation.

4. The Chalky Buttes and South Heart Members of the Chadron Formation both contain mammalian fossils of Chadronian age.

5. The Brule Formation in southwestern North Dakota consists of complexly interbedded lithologies of claystones, calcareous mudstones, siltstones, freshwater limestones, tuffaceous beds, and cross-bedded sandstones. The Antelope Creek tuff in the lower part of the Brule Formation is a useful marker bed.

6. The Brule Formation is unconformably overlain by the Arikaree Formation.

7. Mammalian fossils indicate that the lower part of the Brule Formation is Orellan in age and that the upper part of the Brule is possibly Whitneyan in age.

8. The Arikaree Formation is lithologically variable and consists of concretionary, cross-bedded, calcareous sandstones, siltstones, silty claystones, tuffaceous beds, and carbonates.

9. The Arikaree Formation unconformably overlies either the Chadron or Brule Formations and is overlain by unconsolidated sediments, primarily lag gravels, of uncertain age.

10. Mammalian fossils suggest a late Whitneyan to late Arikareean age for the Arikaree Formation in North Dakota, which is consistent with an age of 25.1 ± 2.2 Ma from fission track dating of volcanic glass from the formation.

ACKNOWLEDGMENTS

This investigation was initiated as a cooperative project between the North Dakota Geological Survey and the U.S. Geological Survey through the Cooperative Geological Mapping Program (Project 14-08-0001-A0805). We thank the following people for assistance with various aspects of this project: Richard LeFever, John Kay, Steven McLean, Eric Harvey, James Sorenson, Glen Izett, Charles Naeser, Nancy Naeser, Roy Knight, James Martin, Phil Bjork, Janet Whitmore, George Lammers, Richard Tedford, Robert Hunt, Tim Heaton, Robert Emry, Dean Pearson, Bob Purdy, Sergiu Mamay, Donald Prothero, Allen Kihm, Johnathan Campbell, and Todd Strand. We also thank Jason Lillegraven, James Martin, and Margaret Stevens for providing useful reviews of the manuscript.

In addition, we thank the following landowners and their families for cordially allowing access to their land: Leonard Buzalsky, Ryan Brooks, Ted Pope, Albert Privratsky, Ben Privratsky, William Schmidt, George Schmidt, Edward Schmidt, Ron Obritsch, Bob Obritsch, Bob Fitterer, Jack Murphy, Alec Dvornak, Lorin Dvornak, Ken Urlacher, Ralph Urlacher, Rick Maxiner, Fred Thomas, Nick Biel, Orion Bruvold, Gene Fisher, Sherman Oakland, and Jeff and Laurie Oakland.

REFERENCES CITED

Benson, W. E., 1952, Geology of the Knife River area, North Dakota: U.S. Geological Survey Open-File Report, 323 p.

Cope, E. D., 1883a, A letter to the Secretary: Proceedings of the American Philosophical Society, v. 21, p. 216–217.

Cope, E. D., 1883b, On a new extinct genus and species of Percidae from Dakota Territory: American Journal of Science, v. 25, p. 414–416.

Delimata, J. J., 1975, Petrology and geochemistry of the Killdeer carbonates [Ph.D. thesis]: Grand Forks, University of North Dakota, 256 p.

Denson, N. M., and Gill, J. R., 1965, Uranium-bearing lignite and carbonaceous shale in the southwestern part of the Williston Basin—a regional study: U.S. Geological Survey Professional Paper 463, 75 p.

Denson, N. M., Bachman, G. O., and Zeller, H. D., 1959, Uranium-bearing lignite in northwestern South Dakota and adjacent states: U.S. Geological Survey Bulletin 1055-B, p. 11–57.

Douglass, E., 1908a, Rhinoceroses from the Oligocene and Miocene deposits of North Dakota and Montana: Carnegie Museum Annals, v. 4, p. 256–266.

Douglass, E., 1908b, Fossil horses from North Dakota and Montana: Carnegie Museum Annals, v. 4, p. 267–277.

Douglass, E., 1909, A geological reconnaissance in North Dakota, Montana, and Idaho; with notes on Mesozoic and Cenozoic geology: Carnegie Museum Annals, v. 5, p. 211–288.

Emry, R. J., Russell, L. S., and Bjork, P. R., 1987, The Chadronian, Orellan, and Whitneyan North American Land Mammal Ages; in Woodburne, M. O., ed., Cenozoic mammals of North America: Berkeley, University of California Press, p. 118–152.

Evans, J. E., and Terry, D. O., 1994, The significance of incision and fluvial sedimentation in the basal White River Group (Eocene-Oligocene), Badlands of South Dakota, U.S.A.: Sedimentary Geology, v. 90, p. 137–152.

Forsman, N. F., 1986, Documentation and diagenesis of tuffs in the Killdeer Mountains, Dunn County, North Dakota: North Dakota Geological Survey Report of Investigation 87, 13 p.

Hares, C. J., 1928, Geology and lignite resources of the Marmarth Field, southwestern North Dakota: U.S. Geological Survey Bulletin 775, 110 p.

Heaton, T. H., and Emry, R. J., 1996, Leptomerycidae, in Prothero, D. R., and Emry, R. J., eds., The terrestrial Eocene-Oligocene transition in North America: New York, Cambridge University Press, p. 581–608.

Hoganson, J. W., 1986, Oligocene stratigraphy of North Dakota, in Clausen, E., and Kihm, A. J., Tertiary and Upper Cretaceous of south-central and western North Dakota, in Field Trip Guidebook: Bismarck, North Dakota, Geological Society, p. 36–40.

Hoganson, J. W., and Lammers, G. E., 1992, Vertebrate fossil record, age and depositional environments of the Brule formation (Oligocene) in North Dakota, in Erickson, J. M., and Hoganson, J. W., eds., Frank D. Holland, Jr., Symposium: Bismarck, North Dakota Geological Survey Miscellaneous Series 76, p. 243–257.

Kepferle, R. C., and Culbertson, W. C., 1955, Strippable lignite deposits, Slope and Bowman Counties, North Dakota: U.S. Geological Survey Bulletin 1015-E, p. 123–182.

Kihm, A. J., 1990, Stratigraphic distribution of *Eumys* in the Brule Formation of North Dakota and its biochronologic implications: Proceedings of the North Dakota Academy of Science, v. 44, p. 69.

Leonard, A. G., 1922, The White River Formation in North Dakota: North Dakota Quarterly Journal, v. 12, p. 27–114.

Lillegraven, J. A., 1970, Stratigraphy, structure, and vertebrate fossils of the Oligocene Brule Formation, Slim Buttes, northwestern South Dakota: Geological Society of America Bulletin, v. 81, p. 831–850.

Lillegraven, J. A., 1993, Correlation of Paleogene strata across Wyoming—a users' guide, in Snoke, A. W., Steidtmann, J. R., and Roberts, S. M., eds., Geology of Wyoming: Geological Survey of Wyoming Memoir 5, p. 414–477.

Moore, E. C., Mehlin, R. E., and Kepferle, R. C., 1959, Uranium-bearing lignite in southwestern North Dakota: U.S. Geological Survey Bulletin 1055-E, p. 147–166.

Murphy, E. C., Hoganson, J. W., and Forsman, N. F., 1993, The Chadron, Brule, and Arikaree Formations in North Dakota: North Dakota Geological Survey Report of Investigation 96, 144 p.

Pearson, D. A., 1993, The Medicine Pole Hills local fauna—the first diverse Chadronian (latest Eocene)-age fauna from North Dakota, in Kihm, A. J., and Hartman, J. H., eds., The Marshall Lambert Symposium: Bismarck, North Dakota Geological Society, p. 24–25.

Pearson, D. A., and Hoganson, J. W., 1995a, The Medicine Pole Hills local fauna: Chadron Formation (Eocene: Chadronian), Bowman County, North Dakota: Proceedings of the North Dakota Academy of Science, v. 49, p. 65.

Pearson, D. A., and Hoganson, J. W., 1995b, The Medicine Pole Hills local fauna: Chadron Formation (Eocene) vertebrate assemblage from Bowman County, southwestern North Dakota: North-Central and South-Central Sections, Geological Society of America Abstracts with Programs, v. 27, no. 3, p. A79.

Powers, W. E., 1945, White River Formation of North Dakota [abs.]: Geological Society of America Abstracts with Programs, v. 56, p. A1192.

Prothero, D. R., and Shubin, N., 1989, The evolution of Oligocene horses; in Prothero, D. R., and Schoch, R. M., The evolution of perissodactyles: New York, Oxford University Press, p. 142–175.

Prothero, D. R., Denham, C. R., and Farmer, H. G., 1983, Magnetostratigraphy of the White River Group and its implications for Oligocene geochronology: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 42, p. 151–166.

Quirke, T. T., 1913, Geology of the Killdeer Mountains, Dunn County, North Dakota [M.S. thesis]: Grand Forks, University of North Dakota, 41 p.

Quirke, T. T., 1918, The geology of the Killdeer Mountains, North Dakota: Journal of Geology, v. 26, p. 255–271.

Retallack, G. J., 1983, Late Eocene and Oligocene fossil soils from Badlands National Park, South Dakota: Geological Society of America Special Paper 193, 82 p.

Seager, O. A., Blackstone, D. L., Jr., Cobban, W. A., Downs, G. R., Laird, W. M., and Sloss, L. L., 1942, Stratigraphy of North Dakota: Bulletin of the American Association of Petroleum Geologists, v. 26, n. 8, p. 1414–1423.

Seeland, D., 1985, Oligocene paleogeography of the northern Great Plains and adjacent mountains, in Flores, R. M., and Kaplan, S. S., eds., Cenozoic paleogeography of the west-central United States, Rocky Mountain Paleogeography Symposium 3: Denver, Colorado, Rocky Mountain Section, Society of Economic Paleontologists and Mineralogists, p. 187–205.

Skinner, M. F., 1951, The Oligocene of western North Dakota, in Bump, J. D., ed., Guidebook, 5th Annual Field Conference, western South Dakota, August-September 1951: Rapid City, Society of Vertebrate Paleontology, p. 51–58.

Stone, W. J., 1973, Stratigraphy and sedimentary history of middle Cenozoic (Oligocene and Miocene) deposits in North Dakota [Ph.D. thesis]: Grand Forks, University of North Dakota, 217 p.

Storer, J. E., 1978, Rodents of the Calf Creek local fauna (Cypress Hills Formation, Oligocene, Chadronian), Saskatchewan: Saskatchewan Museum of Natural History, Natural History Contribution 1, 54 p.

Terry, D. O., LaGarry, H. E., and Wells, W. B., 1995, The White River Group revisited: vertebrate trackways, ecosystems, and lithostratigraphic revision, redefinition, and redescription, in Flowerday, C. A., ed., Geologic field trips in Nebraska and adjacent parts of Kansas and South Dakota, Guidebook 10: Lincoln, Nebraska, Conservation and Survey Division, Institute of Agriculture and Natural Resources, University of Nebraska–Lincoln, p. 43–57.

Wood, H. E., Chaney, R. W., Clark, J., Colbert, E. H., Jepsen, G. L., Reeside, J. B., Jr., and Stock, C., 1941, Nomenclature and correlation of the North American continental Tertiary: Geological Society of America Bulletin, v. 52, p. 1–48.

MANUSCRIPT ACCEPTED BY THE SOCIETY DECEMBER 5, 1997

Printed in U.S.A.

Geological Society of America
Special Paper 325
1998

Arikareean and Hemingfordian faunas of the Cady Mountains, Mojave Desert Province, California

Michael O. Woodburne
Department of Earth Sciences, University of California, Riverside, California 92521

ABSTRACT

The Mojave Desert of southern California contains a number of basins that preserve mammal-bearing sediments, and intercalated volcanic tuffs, of late Oligocene to Miocene age. Of these, only the Cady Mountains preserve well-studied sections with mammal fossils of Arikareean to early and late Hemingfordian age. This report is an update of geochronologic and faunal information contained in these strata, known as the Hector Formation. With that accomplished, these radioisotopic and faunal data aid in assessing the age of presumptively correlative first appearance datum (FAD) of taxa in the type sequences of Nebraska.

The Black Butte Mine and Logan Mine Local Faunas are found in superposition in the Hector Formation on the southwestern flank of the Cady Mountains. The two assemblages appear to best correlate with taxa of late Arikareean age elsewhere and are calibrated by a K-Ar age of ~21.6 Ma found stratigraphically near the upper part of the Black Butte Mine sequence. Black Butte Mine taxa important for regional correlation include *Merychyus calaminthus* and *Stenomylus* cf. *hitchcocki* of late Arikareean (Ar3) age, with *Merychyus* having a lowest stratigraphic datum (LSD) in the Harrison Formation of Nebraska. In the Cady Mountains, *M. calaminthus* is bracketed by tuffs dated at 22.9 and 21.6 Ma, suggesting that the coeval part of the Harrison Formation is ~22 Ma old. Temporally significant taxa in the Logan Mine Local Fauna include *Michenia* cf. *M. agatensis* and *Ustatochoerus* cf. *leptoscelos* (Ar4). *Michenia* (including *M. agatensis*) has a lowest stratigraphic datum (LSD) in the Marsland Formation in Nebraska (Upper Harrison beds, as restricted by Hunt, 1985, 1990) dated at ~19.2 Ma. The deposits that bear the Logan Mine Local Fauna include lithic fragments eroded from the Peach Springs Tuff dated at 18.5 ± 0.2 Ma. An LSD of *Michenia* cf. *agatensis* below the Peach Springs Tuff clasts in the Hector Formation is similar to the LSD of that taxon in the late Arikareean (Ar4) fauna of the Upper Harrison beds of Nebraska. An LSD of *Ustatochoerus* cf. *leptoscelos* of the Logan Mine Local Fauna just above the Peach Springs Tuff clasts correlates to an early Hemingfordian (He1) age in Nebraska (taxa in the Runningwater Formation). Whereas the Black Butte Mine and most of the Logan Mine taxa are consistent with correlations to Ar3 and Ar4, some Logan Mine faunal elements thus range into He1 in the mammalian biochronological hierarchy. The Logan Mine Local Fauna nevertheless records the Arikareean/Hemingfordian boundary in the Mojave Desert Province of California, interpolated at ~18.8 m.y. as bracketed between clasts of the 18.5-Ma-old Peach Springs Tuff (above) and by a basalt considered on paleomagnetic grounds as being 19 to 20 m.y. old (below) in the Northern Cady Mountains.

Woodburne, M. O., 1998, Arikareean and Hemingfordian faunas of the Cady Mountains, Mojave Desert Province, California, *in* Terry, D. O., Jr., LaGarry, H. E., and Hunt, R. M., Jr., eds., Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America): Boulder, Colorado, Geological Society of America Special Paper 325.

The Lower Cady Mountains Local Fauna and the Upper Cady Mountains Local Fauna are found in superposition on the northern flanks of the range. Basal radioisotopic ages of 18.6 and 17.9 Ma, respectively, are provided by dated volcanic units that underlie each faunal unit, but these ages are revised herein to be 19 to 20 and 18.5 Ma, respectively. The Aletomerycinae LSD in the Lower Cady Mountains Local Fauna compares favorably with the LSD of *Aletomeryx* in the Runningwater Formation of Nebraska, interpolated in California at ~18.7 Ma. The presence of *Aepycamelus* in the Upper Cady Mountains Local Fauna, correlated to Barstow Formation faunas having a paleomagnetic age of 16.8 to 17.3 Ma, compares favorably with an LSD of that genus in the Box Butte Formation of Nebraska and helps constrain its age there.

The faunal, radioisotopic, and magnetostratigraphic data in the Cady Mountains, Mojave Desert Province, southern California, suggest that local LSDs of *Merychyus, Michenia, Ustatochoerus*, Aletomerycinae, and *Aepycamelus* comprise respective FADs at the scale of California to Nebraska during the early Miocene, aid in giving a correlated age to faunal units in Nebraska, and confirm that the Arikareean/Hemingfordian boundary in Nebraska is ~18.8 m.y. old.

INTRODUCTION

This chapter provides an update on the faunal and radioisotopic correlations of the Black Butte Mine and Logan Mine Local Faunas of late Arikareean age in the Mojave Desert Province of southern California (Fig. 1) originally described in Woodburne et al. (1974). The new information also leads to a reevaluation of the age and extent of these faunal units and a determination of the Arikareean/Hemingfordian boundary in the Cady Mountains. The chapter also revises information on the Lower Cady Mountains and Upper Cady Mountains Local Faunas found on the north side of the range. Correlations take into account radioisotopic and magnetostratigraphic data acquired since 1974. Terminology appropriate for discussions of correlation vs. calibration is provided. The Mojave Desert Province is taken as the region now bounded on the southwest by the San Andreas fault; on the south by the Pinto Mountain fault; on the north and northwest by the Garlock fault, and on the east by the southern extension of the Death Valley fault zone, the Granite Mountains fault (Fig. 1).

GEOLOGIC OVERVIEW AND PREVIOUS WORK

Woodburne et al. (1974) described the oldest Cenozoic mammal faunas from the Mojave Desert Province, California. The assemblages were grouped as the Black Butte and Logan Mine Local Faunas, and considered to be of late Arikareean and early Hemingfordian age, respectively. Woodburne et al. (1982) and Tedford et al. (1987) revised the name of the Black Butte Mine Local Fauna because the original term was already in use.

The fossil mammals are preserved in the Hector Formation that crops out on the southwestern and northern flanks of the Cady Mountains (Fig. 1). In the Hector type area, a volcaniclastic succession ~500 m thick (H) unconformably overlies a terrane of extrusive volcanic rocks (Tv) and is unconformably overlain by Quaternary alluvium (Fig. 1). The extrusive volcanic succession is composed largely of andesitic lahar, and agglomerate considered to be coeval with a succession of andesite, andesitic basalt, andesite to rhyodacite flows, breccias, and rhyolite domes that crop out within the Cady Mountains and adjacent to the Mojave River to the north (Fig. 1,Tv) (Williamson, 1980, Fig. 1; Moseley, 1978; Woodburne et al., 1982; Woodburne, 1991). This extrusive succession is the first local Cenozoic record of tectonic depositional events on the Mojave Desert Province, although it is underlain in part by red arkosic sandstone or sedimentary breccia of unknown age (see also Byers, 1960, for a similar statement for Alvord Mountain, located ~15 km to the northwest, as shown in Woodburne et al., 1982, Fig. 2). Dokka et al. (1991, and references cited therein) consider the volcanic rocks to represent the initial phase of a regionally extensive interval of crustal extension that lasted from ~26 to 19 m.y. (see also Woodburne, 1991, for a regional summary of rocks of this age within the Mojave Desert Province).

Miller (1980) mapped exposures of the Hector Formation on the northern flank of the Cady Mountains and described the Lower Cady Mountains and Upper Cady Mountains Local Faunas from these strata, of early and late Hemingfordian age, respectively. The base of the Hector Formation succession in the northern Cady Mountains is calibrated at ~22.9 Ma (Fig. 1, K-Ar, biotite), and the Lower Cady Mountains Local Fauna postdates a basalt dated at ~18.6 Ma (K-AR, whole rock; hereafter revised to ca. 19–20 m.y.). The Upper Cady Mountains Local Fauna postdates an ignimbrite dated at ~17.9 Ma (K-Ar, sanidine; herein revised to 18.5 Ma) and is correlated regionally with faunas that are ~16.8 to 17.3 m.y. old (Miller, 1980; MacFadden et al., 1990b; Woodburne et al., 1982; Woodburne, 1991). In this chapter, taxa of late Arikareean age are recognized in the northern Cady Mountains and occur stratigraphically ~21 m (70 ft) below the 19- to 20-m.y.-old basalt. In terms of regional correlations (e.g., Woodburne and Swisher, 1995; this chapter), this basalt lies at about (somewhat older than) the Arikareean/Hemingfordian boundary. See also MacFadden and Hunt, this volume. As well as in its type

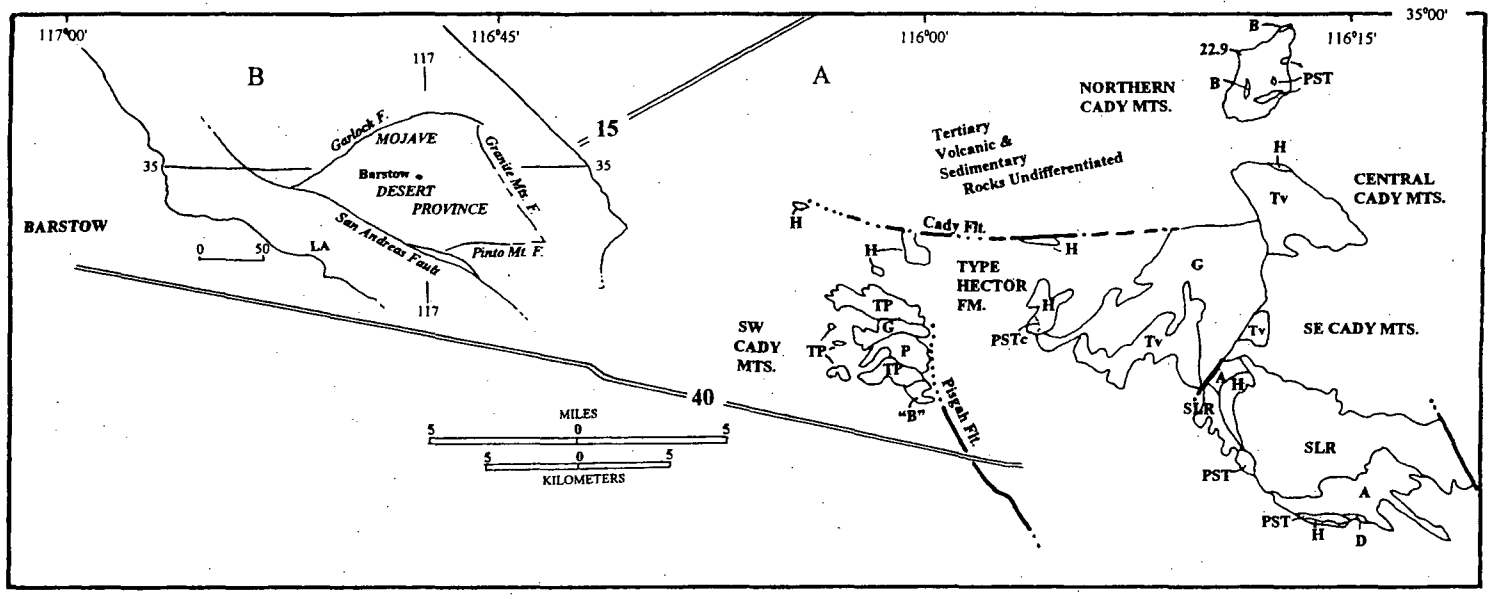


Figure 1. Selected elements of the geology of the Cady Mountains (A), Mojave Desert Province, southern California. Inset map (B) shows limits of Mojave Desert Province. Northern Cady Mountains after Miller (1980); central Cady Mountains after Williamson (1980); southwestern Cady Mountains after Ross (1995); type Hector Formation after Woodburne et al. (1974); southeastern Cady Mountains after Glazner (1988). Unless otherwise noted, all Tertiary units are overlain by Quaternary alluvium. Modified from Geologic Map of California, San Bernardino sheet; scale 1:250,000. Key: Northern Cady Montains 22.9 = locations of tuff yielding that date in Ma; B = basalt dated at 18.6 Ma (revised herein to ca. 19–20 Ma); PST = Peach Springs Tuff dated 17.9 Ma, but revised herein to 18.5 Ma. Central Cady Mountains; Tv = pre–Hector terrane of eruptive volcanic units, lahars, domes; H = Hector Formation (uppermost unit). Southwestern Cady Mountains G = Mesozoic granitic rock; P = Formation of Poe; TP = Formation of Troy Peak. P and TP are regionally correlative with Tv (Tertiary volcanic rock; Williamson, 1980) that underlies the type Hector Formation. "B" = "Barstow Formation," a lithostratigraphic correlation not accepted herein. Type Hector Formation. H = Hector Formation; PSTc = clasts reworked from Peach Springs Tuff. Southeastern Cady Mountains SLR = Formation of Sleeping Beauty Ridge; A = Formation of Argos Station; D = Dacite of Cady Mountains; PST = Peach Springs Tuff; H = Hector Formation. In Glazner (1988), all but H gave radioisotopic ages of ca. 19 to 20 m.y., but age of PST is now considered (Nielson et al., 1990) as 18.5 Ma. H in this area postdates the PST. If the calibrations of the other units are tenable, then all correlate approximately with either the basal part of the type Hector Formation, or with the immediately underlying Tv, and with P and TP of the southwestern Cady Mountains.

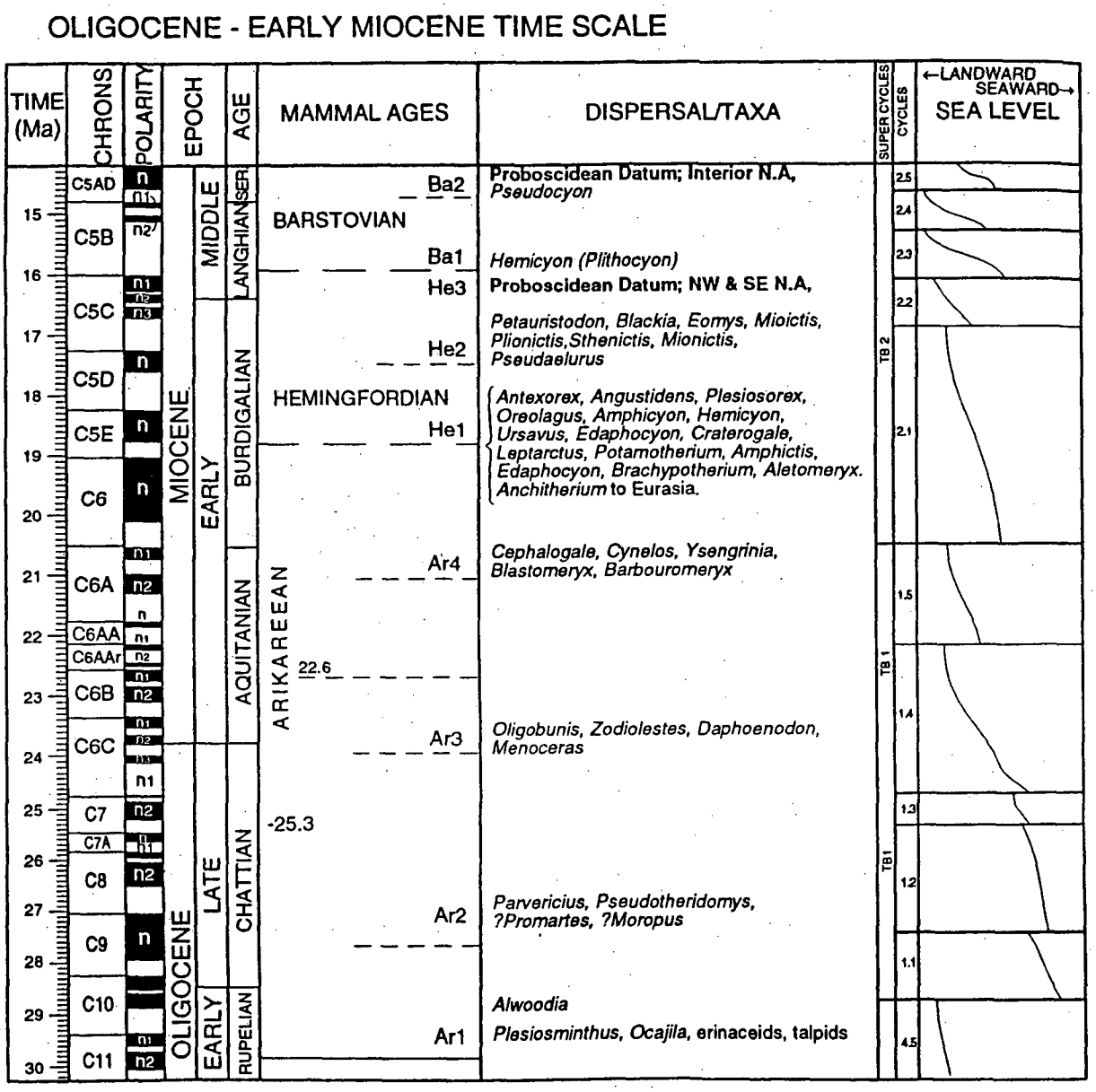


Figure 2. Time scale for the later Oligocene and early Miocene epochs; after Woodburne and Swisher (1995). Ar1–4 are subages of the Arikareean Land Mammal Age; He 1–3 are subages of the Hemingfordian Land Mammal Age; Ba1–2 are subages of the Barstovian Land Mammal Age. Sea-level stands follow Woodburne and Swisher (1995).

area, the Hector Formation of the northern Cady Mountains holds the potential of being a calibrated sequence containing the Arikareean/Hemingfordian boundary in North America.

MacFadden et al. (1990b) showed that the Hector Formation in the northern Cady Mountains (Fig. 1) underwent ~21° of clockwise rotation about vertical axes subsequent to its deposition (~16 Ma). Ross (1992, 1995), Ross et al. (1991), and Dokka et al. (1991) indicated that the area west of the type Hector Formation outcrops underwent a total clockwise rotation of ~130° ± 13° about vertical axes (Fig. 1, southwest Cady Mountains). The sequence of rotation involved two steps: ca. 64° ± 19° clockwise prior to 16.5 Ma that affected both the originally extended and tilted succession (Fig. 1, Formation of Poe; P) and the unconformably overlying Formation of Troy Peak (Fig. 1, TP). The Formation of Troy Peak is correlated to strata that underlie the type Hector Formation, and thus is older than 21.6 Ma (see below). Subsequent to the deposition of the Formation of Troy Peak, another interval of rocks (Fig. 1, "B," "Barstow Formation") was deposited, beginning ~16.5 Ma and continuing to at least 14.0 Ma. After this latest succession was formed and tilted, the region underwent an additional 67° ± 15° of clockwise rotation.

Whether the type Hector Formation rocks experienced this style of rotation still has yet to be determined. A northern extension of the Pisgah fault (Fig. 1) may structurally separate the southwestern Cady Mountains from the deposits in the Hector type area. On the other hand, Ross (1995) and Ross et al. (1989) have suggested that rocks older than ~16 Ma in the central Mojave Desert Province were affected by regional clockwise rotation of ~50°. Yet the Hector Formation of the northern Cady Mountains, which spans an interval from ca. 23 to 16 m.y., shows neither this dichotomy of rotation history nor its magnitude. The northern and southern Cady Mountains Hector Formation terranes are separated by the east-west–trending, left-slip Cady fault (Fig. 1), which may have insulated the northern terrane from some of the regional rotation.

TERMINOLOGY AND ABBREVIATIONS

Institutional abbreviations

UCR: Department of Earth Sciences, University of California, Riverside, CA 92521

AMNH: Department of Vertebrate Paleontology, American Museum of Natural History, New York, 10024

Chronologic terms

Calibration: A statement about the age of a rock or faunal unit based on materials amenable to radioisotopic dating that are directly associated with the unit in question

Correlation: A statement about the age of a rock or faunal unit based on indirect means of calibration (e.g., faunal, stratigraphic, or magnetostratigraphic, to a radioisotopic scale). This includes extrapolation of ages between dated levels and interpolation of ages within dated intervals, as well as to ages assigned to reversal chrons in the Magnetic Polarity Time Scale based on spreading rate calculations. An association of faunal and magnetostratigraphic data may carry an implication of age in the Ma scale, but this is not a calibration

Chron: A magnetic polarity interval in the MPTS; nomenclature follows Berggren et al. (1995)

FAD: First Appearance Datum; a regionally significant change in the fossil record (e.g., Berggren and Van Couvering, 1974)

Fauna: An assemblage of one or more taxa of fossil mammals found in more than one geographic and stratigraphic location (e.g., Tedford et al., 1987, p. 155)

HSD: Highest Stratigraphic Datum, a local stratigraphic last appearance (e.g., Lindsay et al., 1984)

LAD: Last Appearance Datum; a regionally significant temporal change in the fossil record (e.g., Berggren and Van Couvering, 1974)

Local Fauna: An assemblage of one or more taxa of fossil mammals found in a geographically and stratigraphically limited location (e.g., Tedford et al., 1987, p. 155)

LSD: Lowest Stratigraphic Datum; a local stratigraphic first appearance (e.g., Lindsay et al., 1984)

Ma: Megannum in the radioisotopic time scale

MPTS: Magnetic polarity time scale, specifically that of Berggren et al. (1995)

m.y.: Million years in duration or interval but not directly tied to the radioisotopic time scale

Mammal age terms

Calibration and correlation of the Arikareean and Hemingfordian intervals (Fig. 2) of the North American Mammal Age sequence follows Tedford et al. (1987; Figs. 6.2, 6.3), Tedford et al. (1995) and Woodburne and Swisher (1995).

Arikareean: Mammal age as defined by Tedford et al. (1987), modified by Tedford et al. (1995), and as shown in Figure 2. The Arikareean mammal age begins with the immigrant first occurrence of *Plesiosminthus* considered to correlate to an age of ~30 m.y. (Woodburne and Swisher, 1995; *Ocajila*, erinaceid, and talpid insectivores deleted as per R. H. Tedford et al., 1995, who indicated that these taxa are not basal Arikareean). The Arikareean contains four subages, Ar1 through Ar4. Ar2 begins with the immigrant occurrence of *Alwoodia, Parvericius,* and *Pseudotheridomys*, along with *Moropus* and *Promartes* and is ~28.0 to 27.5 m.y. old (R. H. Tedford et al., 1995; Woodburne and Swisher, 1995). Ar3 begins with the immigrant occurrence of *Oligobunis, Zodiolestes, Daphoenodon,* and *Menoceras,* correlated near the Oligocene-Miocene boundary at ~23.7 m.y. Ar4 begins with the immigrant occurrence of *Cephalogale, Cynelos, Ysengrinia, Blastomeryx,* and *Barbouromeryx*, correlated as ~20.7 m.y. (Woodburne and Swisher, 1995).

Hemingfordian: Mammal age as defined by Tedford et al. (1987), and as shown in Figure 2 (after Woodburne and

Swisher, 1995). The Hemingfordian mammal age begins with the immigrant occurrence of *Antesorex, Angustidens, Plesiosorex, Oreolagus, Amphicyon, Hemicyon, Ursavus, Edaphocyon, Craterogale, Leptarctus, Potamotherium, Amphictis, Edaphocyon, ?Brachypotherium* (query from R. H. Tedford, personal communication, 1995), and *Aletomeryx*, correlated as being ~18.8 m.y. old (Fig. 2; age follows MacFadden and Hunt, this volume). This major immigration episode (He1) is followed by another immigration event, He2. He2 is heralded by the arrival at ~17.5 m.y. of the immigrants *Petauristodon, Blackia, Eomys, Plionictis, Sthenictis, Mionictis,* and *Pseudaelurus*. Neither He1 nor He2 is correlated with a specific sea-level low stand (Woodburne and Swisher, 1995) (Fig. 2). The latest episode of the Hemingfordian, He3, is recognized on the basis of the Proboscidean Datum in the northwest and southeastern extremities of the United States and correlated as being ~16.3 m.y. old (Woodburne and Swisher, 1995).

Lithostratigraphic terms

Barstow Formation vs. Hector Formation: Woodburne et al. (1982) summarized the potential regional distribution of the Barstow and Hector formations in the Mojave Desert Province. The term Barstow Formation was extended eastward to include strata of Alvord Mountain (Byers, 1960), who proposed this correlation primarily on similarity in age, as shown by faunal and (later: Woodburne et al., 1982) radioisotopic data, rather than on lithologic details. Woodburne et al. (1982) did not include strata now embraced within the Hector Formation in the northern Cady Mountains as "Barstow Formation." Subsequently, authors attempting regional syntheses (e.g., Dokka et al., 1991; Ingersoll et al., 1996) have utilized the term "Barstow Formation" as shorthand for strata that are both younger than ~16 Ma and also signify postextensional deposition on the Mojave Desert Province. However, Woodburne et al. (1990) reappraised the lithostratigraphy of the type Barstow Formation (Mud Hills, ~15 km northwest of Barstow; Fig. 1 herein; also Woodburne et al., 1982; Fig. 2) and determined that it ranges in age from ~19 to 13 m.y., based on a combination of faunal, radioisotopic, and magnetostratigraphic data. First, the Barstow Formation is likely to be partly coeval with extension within the Mojave Desert Province (which is considered to have terminated prior to the deposition of the Peach Springs Tuff at ca. 18.5 Ma, e.g., Dokka et al., 1991), and second, has a plausible regional occurrence (Woodburne et al., 1982) that excludes rock units that clearly have their own regional context and characteristic (different from Barstow Formation) lithology. For this chapter, the extension of the term Barstow Formation for strata in the southwestern Cady Mountains is utilized in the sense presented by Ross (1995). This usage does not indicate formal endorsement of this nomenclature for those strata. In fact, due to the numerous informal stratigraphic conventions utilized in the literature cited for Figure 1 (i.e., "formation of . . ."), it is timely to undertake studies that would present a regionally coherent and consistent stratigraphic nomenclature based on objective lithostratigraphy, and not on interpretive tectonostratigraphy. Until such occurs, the terms Barstow Formation and Hector Formation are restricted to the extent given in Woodburne et al. (1990), and the use of Barstow Formation by Ross (1995) is not supported formally.

BIOSTRATIGRAPHY AND GEOCHRONOLOGY

The following sections consider the biostratigraphy and regional correlation of the mammalian faunas of the Hector Formation in the Cady Mountains. Discussion begins with the type Hector Formation and is followed by that of the northern part of the range.

Type area of Hector Formation, Southern Cady Mountains

Black Butte Mine Local Fauna. This unit originally was named the Black Butte Fauna. Due to the term being preoccupied by a fauna of later Miocene age in Oregon (Shotwell et al., 1963), Tedford et al. (1987) supplied the amended name given above. As shown in Figures 3, 4, and 5, the Black Butte Mine Local Fauna is locally restricted in geographic and stratigraphic occurrence and is based on the stenomyline camel, *Stenomylus* cf. *S. (S.) hitchcocki,* and the oreodont *Merychyus calaminthus*. Both taxa co-occur near the base of the type Hector Formation ~100 m stratigraphically below the position of a tuff dated at 21.6 Ma (revised from 21.0 Ma in Woodburne et al., 1974, according to the calculations presented in Dalrymple, 1979). The camel, but not the oreodont, ranges above that tuff to a position just below a tuff that is used as a convenient lithologic marker to separate the Black Butte Mine and Logan Mine local faunas.

Logan Mine Local Fauna. The Logan Mine Local Fauna contains the camels, *Michenia* cf. *M. agatensis*, and cf. *Protolabis*, the oreodont, *Ustatochoerus* cf. *leptoscelos*; the amphicyonid, cf. *Daphoenodon*, and the mustelid, *Promartes*. The Logan Mine Local Fauna begins ~53 m (175 ft) stratigraphically above the marker tuff and ranges ~300 m (1000 ft) stratigraphically above that tuff. About 42 m (140 ft) above the marker tuff (adularescent sanidine clasts) (Fig. 3), the sediments contain clasts of a tuff bearing distinctive crystals of adularescent sanidine. Although not present in outcrop in the Hector type area, these clasts apparently were reworked from the similarly distinctive Peach Springs Tuff, a regionally widespread marker unit in western Arizona and the Mojave Desert of California (Nielson et al., 1990). The Peach Springs Tuff is 18.5 ± 0.2 Ma old (Nielson et al., 1990), so that if the correlation is viable, then this part of the Hector Formation appears to be ~18.5 Ma old, as well.

The Peach Springs Tuff crops out in the southeastern Cady Mountains (Fig. 1) (also see Glazner, 1988). The westernmost occurrence of the Peach Springs Tuff is in Daggett Ridge (Hillhouse and Wells, 1986; Wells and Hillhouse, 1989), ~8 km south of Barstow (Fig. 1), and Buesch (1994) recorded the presence of the Peach Springs Tuff in the Hector Formation of the

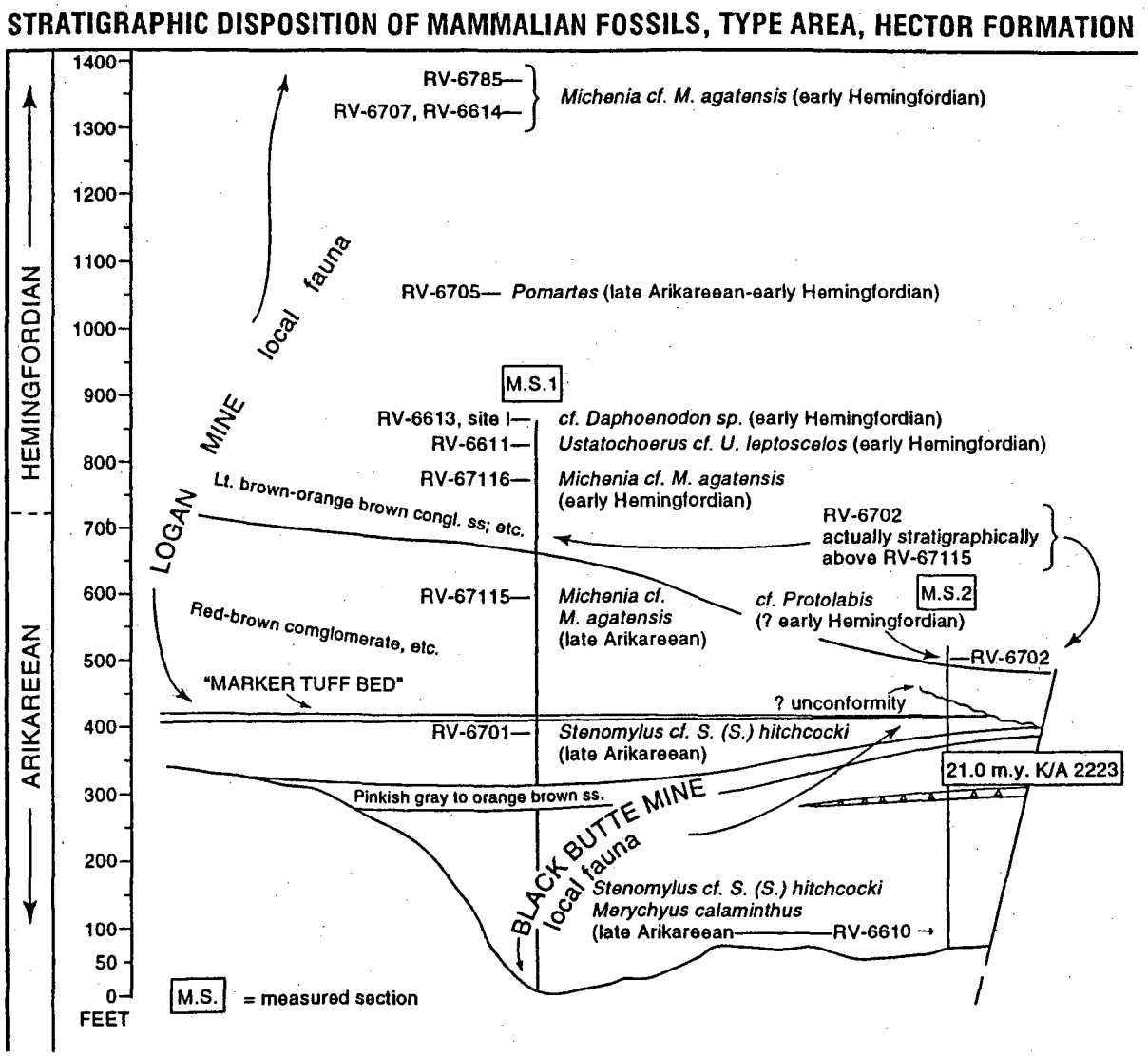


Figure 3. Stratigraphic disposition of mammalian fossils and selected lithostratigraphic features; modified from Woodburne et al. (1974; Fig. 4) to reflect changes discussed in the text. Stratigraphic thinning of the red-brown conglomerate above the "Marker Tuff Bed" is shown by the line that slopes downward between M.S. 1 and M.S. 2 (measured sections), so that, for example, locality RV-6702 is stratigraphically higher than RV-67115. The sloping line also corresponds to the stratigraphic position of an influx of adularescent sanidine clasts, considered to have been derived from the Peach Springs Tuff, dated at ca.18.5 Ma. Taxa that occur stratigraphically above this boundary are considered to correlate with faunas of early Hemingfordian age, but of these only *Ustatochoerus* cf. *leptoscelos* is an early Hemingfordian indicator.

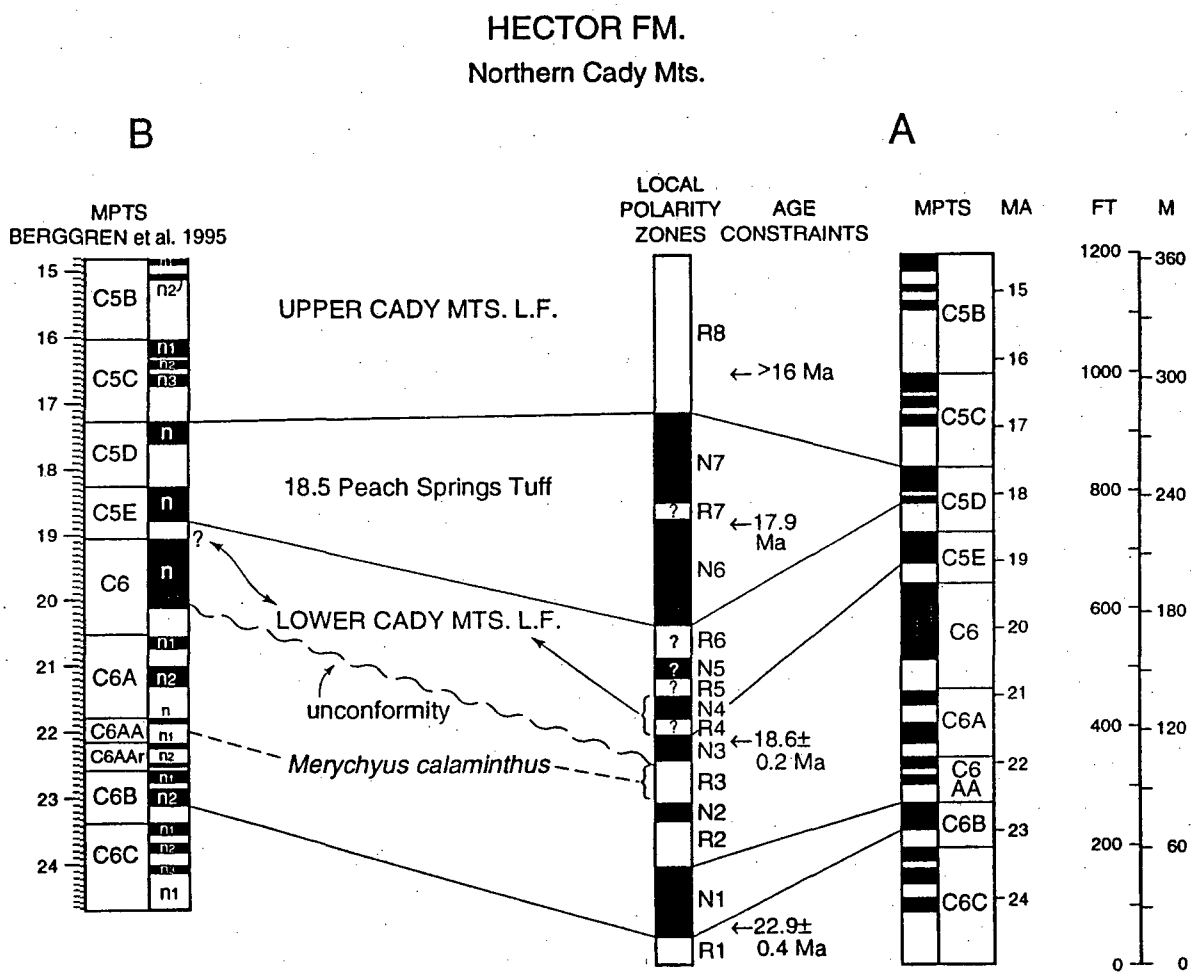


Figure 4. Lithostratigraphic, biostratigraphic, and magnetostratigraphic relationships of the Hector Formation in the northern Cady Mountains. A, After MacFadden et al. (1990). B, new interpretations discussed in the text. Time scale in B follows Berggren et al. (1995). *Merychyus calaminthus* in magnetozone R3 of A is considered to most likely correlate with Chron C6AA in B. Because of apparently lost magnetozones in A, an unconformity is proposed to be present at the base of N3 (below the basalt originally dated at 18.6 Ma) in A, so that N3 correlates with Chron C6n in B. Note that the basalt in A dated at 18.6 ± 0.2 Ma is considered to correlate to Chron C6n, based on discussions in the text, and to actually be 19 to 20 m.y. old. Similarly, the tuff in A dated at 17.9 Ma is considered to be a correlative of the Peach Springs Tuff that is 18.5 Ma old. Thus N6 in A most likely correlates to Chron C5En in B. The Lower Cady Mountains Local Fauna, taken from R4–N4 in A, thus is considered to correlate with C5Er? and n in B, and the intervals R5–R6 in A cannot be reconciled in B. The Upper Cady Mountains Local Fauna considered to be greater than about 16 Ma old in A correlates on paleontologic grounds with Chron C5Cr. Thus, correlation of the tuff at 17.9 Ma to the Peach Springs Tuff results in a changed interpretation of its age to 18.5 Ma. Recognition that the basalt in A dated at 18.6 Ma was extruded into a lake and thus that the age likely is a minimum one, plus considerations developed in the text, combine to suggest that its age actually is older than about 19 m.y.

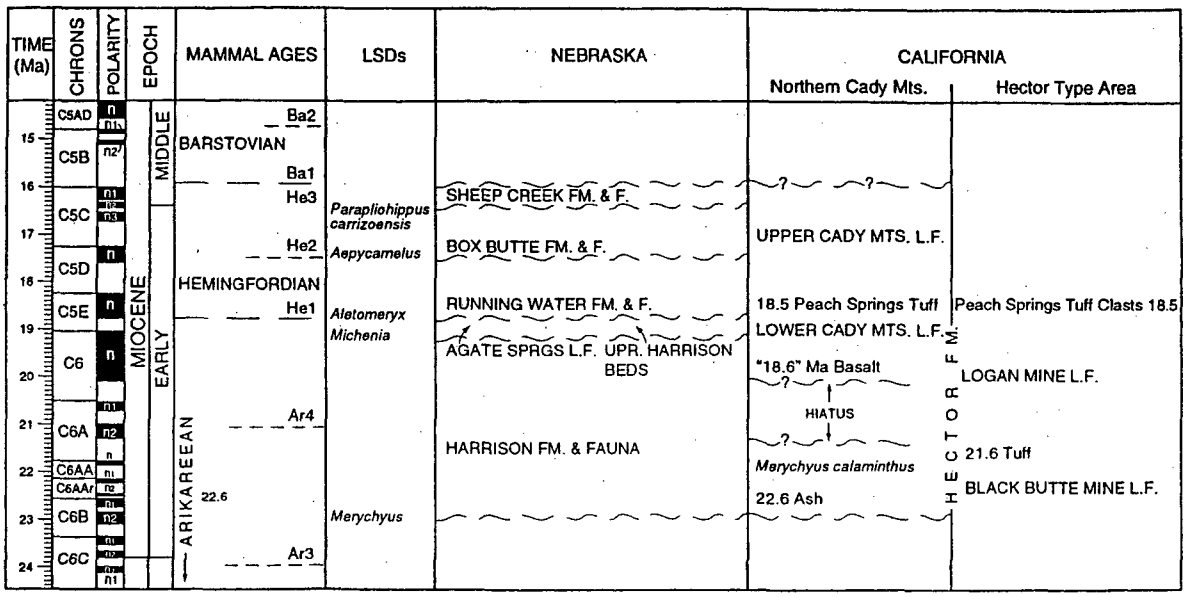


Figure 5. Regional correlation of the late Arikareean and Hemingfordian mammalian faunas of the Cady Mountains district to typical sequences in Nebraska. Time scale is after Woodburne and Swisher (1995), LSDs mainly after Tedford et al. (1987); see text.

northern Cady Mountains (PST) (Fig. 1). These regions surround the type area of the Hector Formation and support the suggestion that the adularescent clasts in the Hector Formation (Hector type area) (Fig. 1) are derived from the Peach Springs Tuff, but petrographic analyses have not been conducted to confirm this correlation.

Regional correlation

Woodburne et al. (1974) noted that the Black Butte Mine and Logan Mine Local Faunas contained taxa also preserved in the Harrison Formation and Upper Harrison beds of Nebraska, then considered to be of late Arikareean to early Hemingfordian age (for use of Upper Harrison beds, see MacFadden and Hunt, this volume). Hunt (1985, 1990) subsequently revised the stratigraphy and chronology of these units such that the famous Agate Springs Quarry Fauna occurs in the base of the Upper Harrison beds and is of latest Arikareean age. Tedford et al. (1987) also gave revised summaries of the key mammalian taxa from faunas of Arikareean and Hemingfordian age in Nebraska (as well as for other parts of North America), and provided important information on the stratigraphy and radioisotopic chronology of the rock units. Thus it now is possible to reappraise the correlation of the California faunas to those in Nebraska that include the units on which these mammal ages were based originally (Wood et al., 1941), and to suggest correlations of these to the radioisotopic time scale as recognized in plausibly coeval faunal units in California.

Black Butte Mine Local Fauna. Figures 4 and 5 show the correlations proposed herein, along with LSDs of key mammalian taxa. M. S. Stevens (in Woodburne et al., 1974) discussed the systematic paleontology of the oreodonts of the Hector Formation. She noted the strong similarity between *Merychyus calaminthus* from the Black Butte Mine Local Fauna and *M. crabilli* described by Schultz and Falkenback (1947) from the Harrison Fauna as utilized by Tedford et al. (1987). Stevens, in fact, suggested that the two species "most likely are regional varieties of a single widespread species population . . ." (Woodburne et al., 1974, p. 19; Stevens, 1977, p.36, 37). The Harrison Formation contains the LSD of *Merychyus*, as shown in Figure 5, following Tedford et al. (1987). The correlation proposed between the Black Butte Mine and Harrison faunas reflects this similarity, with a suggested FAD of *Merychyus* in Nebraska being ca 22.9 Ma old, based on the calibration in California.

B. E. Taylor (in Woodburne et al., 1974) discussed the faunal affinity of the stenomyline camel from the Black Butte Mine Local Fauna. This taxon, *Stenomylus* cf. *S. (S.). hitchcocki*, compares most favorably with *S. (S.) hitchcocki* from Stenomylus Quarry of the Harrison Formation as restricted by Hunt (1985, 1990).

Logan Mine Local Fauna. Tedford et al. (1987) noted that the camelid *Michenia* has an LSD in the Upper Harrison beds, which contains the fauna from Agate Springs Quarry at its base. In describing the camels of the Hector Formation, B. E. Taylor (in Woodburne et al., 1974) indicated that the Logan Mine *Michenia* was most similar to the genotypic species, *M. agatensis,* and differed dentally (lower crowned) from Harrison Formation material allocated to *Oxydactylus* and postcranially (longer and more

strongly fused metapodials) from the likely morphology of the comparably aged *Miolabis californicus* from the Tick Canyon Formation of California. The correlation shown in Figure 5 incorporates the LSD of *Michenia agatensis* in California and Nebraska. In that these LSDs occur stratigraphically above chronologically older mammalian faunas in both places, they approach a potential FAD at that regional scale, consistent with an age of ~19.2 Ma in Nebraska and older than a possible FAD of *Aletomeryx,* that is in turn older than ~18.5 Ma in California. A potential regional FAD of *Michenia* from Nebraska to California is consistent with its age being ~19 m.y. old.

Stevens (in Woodburne et al., 1974) also considered the phyletic affinities of the oreodont, *Phenacocoelus* cf. *P. leptoscelos* and considered that the Logan Mine form was closest to the species found in the lower part of the Delaho Formation of Texas, in the Castolon Local Fauna of late Arikareean age (Stevens et al., 1969; Stevens, 1977). Stevens (*in* Busbey and Lehman, 1989) indicates that the Castolon Local Fauna occurs stratigraphically above a basalt dated at ~23 Ma and (personal communication, 1996) that lateral equivalents of the Castolon Local Fauna are overlain by rocks dated at ~22 Ma. The late Arikareean age of the Castolon Local Fauna appears to be consistent with its radioisotopic calibration.

Stevens (1977) showed that the Texas sample, identified as *Phenacocoelus leptoscelos*, was more specialized than, e.g., *P. typus* from the Harrison Formation of Nebraska and, in fact, more specialized than any other species of *Phenacocoelus*. M. S. Stevens (*in* Busbey and Lehman, (1989) indicates that this species is, in fact, referable to the genus *Ustatochoerus*. Using the present taxonomic terminology, Stevens (1977) also indicated that the Logan Mine Local Fauna *Ustatochoerus* cf. *U. leptoscelos* is somewhat larger than the Texas form (*U. leptoscelos*) and suggested (personal communication, 1996) that the morphology of the California specimens is comparable to *Ustatochoerus* (formerly designated as *Phenacocoelus*) that has an LSD in the Runningwater Formation of Nebraska. This is consistent with the suggestion of Stevens (1977, p. 39–41) that "*P.*" *leptoscelos* differs from other members of *Phenacocoelus* in showing affinities to the derived genus, *Ustatochoerus*, then considered to be of Barstovian and younger age. In that "*P.*" *leptoscelos* is not closely allied to forms typical of the Harrison Formation and Upper Harrison beds of Nebraska, and appears to be comparable to specimens of *Ustatochoerus* in the Runningwater Formation (M. S. Stevens, personal. communication, 1996), its presence in the Logan Mine Local Fauna succession stratigraphically above clasts 18.5 m.y. old appears to be consistent with its evolutionary status, as suggested by Stevens (1977, and personal communication, 1996), and consistent with a correlation to an early (but not earliest) Hemingfordian age, as stated in Figure 3, 5. In this context, those units in the Runningwater Formation that contain an LSD of *Ustatochoerus* (= "*Phenacocoelus*") *leptoscelos* should be ~18.5 Ma old or slightly younger, but not as young as ~17.3 Ma (see below). This is consistent with a correlation to the early Hemingfordian mammal age (He1), but not necessarily to its earliest part.

Tedford (*in* Woodburne et al., 1974, p. 16–18) considered that the amphicyonid, cf. *Daphoenodon* was most similar to forms from the Upper Harrison beds of Nebraska. The Logan Mine mustelid, *Promartes*, is less useful in biochronologic correlation, with the Logan Mine remains being similar to species known from faunas of late Arikareean age.

Based on the correlations above, and on the radioisotopic dates associated with faunal- bearing units in the two areas, it appears that the Black Butte Mine Local Fauna can be attributed to interval Ar3, the Logan Mine Local Fauna to Ar4, in part. In this regard, the presence (J. Nielson, personal communication, 1996) of materials eroded from the Peach Springs Tuff in the strata with the Logan Mine Local Fauna suggests that the beds from this location (Figs. 3, 5) and above are younger than 18.5 m.y. In that case, *Michenia* cf. *M. agatensis*, which ranges ~330 m stratigraphically above the Peach Springs Tuff clasts, would extend into early Hemingfordian time. *Michenia* species are notoriously long-ranging (R. H. Tedford, personal communication, 1995). The other taxa (Fig. 3) that range above this stratigraphic position and are also of ostensible late Arikareean age elsewhere, apparently range later in time in this part of California than their individual LSDs. In summary, the only taxon of the Logan Mine Local Fauna consistent with an early Hemingfordian age as suggested by its presence above reworked Peach Springs Tuff clasts is the Hector Formation representative of *Ustatochoerus* cf. *U. leptoscelos*.

Northern Cady Mountains

Miller (1980) discussed the mammalian faunas of the northern Cady Mountains. As shown in Figures 4 and 5, these can grouped as the Lower Cady Mountains Local Fauna and the Upper Cady Mountains Local Fauna (see also Woodburne et al., 1982; Woodburne, 1991). A tuff near the local base of the Hector Formation in the northern Cady Mountains yielded a radioisotopic age of 22.9 ± 0.4 Ma (K-Ar, biotite Miller, 1980) (22.9 in Fig. 1), and the succession continues upward stratigraphically for ~100 m to a basalt flow dated at 18.6 ± 0.2 Ma (K-Ar, whole-rock; Miller, 1980) (as discussed below, the age of this basalt is considered to be between 19 and 20 m.y.). About 120 m stratigraphically above this basalt, a Hector Formation ignimbrite yielded an age of 17.9 Ma (K-Ar; sanidine; Miller, 1980; MacFadden et al., 1990b) (Fig. 2). Buesch (1994) , however, included this unit within the Peach Springs Tuff, based on feldspar geochemistry, heavy mineral suites and phenocryst assemblages. This correlation carries the interpretation that the northern Cady Mountains ignimbrite is 18.5 Ma old, revised from the age obtained by Miller (1980). This interpretation also drives a reconsideration of the age of the basalt that occurs stratigraphically ~120 m below the ignimbrite, dated at 18.6 Ma (Miller, 1980). This is discussed further in the following section.

Lower and Upper Cady Mountains Local Faunas. The Lower Cady Mountains Local Fauna occurs ~15 m above the dated basalt, which correlates to interval R4 and part of N4 of

MacFadden et al. (1990b, Fig. 7) (Fig. 4A). Miller (1980) also indicated that a skull of the oreodont *Merychyus calaminthus* occurs ~21 to 24 m below the dated basalt, correlated herein to magnetozone R3 (Fig. 4A). Woodburne et al. (1982) and Woodburne (1991) included this oreodont in the Lower Cady Mountains Local Fauna, but that is revised here to remove that taxon from this fauna and to restrict the fauna to its original stratigraphic occurrence (above the dated basalt).

Figure 4A shows the correlation of the northern Cady Mountains succession to the MPTS developed by MacFadden et al. (1990b). This correlation is revised here (Fig. 4B) due to the new faunal, radioisotopic, and stratigraphic interpretations discussed below. The revised correlation was developed as follows. Magnetozone N1 is associated with a tuff dated at 22.9 Ma. On that basis, the base of N1 appears to correlate with the base of Chron C6Bn.2 in Figure 4B. *Merychyus calaminthus* is bracketed in the Cady Mountains as between radioisotopic dates of 22.9 (northern Cady Mountains.) and 21.6 Ma (Black Butte Mine Local Fauna). Miller (1980 p. 58) reported that *M. calaminthus* of the northern Cady Mountains succession occurs ~21 to 24 m (~70 ft) below the basalt dated at 18.6 Ma. Although not part of the magnetostratigraphic column in MacFadden et al. (1990b), such a stratigraphic position would be equivalent to magnetozone R3. The oreodont is important, nevertheless, in showing that a taxon of late Arikareean age occurs stratigraphically 21 to 24 m below this basalt. In addition, Whistler (1984, p. 31) indicated that *Merychyus calaminthus* is more primitive than *Merychyus* cf. *M. minimus* from the early Hemingfordian Boron Local Fauna of the western Mojave Desert, California. This is consistent with the apparent species-level similarity between *M. calaminthus* (California) and *M. crabilli* (Nebraska), the suggested temporal correlation of the rock units in the two areas (Fig. 5), and the pre–*M. minimus* evolutionary stage for *M. calaminthus-crabilli*. These data combine to suggest that the overlying basalt is no older than ~22 Ma. The radioisotopic calibration and magnetostratigraphic correlation of this occurrence of *M. calaminthus* is consistent with a correlation to Chron C6AA, although correlation to Chron C6Ar.2 cannot be ruled out. The preferred correlation (to C6AA), however, preserves the relative spacing and thickness of reversed magnetozones R2 and R3 relative to chrons C6AAr and C6AA, respectively.

The next step in this analysis involves magnetozone N6. This zone includes the Peach Springs Tuff, originally dated here at 17.9 Ma, but revised to 18.5 Ma (see above). That revision suggests that magnetozone N6 correlates with Chron C5En in Figure 4B. Magnetozone R8 contains strata yielding the Upper Cady Mountains Local Fauna of late Hemingfordian age (below). Specifically, the presence of *Parapliohippus carrizoensis* and *Paramiolabis tenuis* in this local fauna compares favorably with the occurrence of those taxa in the Barstow Formation in a reversed magnetochron correlated to the MPTS at ~16.8 Ma in Woodburne et al. (1990). This suggests that magnetozone N7 correlates to Chron C5Dn, and R7 to C5Dr. Magnetozone R7 is questioned (Fig. 4A) only because it is a single site determination (see MacFadden et al., 1990b).

Magnetozone N3 includes the basalt that underlies the Lower Cady Mountains Local Fauna (as revised herein) and originally dated at ca. 18.6 Ma. Based on the preceding analysis, however, N3 must be older than Chron C5En. The next oldest candidate in Figure 4B is Chron C6N. This is consistent with aletomerycine artiodactyl fossils being recovered from the Lower Cady Mountains Local Fauna within 15m (50 ft) stratigraphically above the dated basalt, including the span of magnetozone R4. Again, the query at this interval (Fig. 4A) is based on its being represented by only a single site (MacFadden et al., 1990b). In Nebraska, the LSD of aletomerycine artiodactyls (*Aletomeryx*) (Fig. 5) begins with the early Hemingfordian, within Chron C5Er (see MacFadden and Hunt, this volume). Based on the correlation to the MPTS shown in Figure 4B, this suggests that the underlying basalt is no younger than 19 Ma and likely is between 19 and 20 Ma old, as shown in Figure 4B. If this is tenable (J. E. Nelson, personal communication, 1996, indicated that the basalt was extruded into a body of water so that its age is likely a minimum figure), then the correlation of magnetozones N4–R6 cannot be accounted for with respect to Figure 4B. The preferred correlation also suggests that there is an unconformity between magnetozones R3 and N3 to account for the missing portions of Chrons C6, C6A, and C6AA. This is at least plausible in that, virtually by definition, any basalt rests on an unconformity relative to underlying strata in nonmarine successions. The extent to which this possibility is true has not been documented to date in the Cady Mountains and, as summarized here, there are a number of other unanswered questions posed by the correlation in Figure 4B. Still, it seems to be the best that can be accomplished at present with respect to the available data.

Lower Cady Mountains Local Fauna. As now restricted, the Lower Cady Mountains Local Fauna includes only sparse and fragmentary remains of Merychyinae and Aletomerycinae. The correlation shown in Figures 4 and 5 is predicated on the LSD of *Aletomeryx* reported (Tedford et al., 1987) for the Runningwater Formation of Nebraska. This suggests that the Lower Cady Mountains Local Fauna is of early Hemingfordian age, consistent with its radioisotopic calibration in Nebraska as no older than ~19 Ma. (e.g., Woodburne and Swisher, 1995; MacFadden and Hunt, this volume). In fact, it is suggested by MacFadden and Hunt (this volume) that this boundary is ~18.8 m.y. old. The early Hemingfordian Boron Local Fauna also shows the presence of species of *Aletomeryx* above the Saddleback Basalt that yielded radioisotopic ages of 18.3 and 20.3 m.y. (Whistler, 1984, Fig. 60). Furthermore, in its occurrence above the dated basalt, the age of the Lower Cady Mountains Local Fauna is compatible with the ~19 m.y. age interpreted by Tedford et al. (1987, text but not Figs. 6.2, 6.3) and followed by Woodburne and Swisher (1995) for the age of the Arikareean/Hemingfordian boundary. See also MacFadden and Hunt (this volume) for a refined age of the Arikareean/Hemingfordian boundary.

Upper Cady Mountains Local Fauna. This faunal unit occurs ~65 m stratigraphically above an ignimbrite dated at 17.9 Ma (Miller, 1980) but more likely is ~18.5 Ma (see above).

The fauna is diverse, and contains the equid *Parapliohippus carrizoensis*; the beaver cf. *Anchitheriomys*; a rhino; a dog, *Tomarctus* cf. *T. hippophagus*; a rodent, *Proheteromys sulculus*; the camels *Paramiolabis tenuis* (Kelly, 1992) and cf. *Aepycamelus*; and the pronghorn antelope *Merycodus*. Tedford et al. (1987) reported an LSD for *Aepycamelus* in the Box Butte Formation of Nebraska of late Hemingfordian age. In southern California, *Parapliohippus carrizoensis* is known from the Phillips Ranch Local Fauna of the Bopesta Formation (as revised by Quinn, 1987) in the Tehachapi Mountains (Buwalda and Lewis, 1955; Quinn, 1987; Woodburne, 1991), from the Caliente Formation, Cuyama Valley (Kelly, 1995) and from the Red Division Fauna of the Barstow Formation in the Mud Hills (Miller, 1980; Woodburne et al. 1982; Tedford et al., 1987; Woodburne et al., 1990). In the Mud Hills, *P. carrizoensis* is associated with *Paramiolabis tenuis*, and occurs stratigraphically ~30 m below the Rak Tuff, dated at about 16.3 Ma.

The Red Division Fauna is correlated on paleomagnetic data (MacFadden, et al., 1990a) to Chron C5Cr, which is considered to be ~16.7 to 17.25 m.y. old (Berggren et al., 1995). MacFadden et al. (1990b, Fig. 7) (Fig. 4A) indicated that the Upper Cady Mountains Local Fauna falls within a reversed chron, R8, likely correlative with Chron C5Cr, or equivalent to the proposed age for the Red Division Fauna of the Barstow Formation. The similarity between the Upper Cady Mountains Local Fauna and the Red Division Fauna (especially on the shared occurrence of *Parapliohippus carrizoensis* and *Paramiolabis tenuis*) supports this correlation, as well, and reinforces suggestions (i.e., Woodburne and Swisher, 1995; Tedford et al., 1987, Fig. 6.2; and above) that the Box Butte Formation is correlative with these units and thus is ~17.3 m.y. old.

SUMMARY

Mammal faunas from the Hector Formation in the Cady Mountains, Mojave Desert, California, range in age from late Arikareean (Ar3) to late Hemingfordian (He2). Although taxonomically sparse, the homotaxial and other relationships discussed herein suggest that these faunal elements are comparable to those of the Harrison, Upper Harrison beds, Runningwater, and Box Butte Formations in the type sequences of Nebraska. In addition, the associated radioisotopic and paleomagnetic data help constrain the ages not only of the Californian faunas, but also those of Nebraska. In California, *Merychyus* has an LSD bracketed by tuffs having ages of 22.9 and 21.6 Ma, and this suggests that coeval parts of the Harrison Formation, Nebraska, are ~22 Ma old. An LSD of *Michenia* at ~19.2 Ma in Nebraska compares favorably with the LSD of that taxon being older than the 18.5-Ma Peach Springs Tuff and younger than a tuff dated at 21.6 Ma in California. The 18.8-Ma age adopted here for the Arikareean/Hemingfordian boundary in Nebraska (derived from the work of MacFadden and Hunt, this volume) is similar to the post-19.0-m.y. and pre-18.5-m.y. age of the early Hemingfordian Lower Cady Mountains Local Fauna (Aletomerycinae) and the likely post–Peach Springs Tuff LSD (18.5 Ma) of *Ustatochoerus* in the type Hector Formation and Runningwater Formation correlate in Nebraska. This correlation is consistent with the likely pre–Peach Springs Tuff age (18.5 m.y.) for the LSD of *Michenia* cf. *M. agatensis* in the Logan Mine Local Fauna, correlated to the Agate Springs Local Fauna in the Upper Harrison beds, Nebraska, with a basal age of ~19.2 Ma. The post-18.5-m.y. (and likely younger) age of the Upper Cady Mountains Local Fauna is consistent with its suggested correlation to Chron C5Cr in the MPTS, which is in turn consistent with a basal age of the taxonomically similar (*Aepycamelus*, initial radiation of early hypsodont equids) in late Hemingfordian faunas of western Nebraska (Box Butte Formation). The combined presence of *Paramiolabis tenuis* and *Parapliohippus carrizoensis* in the Upper Cady Mountains Local Fauna resembles those taxa in the Red Division Fauna, dated radioisotopically as older than ~16.3 Ma and on paleomagnetic grounds to from ~16.7 to 17.25 Ma, suggesting the Box Butte Formation faunas are about this old.

Arikareean and Hemingfordian mammal faunas of the Cady Mountains also yield paleoecologic information. Although mammalian taxa are not numerous until late Hemingfordian time, the ungulate component of the fauna is dominated by relatively hypsodont camels and oreodonts in the interval from Ar3 to He1, and equids are distinctive by their absence. The late Hemingfordian Upper Cady Mountains Local Fauna has the earliest occurrence of hypsodont Equidae (*Parapliohippus carrizoensis*). This taxon is the first of a number of progressively hypsodont Equidae that are recorded in late Hemingfordian and Barstovian faunas of the Mojave Desert Province, along with a diversity of other taxa comparable to, or greater than, that shown in the Upper Cady Mountains Local Fauna. This diversity, coupled with the conspicuously increased regional occurrence of deposits indicative of lacustrine sedimentation in the Barstow Formation (e.g., Woodburne et al., 1990; Woodburne, 1991), is consistent with an interpretation that the older, Arikareean and early Hemingfordian, mammal faunas lived during conditions that were substantially more xeric than those of later Hemingfordian and Barstovian times. This is compatible with the interpretations of Stevens (1977) for the relatively xeric conditions that prevailed during the span of the Black Butte Mine and Logan Mine Local Faunas in California, in agreement with inferences based on mammalian faunas from the Delaho Formation of Texas, and with the presence in the early Hemingfordian Boron Fauna of the western Mojave Desert (Whistler, 1984) and the Hackberry Local Fauna of the eastern Mojave Desert (Reynolds et al., 1995) of relatively hypsodont rodents and aletomerycine artiodactyls, but still lacking hypsodont equids.

ACKNOWLEDGMENTS

It is a pleasure to acknowledge W. R. Daily, then of Victorville, California, and C. F. Rader, Lucerne Valley, California, who donated the original specimens on which this chapter (and Woodburne et al., 1974) was based to the Department of

Earth Sciences, University of California, Riverside (UCR). Research leading to this chapter was supported by grants from the Academic Senate, UCR, and aided by numerous enthusiastic undergraduate and graduate students of the Department of Earth Sciences, as well as continued interest and help of R. H. Tedford, M. S. Stevens, and R. M. Hunt. R. H. Tedford and J. E. Nielson made important reviews of an early version of the manuscript and materially improved it.

REFERENCES CITED

Berggren, W. A., and Van Couvering, J. A., 1974, The Late Neogene; biostratigraphy, geochronology, and paleoclimatology of the last 15 million years in marine and continental sequences: Palaeogeography, Palaeoclimatology and Palaeoecology, v. 16, p. 1–216.

Berggren, W. A., Kent, D. V., Swisher, C. C., III, and Aubry, M.-P., 1995, A revised Cenozoic geochronology and chronostratigraphy, *in* Berggren, W. A., Kent, D. V., Aubry, M.-P. and Hardenbol, J., eds., Geochronology, time-scales and stratigraphic correlation; framework for an historical geology: Society of Stratigraphic Geology, Special Publication 54, p. 129–212.

Buesch, D. C., 1994, Feldspar geochemistry of four Miocene ignimbrites in southeastern Calif. and western Ariz., *in* Sherrod, D. R., and Nielson, J. E., eds., Tertiary stratigraphy of highly extended terranes, California, Arizona, and Nevada: U.S. Geological Survey Bulletin 2053, p. 55-69.

Busbey, A. B., and Lehman, T. M., eds., 1989, Vertebrate paleontology, biostratigraphy and depositional environments, latest Cretaceous and Tertiary, Big Bend area, Texas. Guidebook, Field Trip Nos. 1A, B, and C. Austin, Texas, 49th Annual Meeting, Society of Vertebrate Paleontology.

Buwalda, J. P., and Lewis, G. E., 1955, A new species of *Merychippus*: U.S. Geological Survey Professional Paper 264G, p. 147–152.

Byers, F. M., Jr., 1960, Geology of the Alvord Mountain Quadrangle, San Bernardino County, California: U.S. Geological Survey Bulletin 1089-A, 71 p.

Dalrymple, G. B., 1979, Critical tables for conversion of K-Ar ages from old to new constants: Geology, v. 7, no. 11, p. 558–560.

Dokka, R. K., and 10 others, 1991, Aspects of the Mesozoic and Cenozoic geologic evolution of the Mojave Desert: Geological Society of America, Annual Meeting Field Trip Guidebook, p. 1–43.

Glazner, A. F., 1988, Stratigraphy, structure, and potassic alteration of Miocene volcanic rocks in the Sleeping Beauty area, central Mojave Desert, California: Geological Society of America Bulletin, v. 100, p. 424–435.

Hillhouse, J. W., and Wells, R. E., 1986, Assessment of rotations in the eastern Mojave Desert, California and Arizona, from paleomagnetism of the Peach Springs Tuff [abs.]: Eos (Transactions, American Geophysical Union), v. 67, p. 922.

Hunt, R. M., Jr., 1985, Faunal succession, lithofacies, and depositional environments in Arikaree (lower Miocene) of the Hartville Table, Nebraska and Wyoming, *in* Martin, J. E., ed., Fossiliferous Cenozoic deposits of western South Dakota and northwestern Nebraska: Dakoterra, v. 2, p. 155–204.

Hunt, R. M. Jr., 1990, Taphonomy and sedimentology of Arikaree (lower Miocene) fluvial, eolian, and lacustrine paleoenvironments, Nebraska and Wyoming; a paleobiota entombed in fine-grained volcaniclastic rocks. Geological Society of America Special Paper 244, p. 69–111.

Ingersoll, R. V., and 9 others, 1996, The Mud Hills, Mojave Desert, California: Structure, stratigraphy, and sedimentology of a rapidly extended terrain. Geological Society of America Special Paper 303, p. 61–84.

Kelly, T. S., 1992, New middle Miocene camels from the Caliente Formation, Cuyama Valley badlands, California: PaleoBios, v. 13, no. 32, p. 1–22.

Kelly, T. S., 1995, New Miocene horses from the Caliente Formation, Cuyama Valley badlands, California. Natural History Museum (Los Angeles). Contributions in Science, v. 455, 33 p.

Lindsay, E. H., Opdyke, N. D., and Johnson, N. M., 1984, Blancan-Hemphillian land mammal ages and late Cenozoic mammal dispersal events: Annual Review, Earth and Planetary Science, v. 12, p. 445–488.

MacFadden, B. J., Opdyke, N. D., Swisher, C. C., III, and Woodburne, M. O., 1990a, Paleomagnetism, geochronology and possible tectonic rotation of the middle Miocene Barstow Formation, Mojave Desert, California: Geological Society of America Bulletin, v. 102, p. 478–493.

MacFadden, B. J., Woodburne, M. O., and Opdyke, N. D., 1990b, Paleomagnetism and Neogene clockwise rotation of the northern Cady Mountains, Mojave Desert of southern California: Journal of Geophysical Research, v. 95(B4), p. 4597–4608.

Miller, S. T., 1980, Geology and mammalian biostratigraphy of a part of the northern Cady Mountains, California: U.S. Geological Survey Open-File Report 80-978, p. 1–121.

Moseley, C. G., 1978, Geology of a portion of the northern Cady Mountains, Mojave Desert, California [M. S. thesis]: University of California, Riverside, 131 p.

Nielson, J. E., Lux, D. R., Dalrymple, G. B., and Glazner, A. F., 1990, Age of the Peach Springs Tuff, southeastern California and western Arizona: Journal of Geophysical Research, v. 95, p. 571–580.

Quinn, J. P., 1987, Stratigraphy of the middle Miocene Bopesta Formation, southern Sierra Nevada, California: Natural History Museum (Los Angeles) Contributions in Science, v. 393, p. 1–31.

Reynolds, R. E., Hunt, R., and Albright, B., 1995, Rhinoceros in Lanfair Valley: San Bernardino County Museum Quarterly, v. 42, no. 3, p.107–110.

Ross, T. M., 1992, Geologic and palomagnetic constraints on the timing of initiation and amount of slip on the Rodman and Pisgah faults, central Mojave Desert, California, *in* Richard, S. M., ed., Deformation associated with the Neogene eastern California shear zone, southeastern California and southwestern Arizona: San Bernardino County Museum Special Publication 92-1, p. 75–77.

Ross, T. M., 1995, North-south directed extension, timing of extension, and vertical axis rotation of the southwest Cady Mountains, Mojave Desert, California: Geological Society of America Bulletin, v. 107, p. 793–811.

Ross, T. M., Luyendyk, B. P., a Haston, R. B., 1989, Paleomagnetic evidence for Neogene clockwise tectonic rotations in the central Mojave Desert, California: Geology, v. 17, p. 470–473.

Ross, T. M., Dokka, R. K., Travis, C. J., and Woodburne, M. O., 1991, Two-phase rotational history of the SW Cady Mountains and implications for the tectonic evolution of the Mojave Desert: Geological Society of American Abstracts with Programs, v. 23, no. 2, p. A93.

Schultz, C. B., and Falkenback, C. H., 1947, Merychyinae, a subfamily of oreodonts: American Museum of Natural History Bulletin, v. 88, article 4, p. 157–286.

Shotwell, J. A., Bowen, R. G., Gray, W. L., Gregory, D. C., Russell, D. E., and Taylor, D. W., 1963, The Juntura Basin; studies in early history and paleoecology: American Philosophical Society Transactions, n.s., v. 53, p. 5–77.

Stevens, M. S., 1977, Further study of Castolon local fauna (early Miocene) Big Bend National Park, Texas: The Pearce-Sellards Series, Texas Memorial Museum, no. 28, p. 1–69.

Stevens, M. S., Stevens, J. B., and Dawson, M. R., 1969, New early Miocene formation and vertebrate local fauna, Big Bend National Park, Brewster County, Texas: The Pearce-Sellards Series, Texas Memorial Museum, no. 15, p. 1–35.

Tedford, R. H., and 8 others, 1987, Faunal succession and biochronology of the Arikareean through Hemphillian interval (late Oligocene through earliest Pliocene epochs) in North America, *in* Woodburne, M. O., ed., Cenozoic mammals of North America, geochronology and biostratigraphy: Berkeley, University of California Press, p. 153–210.

Tedford, R. H., Swinehart, J. B., wisher, C. C., III, Prothero, D. R., King, S. A., and Tierney, T. E., 1995, The Whitneyan-Arikareean transition in the High Plains, *in* Prothero, D., and Emry, R., eds., The terrestrial Eocene-Oligocene transition in North America: New York, Cambridge University Press, p. 278–311.

Wells, R. E., and Hillhouse, J. W., 1989, Paleomagnetism and tectonic rotation of the lower Miocene Peach Springs Tuff: Colorado Plateau, Arizona, to Barstow, California: Geological Society of America Bulletin v. 101, p. 846–863.

Whistler, D. P., 1984, An early Hemingfordian (early Miocene) fossil vertebrate fauna from Boron, western Mojave Desert, California: Natural History Museum (Los Angeles) Contributions in Science, v. 355, p. 1–36.

Williamson, D. A., 1980, The geology of a portion of the eastern Cady Mountains, Mojave Desert, California [M.A. thesis]: University of California, Riverside, 141 p.

Wood, H. E., III, Chaney, R. W., Clark, J., Colbert, E. H., Jepsen, G. L., Reeside, J. B., and Stock, C., 1941, Nomenclature and correlation of the North American continental Tertiary: Geological Society of America Bulletin, v. 52, p. 1–48.

Woodburne, M. O., 1991, The Mojave Desert Province, *in* Woodburne, M. O., Reynolds, R. E., and Whistler, D. P., eds., Inland Southern California; the last 70 million years: San Bernardino County Museum Association Quarterly, v. 38, no. 3–4 p. 60–77.

Woodburne, M. O., and Swisher, C. C., III, 1995, Land mammal high resolution geochronology, intercontinental overland dispersals, sea-level, climate, and vicariance; *in* Berggren, W. A., Kent, D. V., Aubry, M.-P., and Hardenbol, J., eds., Geochronology, time-scales and stratigraphic correlation; framework for an historical geology: Society of Stratigraphic Geology Special Publication 54, p. 335–364.

Woodburne, M. O., Tedford, R. H., Stevens, M. S., and Taylor, B. E., 1974, Early Miocene mammalian faunas, Mojave Desert, California: Journal of Paleontology, v. 48, p. 6–26.

Woodburne, M. O., Tedford, R. H., and Miller, S. T., 1982, Stratigraphy and geochronology of Miocene strata in the central Mojave Desert, California, *in* Cooper, J. D., compiler, Geologic Excursions in the California Desert, volume and guidebook: 78th Annual Meeting, Cordilleran Section, Geological Society of America, p. 47–64.

Woodburne, M. O., Tedford, R. H., and Swisher, C. C., III, 1990, Lithostratigraphy, biostratigraphy and geochronology of the Barstow Formation Mojave Desert, southern California: Geological Society of America Bulletin, v. 102, p. 459–477.

MANUSCRIPT ACCEPTED BY THE SOCIETY DECEMBER 5, 1997

Printed in U.S.A.

Index

[Italic page numbers indicate major references]

D

E

F

G

M

N

O

P

R

S

Printed in U.S.A.

The Geological Society of America
3300 Penrose Place • P.O. Box 9140 • Boulder, Colorado 80301

Contents

ISBN 0-8137-2325-6
9 780813 723259