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PALEOENVIRONMENTS AND TECTONIC SIGNIFICANCE OF THE UPPER JURASSIC MORRISON/LOWER CRETACEOUS CLOVERLY FORMATIONS, BIGHORN BASIN, WYOMING

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Paleoenvironments and tectonic significance of the Upper Jurassic Morrison/Lower Cretaceous Cloverly formations, Bighorn Basin, Wyoming

by

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TABLE OF CONTENTS

LIST OF FIGURES

Page

LIST OF TABLES

INTRODUCTION

Method of Study

The Morrison and Cloverly formations were studied in detail along the northeastern margin of the Bighorn Basin, northern Wyoming (Figure 1). Complete detailed descriptions of the alluvial architectural elements of the formations were made north of the town of Shell near Horse Creek and in the badlands in the area of confluence of Beaver Creek and Cedar Creek just northwest of Horse Creek. Complete sections of Morrison and Cloverly were also described in detail in the Sheep Mountain region, north of Greybull. These sections were utilized in a general examination of other outcrops of Morrison and Cloverly in the Bighorn Basin and compared with work done by previous investigators in an attempt to: 1) determine the environments of deposition of the two formations, 2) clarify the established formal and informal stratigraphic nomenclature used in the literature, and 3) develop the sedimentary tectonic model proposed in this paper.

Special attention was paid to a more detailed basin-wide analysis of the "Greybull sandstone" of the upper Cloverly. This was done because of the potential of the Greybull as a reservoir for hydrocarbons, water, and perhaps uranium, but whose exploitation has been hampered by a poor

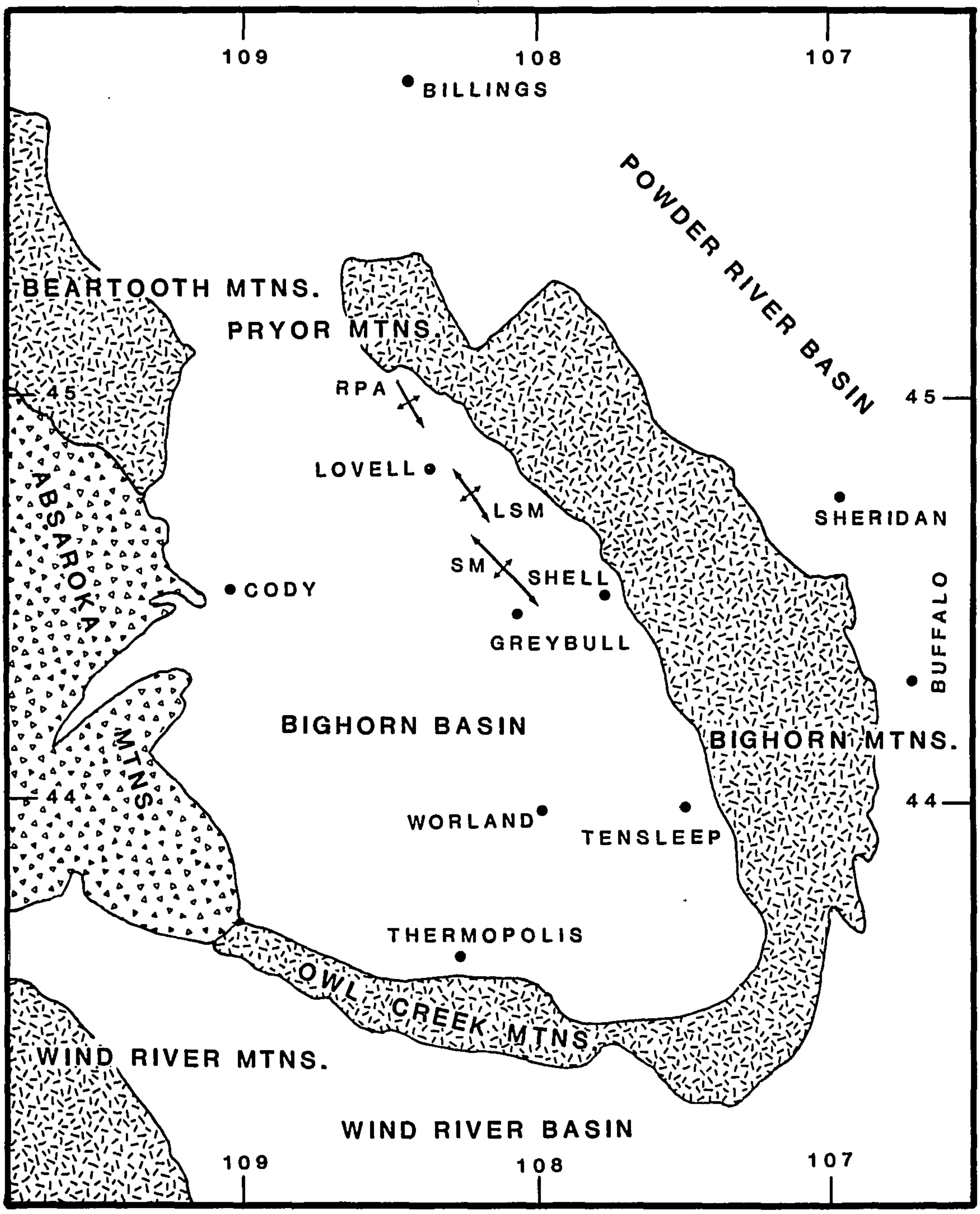


Figure 1. Map of Bighorn Basin, northern Wyoming; RPA = Red Pryor anticline, LSM = Little Sheep Mountain, SM = Sheep Mountain

understanding of its depositional environments. The following is a summary of the mineralogy, stratigraphy, sedimentology and tectonic significance of the Morrison and Cloverly formations in the northeastern part of the Bighorn Basin with generalizations about the rest of the basin based on reconnaissance field notes and previous work of others.

Geologic Setting

The Upper Jurassic Morrison and Lower Cretaceous Cloverly formations of the Bighorn Basin of northern Wyoming are a part of an extensive sequence of fluvial and lacustrine sediments which mark a break in marine sedimentation in the western interior of North America. These formations were deposited in a foreland basin which formed as a result of an eastward dipping Andean type subduction complex to the west (Burchfield and Davis, 1975; Dickinson, 1976).

The sediments accumulated in a clastic wedge, thinning from eastern Idaho, western Wyoming, and Montana to central Wyoming then thickening again towards the Black Hills region (Figure 2). Sediments were derived from the Sevier fold-thrust belt and volcanic vents in eastern and central Idaho, the craton to the east and from intraforeland basin uplifts and volcanic vents.

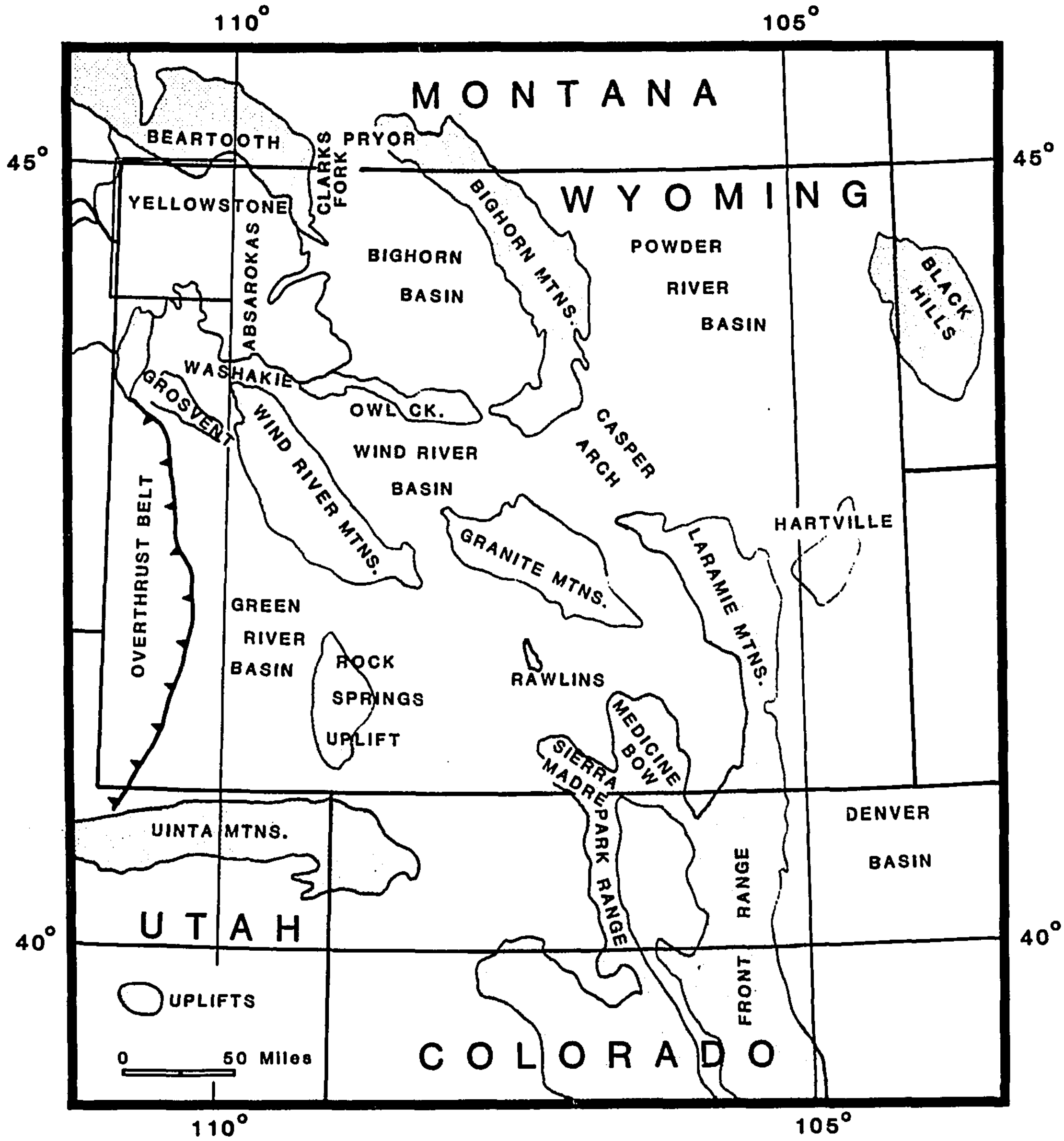


Figure 2. Map of western interior

PREVIOUS WORK

General Background

Geologic investigations of the Upper Jurassic/Lower Cretaceous nonmarine sequence of the western interior date back to the end of the last century. The initial studies were by paleontologists and stratigraphers, the former of which were interested in dinosaur remains preserved in the mudstones and sandstones and the later in more of the general lithologies and regional extent of the units. The stratigraphic studies served to establish formal stratigraphic nomenclature within the various intermontane basins of Wyoming and Colorado. However, correlations between basins, and even within basins, proved to somewhat tenuous, in part, because of the lack of detailed studies necessary to truly understand the fluvial character of the interval with its rapid lateral and vertical variability. The general lack of diagnostic fossils and datable horizons and the presence of multiple source areas also compounded the problems. The difficulties are illustrated by the resignation of many authors to simply map the formations as "Morrison and Cloverly, Undifferentiated" (Andrews *et al.*,1947; Love and Christiansen, 1985; Rioux, 1958; Weitz *et al.*, 1954).

By the late 1950s and early 1960s petrologic studies and paleocurrent studies were being undertaken in Wyoming and Utah in an attempt to determine paleodrainage patterns and source areas of the sediments (e.g., Cadigan, 1967; Chisholm, 1963; Craig _et al_., 1955; MacKenzie and Ryan, 1962; Moberly, 1960; 1962). Most recently the Upper Jurassic/Lower Cretaceous nonmarine interval in Wyoming, Idaho, and Utah has been studied in an attempt to understand the tectonics of the early Sevier foreland basin and associated fold-thrust belt (e.g., Beck _et al_., 1986; Dickinson, 1976; Jordan, 1981; Wiltschko and Dorr, 1983). Unfortunately, these studies have had to rely entirely or in part on the less detailed regional studies which predate the mid-1960s. As a result, correlations between basins still remain confused because of the complexity of the basic fluvial architectural elements within the nonmarine interval.

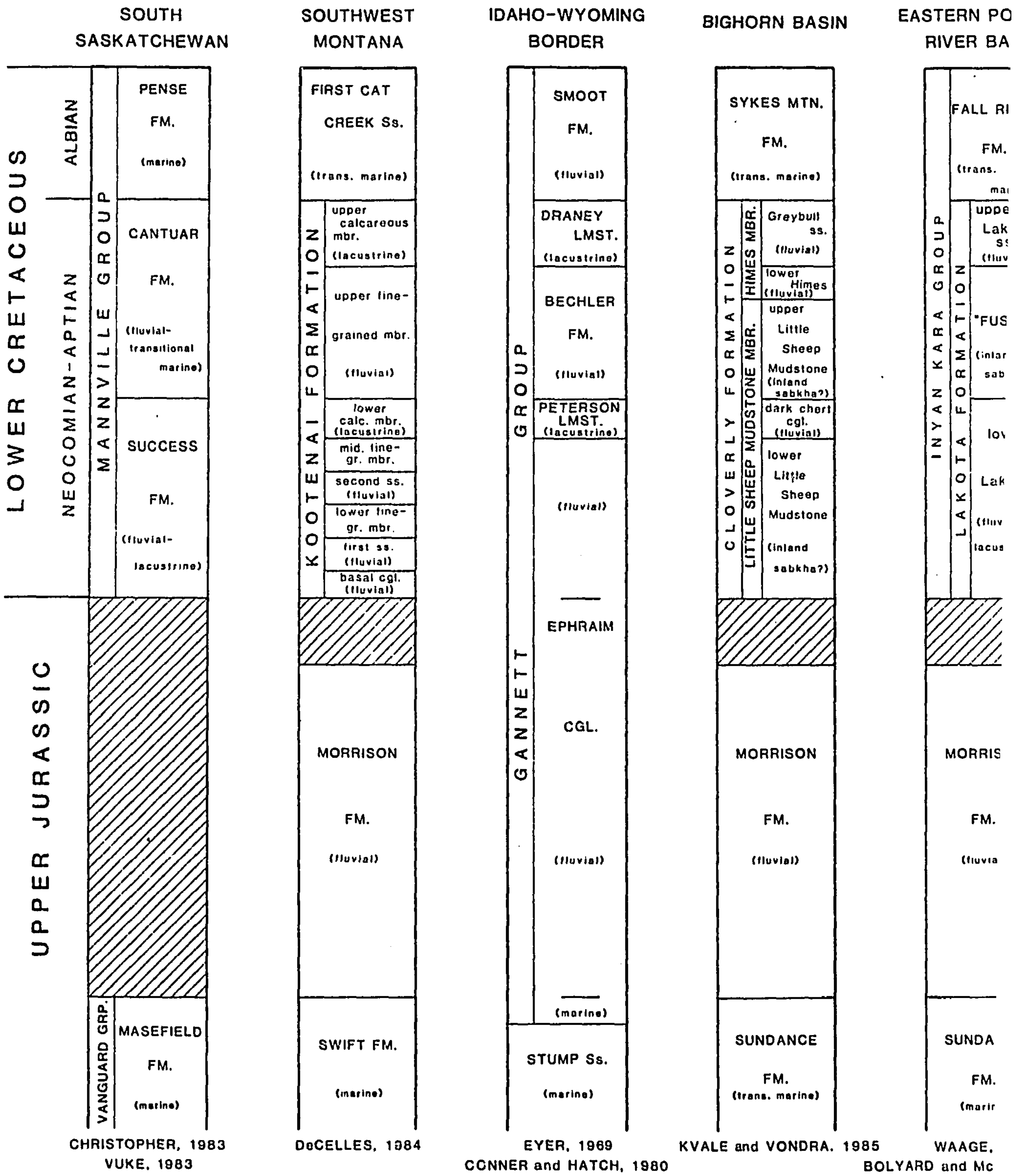
Morrison Formation

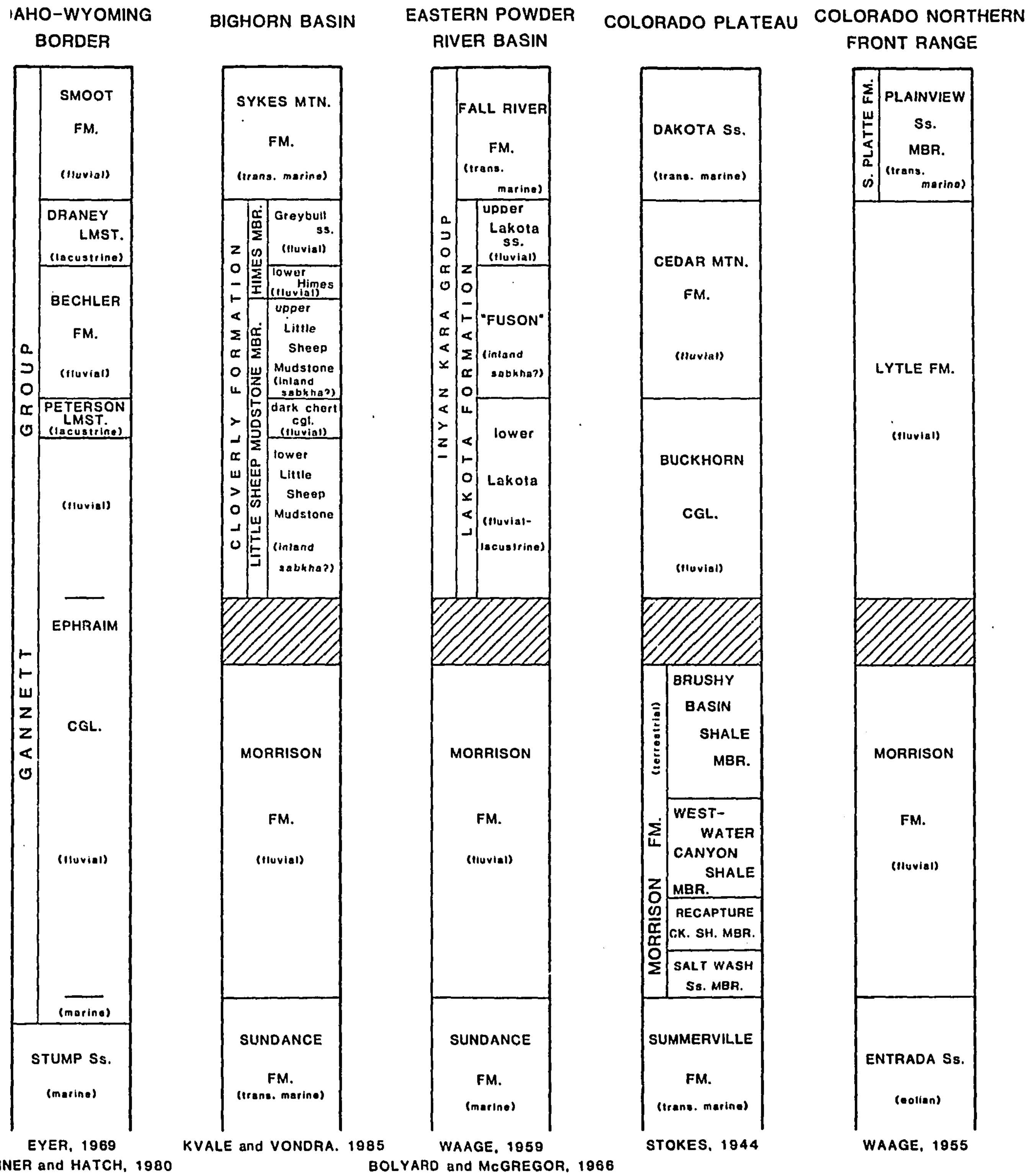
The term Morrison was first used by Cross (1894), but Eldridge (1896) was the first to formally describe and define the formation for exposures north of Morrison, Colorado. Eldridge defined the formation to include the Atlantosaurus Clays of Marsh and an overlying thin conglomerate with "saurian" remains and a sequence of interbedded sandstones and marls. The upper contact of the unit was revised by Lee (1920) by placing it at the base of the "saurian" conglomerate after Knowlton (1920) described a

Dakota flora collected from the overlying interbedded sandstones and marls. The type section was later redefined by Waldschmidt and LeRoy (1944) to consist of the outcrops in the roadcuts along the West Alameda Parkway where it passes through the Dakota hogback. The Morrison Formation is one of the most widespread units in the western interior (Table 1). The name has been applied to the Upper Jurassic nonmarine deposits in Colorado, Utah, New Mexico and throughout much of Wyoming and South Dakota. However, in westernmost Wyoming the lower part of the Ephraim Conglomerate has been equated with the Morrison (Eyer, 1969; Furer, 1970; Wanless et al. 1955). In western South Dakota along the southeast flank of the Black Hills uplift, the Unkapapa Sandstone has been correlated with the lower part of the Morrison (Imlay, 1947).

N. H. Darton was the first to describe Morrison sediments in northern Wyoming. He did so for units in the Powder River Basin (1901) and later in the Bighorn Basin (1906). He based his correlations on the similarity of these units to Morrison deposits in the Colorado Front Range, which he had studied. He described the formation as consisting of terrestrial deposits of sandy shales and claystones with interbedded sandstones and freshwater limestones. The depositional environments of the Morrison are poorly understood. Some workers have suggested the

Table 1. Western interior stratigraphic correlation chart


SOUTH SASKATCHEWAN
SOUTHWEST MONTANA
IDAHO-WYOMING BORDER
BIGHORN BASIN
EASTERN PO
RIVER BA
LOWER CRETACEOUS
ALBIAN
NEOCOMIAN-APTIAN
UPPER JURASSIC
MANNVILLE GROUP
PENSE FM. (marine)
CANTUAR FM. (fluvial-transitional marine)
SUCCESS FM. (fluvial-lacustrine)
VANGUARD GRP.
MASEFIELD FM. (marine)
FIRST CAT CREEK Ss. (trans. marine)
KOOTENAI FORMATION
upper calcareous mbr. (lacustrine)
upper fine-grained mbr. (fluvial)
lower calc. mbr. (lacustrine)
mid. fine-gr. mbr.
second ss. (fluvial)
lower fine-gr. mbr.
first ss. (fluvial)
basal cgl. (fluvial)
MORRISON FM. (fluvial)
SWIFT FM. (marine)
SMOOT FM. (fluvial)
GANNETT GROUP
DRANEY LMST. (lacustrine)
BECHLER FM. (fluvial)
PETERSON LMST. (lacustrine)
(fluvial)
EPHRAIM
CGL.
(fluvial)
(marine)
STUMP Ss. (marine)
SYKES MTN. FM. (trans. marine)
CLOVERLY FORMATION
HIMES MBR.
Greybull ss. (fluvial)
lower Himes (fluvial)
LITTLE SHEEP MUDSTONE MBR.
upper Little Sheep Mudstone (inland sabkha?)
dark chert cgl. (fluvial)
lower Little Sheep Mudstone (inland sabkha?)
MORRISON FM. (fluvial)
SUNDANCE FM. (trans. marine)
FALL RI
FM.
(trans.
mar
INYAN KARA GROUP
LAKOTA FORMATION
uppe
Lak
(fluv
"FUS
(inlar
sab
lov
Lak
(fluv
lacus
MORRIS
FM.
(fluvia
SUNDA
FM.
(marir
CHRISTOPHER, 1983
VUKE, 1983
DeCELLES, 1984
EYER, 1969
CONNER and HATCH, 1980
KVALE and VONDRA, 1985
WAAGE,
BOLYARD and Mc



IAHO-WYOMING BORDER
BIGHORN BASIN
EASTERN POWDER RIVER BASIN
COLORADO PLATEAU
COLORADO NORTHERN FRONT RANGE
SMOOT FM. (fluvial)
DRANEY LMST. (lacustrine)
BECHLER FM. (fluvial)
PETERSON LMST. (lacustrine)
(fluvial)
EPHRAIM CGL. (fluvial)
(marine)
GANNETT GROUP
STUMP Ss. (marine)
SYKES MTN. FM. (trans. marine)
CLOVERLY FORMATION
HIMES MBR.
Greybull ss. (fluvial)
lower Himes (fluvial)
LITTLE SHEEP MUDSTONE MBR.
upper Little Sheep Mudstone (inland sabkha?)
dark chert cgl. (fluvial)
lower Little Sheep Mudstone (inland sabkha?)
MORRISON FM. (fluvial)
SUNDANCE FM. (trans. marine)
FALL RIVER FM. (trans. marine)
INYAN KARA GROUP
LAKOTA FORMATION
upper Lakota ss. (fluvial)
"FUSON" (inland sabkha?)
lower Lakota (fluvial-lacustrine)
MORRISON FM. (fluvial)
SUNDANCE FM. (marine)
DAKOTA Ss. (trans. marine)
CEDAR MTN. FM. (fluvial)
BUCKHORN CGL. (fluvial)
MORRISON FM. (terrestrial)
BRUSHY BASIN SHALE MBR.
WEST-WATER CANYON SHALE MBR.
RECAPTURE CK. SH. MBR.
SALT WASH Ss. MBR.
SUMMERVILLE FM. (trans. marine)
S. PLATTE FM.
PLAINVIEW Ss. MBR. (trans. marine)
LYTLE FM. (fluvial)
MORRISON FM. (fluvial)
ENTRADA Ss. (eolian)
EYER, 1969
ONNER and HATCH, 1980
KVALE and VONDRA, 1985
WAAGE, 1959
BOLYARD and McGREGOR, 1966
STOKES, 1944
WAAGE, 1955

presence of large lakes and tropical climates during Morrison times (Colbert, 1961; 1980; Dodson *et al*., 1980). These interpretations were based on what they deemed a necessary habitat for large sauropods, the remains of which had been found within the Morrison. The Upper Jurassic/Lower Cretaceous nonmarine deposits have long been known as important fossil-bearing strata in the western interior of North America. Extensive fossil reptile collections have been obtained from the Morrison Formation and to a much lesser degree the Cloverly Formation and their equivalents in Wyoming, Utah, Montana and Colorado (Brown, 1934; 1937; Colbert, 1980; Dodson *et al*., 1980; Hallam, 1975; Moberly, 1960; Ostrom, 1970). Unfortunately, controversy exists in the literature concerning climatic conditions and depositional environments which existed during Morrison times. Mook (1916) suggested that the Morrison reflected deposition on a broad alluvial plain with lakes only locally present. Stokes (1944) in his report on the Morrison and related deposits in and adjacent to the Colorado Plateau suggested that semiarid conditions developed through Morrison time. He indicated that the Salt Wash Member constituting the lower part of the Morrison in Utah represented lacustrine and fluvial conditions which gave way to more ephemeral conditions recorded by the Brushy Basin Member of the Upper Morrison. Stokes cites the abundance of volcanic ash, the lack of large channel systems, and the predominance of fine-grained deposits as evidence for semiarid conditions. Later workers have also suggested an increase in aridity through Morrison time. Bilbey and others (1974) suggested this for northeastern Utah as did Brown (1934) for northcentral Wyoming. Other workers have indicated just the opposite for the Colorado Plateau region (Craig et al., 1955) and central Colorado (Brady, 1969).

Floral and faunal evidence suggests that some regions of the Morrison depositional area were favorable enough to support abundant vegetation. Coal deposits are present in the Morrison in western Montana (Peterson, 1972), and remains of aquatic animals such as turtles, crocodiles and fish have been reported (Moberly, 1960). However, concentrated remains of these animals are rare (Dodson et al., 1980). Moberly (1960) has suggested that the Morrison paleoclimate was seasonal and used the Gran Chaco region of South America as a modern analog. Faunal remains also suggest a seasonal climate (Dodson et al., 1980).

Cloverly Formation

Most of the geologic investigations of the Lower Cretaceous nonmarine interval have centered in the Black Hills region in South Dakota, the Powder River and Bighorn basins in Wyoming and the Colorado Plateau in Utah. The term Lakota Formation (from the Black Hills and Powder River Basin terminology) is perhaps the most widely used of the formalized terms for this sequence, but it is the Lakota's Bighorn Basin equivalent, the Cloverly Formation, which has been studied in the most detail.

The term Cloverly Formation was first introduced by Darton (1904) for a sequence of conglomeratic sandstones, sandstones and claystones which overlie the Morrison Formation in the Bighorn Basin. He designated the type area to be in the vicinity of the now abandoned Cloverly post office north of the town of Shell, Wyoming. Darton believed the Cloverly to be roughly correlateable with the Lakota Formation and Fuson Shale, which he had defined in the Black Hills region (Darton, 1901). However, Darton lacked what he considered to be definitive proof of their stratigraphic equivalency. Darton also defined a third formation; the Minnewaste Limestone. This unit is a fresh water limestone which is confined to the southern Black Hills region. It occurs stratigraphically between the Lakota Formation and Fuson Shale. Darton defined the transitional marine Dakota Sandstone for units overlying the Fuson Shale. His Dakota Sandstone was later demonstrated to be older than the type Dakota Sandstone exposed in northeast Nebraska and was renamed the Fall River Formation by Russell (1928). A more detailed study of the Lower Cretaceous strata in the Black

Hills region prompted Waage (1959b) to redefine the terminology of Darton and to combine the Lakota, Minnewaste and the Fuson into a single unit, the Lakota Formation, Waage also proposed that the newly defined Lakota be combined with the Fall River Formation into a redefined Inyan Kara Group. The Inyan Kara Group was originally defined by Rubey (1931) and included the Lakota Formation (Darton's Lakota and Minnewaste Limestone) and the Fuson Shale. Unfortunately, the terms Lakota, Cloverly, Sykes, Mountain (formation which overlies the Cloverly in the Bighorn Basin and named by Moberly in 1960), Dakota, Fuson and Fall River are often used interchangeably especially in petroleum related studies (e.g., Shelton, 1972; Stone, 1983) and the terminology remains confused.

The equivalent of the Cloverly/Lakota Formation along the western Wyoming border was referred to as the Gannett Group by Mansfield and Roundy (1916) (named for outcrops in the Gannett Hills region of the Idaho-Wyoming border). The group includes the basal Ephraim Conglomerate which is overlain by the Peterson Limestone, Bechler Formation, Draney Limestone and Smoot Formation. Most workers consider the basal Ephraim Conglomerate to be equivalent to the Morrison Formation to the east (Eyer, 1969; Furer, 1970).

The exact contact relationships between the Cloverly Formation (and lateral equivalents) and the underlying

Morrison Formation and the overlying marine to transitional marine sequence are controversial. Darton (1906) found no evidence of an unconformity between the Morrison and Cloverly in the Bighorn Basin, yet reported the contact to be somewhat abrupt. Waage (1959b) noted that the Morrison-Lakota contact in the Black Hills region does not interfinger while Bolyard and McGregor (1966) considered the contact to be unconformable. Mirsky (1962) considered the Morrison- Cloverly contact along the south flank of the Bighorn Mountains to represent a major hiatus. Moberly (1960, 1962) concluded that the Morrison-Cloverly contact was generally conformable and only locally unconformable in the northern Bighorn Basin. The upper contact of the Cloverly (Lakota) with the overlying transitional marine units was considered to be disconformable in the Bighorn Basin by Moberly (1962), a paraconformity by Curry (1962) in central Wyoming and a "transgressive disconformity" in the Powder River Basin by Waage (1959b). However, recent work by Kvale and Vondra (1985b) has shown that the contact is conformable in the Bighorn Basin.

The exact nature of the depositional environments of the Cloverly and equivalents are still uncertain because detailed sedimentological analyses of the sediments are generally lacking. A fluvial and lacustrine origin for the Cloverly is generally accepted by most workers (e.g., Eyer, 1969; Furer, 1970; Kvale and Vondra, 1985a and b; Mirsky, 1962; Moberly, 1960; 1962; Stokes, 1944). The Peterson and Draney limestones of western Wyoming are considered to have been deposited in fresh water lakes (Eyer, 1969; Brown and Wilkinson, 1981). However, there is evidence that the uppermost Draney Limestone in southwestern Montana may have been intermittently connected to the Cretaceous sea to the north by tidal inlets (Holm et al., 1977).

The best attempt at interpreting the Early Cretaceous regional climatic conditions is that by Moberly (1960; 1962). His interpretations however are based mostly on mineralogy and preliminary paleomagnetic studies. A more localized and time restricted investigation of the climatic conditions present during deposition of the Draney Limestone was made by Brown and Wilkinson (1981). Their studies were based on the composition, fabric and paleontology of the unit. Both of these studies suggest a temperate Early Cretaceous climate with Moberly inferring a seasonal drying.

The Lower Cretaceous fluvial deposits are represented by the Buckhorn Conglomerate and the Cedar Mountain Formation in the Colorado Plateau in Utah, the Burro Canyon Formation along the eastern margin of the Colorado Plateau and the flanks of the Uncompaghre Uplift in western Colorado, and the Lytle Formation or Lytle Sandstone Member of the Purgatoire Formation along the Front Range in central

Colorado. Stokes (1944) introduced the Buckhorn Conglomerate and the Cedar Mountain Formation for units which had previously been included in the uppermost part of the Morrison in the Colorado Plateau. The term Buckhorn Conglomerate was applied to a 20 to 30 foot thick chert pebble and cobble conglomerate exposed at Cedar Mountain. Buckhorn Flat on the southwest flank of Cedar Mountain in Emery County Utah was designated as its type section. The variegated shales overlying the conglomerate were defined as the Cedar Mountain Formation. Stokes selected the exposures along the southwest flank of Cedar Mountain just north of Buckhorn Reservoir as its type section.

Stokes correlated the Buckhorn with the lower part of the Ephraim Conglomerate in southeastern Idaho and with the "lower conglomerate" now known as the Pryor Conglomerate (Moberly, 1960) of the Cloverly Formation in Wyoming. He originally included the Buckhorn and Cedar Mountain in the Lower Cretaceous on the basis of little or no evidence (Stokes, 1944). However, in 1952 Stokes demonstrated an Early Cretaceous age with fossil evidence for the Cedar Mountain Formation and the Burro Canyon Formation, a unit he and Phoenix had named in 1948 (Stokes and Phoenix, 1948). They applied the term Burro Canyon to strata which had previously been described as the "Post McElmo" by Coffin (1921). The unit was considered to be equivalent to the

Cedar Mountain but was arbitrarily restricted in use to the region of the Colorado Plateau to the east and south of the Colorado River.

The Buckhorn Conglomerate rests unconformably on the Morrison Formation in the Colorado Plateau. It is a thin sheet-like deposit of great aerial extent consisting of chert and occasional silicified wood pebbles (Stokes, 1944). The chert pebbles contain fossils which indicate that they were derived from Carboniferous and Permian carbonates while the silicified wood is similar in appearance to that occurring in the Triassic strata. The grain size of the pebbles and the thickness of the deposit decrease from west to east suggesting a transport direction to the east. Stokes (1944) suggests that the Buckhorn was deposited and periodically reworked by sheet floods in an arid or semiarid climate where the winnowing effect of the wind played an important role in concentrating the pebbles.

The varicolored mudstones of the Cedar Mountain Formation in the Colorado Plateau are quite similar to those of the Brushy Basin Member of the Morrison. They may have been derived from reworked Morrison to the west and southwest (Stokes, 1944). Stokes suggests that they may have been deposited under desert conditions. Fine sediments of volcanic origin were brought in and reworked by eolian processes. Ephemeral lakes or inland sabkhas were undoubtedly important sites of deposition.

Following the publication of reports of the Hayden and the King surveys the oldest Cretaceous (pre-Benton Cretaceous) strata exposed along the Front Range of central Colorado were included in the Dakota Group. At that time, the term group was used interchangeably with the term formation. Consequently, after stratigraphic nomenclature was standardized, the Dakota Group became the Dakota Formation or Dakota Sandstone without any revision in its original meaning.

Gilbert, in 1897, informally subdivided the Dakota in southcentral Colorado into two units while Hills (1899, 1900) noted that it consisted of three - a basal conglomeratic sandstone, a middle shale and an upper fine grained sandstone with some interbedded shale. The threefold division of the Dakota was recognized by other geologists. Darton (1905) and Stanton (1905) discovered fossils of Early Cretaceous age in exposures of the middle shale near La Junta and Canon City. This created some difficulty in stratigraphic taxonomy since, until then, the Dakota Formation was considered to be Late Cretaceous in age on the basis of fossil plants which had been collected from the upper sandstone.

In 1912, Stose formally subdivided the Dakota Formation. He named the Early Cretaceous part, which included the basal sandstone and middle shale, the Purgatoire Formation after exposures along the Purgatoire River in Prowers County. The upper sandstone of the original three-fold division of Hills (1899) retained the name Dakota Sandstone.

Finlay (1916) subsequently assigned member status to the two units now comprising the Purgatoire Formation. The sandstone and shale became the Lytle Sandstone and the Glencairn Shale members, respectively. It is important to note that the Glencairn is marine and is separated from the nonmarine Lytle by a disconformity termed a transgressive disconformity by Waage (1955, 1959a).

The term Dakota Formaton continued to be used in the northern Front Range. Lee's (1923, 1927) study of the northern foothills resulted in the division of the Dakota into five informal units consisting of a lower sandstone, lower shale, middle sandstone, upper shale, and upper sandstone. Lee elevated the Dakota to group status. This terminology remained unchanged until 1955 when Waage revised the Dakota Group to the twofold division of a lower nonmarine unit and an upper marginal marine to marine unit characteristic of the group elsewhere in Colorado. Waage (1955) applied the term Lytle Formation to the lower nonmarine unit of sandstone, conglomeratic sandstone and variegated claystone and the new name South Platte Formation to the upper marine unit of alternating sandstone and dark colored shale. Waage (1955, 1959a) emphasized the occurrence of a prominent disconformity between the two formations. Waage also discussed the contact relationship between the Lytle Formation and the underlying Morrison and clarified the confusion concerning Eldridge's (1896) original definition of the Morrison and the misinterpretation of the position of its contact with the overlying Dakota Group (Lee, 1920; Stokes, 1944; Waldschmidt and LeRoy, 1944).

Waage (1955) demonstrated that the Lytle Formation undergoes a lateral facies change northward to an alternating sequence of claystones and lenticular sandstones. Where claystones of the Lytle rest directly on Morrison claystones, the contact between the two formations, although disconformable, is difficult to delineate.

FACIES AND ENVIRONMENTS OF DEPOSITION

The Morrison and Cloverly Formations crop out along the margins of the Bighorn Basin. At Beaver Creek near its confluence with Cedar Creek and at Horse Creek along the west flank of the Bighorn Mountains, the Morrison conformably overlies the marine Jurassic Sundance Formation. It can be informally divided into two units: 1) a marginal coastal plain supratidal to intertidal depositional sequence and 2) a more distal fluvial coastal plain deposit.

The Cloverly Formation as defined by Moberly (1960) consists of three members, in ascending order: 1) the Pryor Conglomerate, 2) the Little Sheep Mudstone Member, and 3) the Himes Member. According to Moberly, the Pryor Conglomerate is a thick massive to trough cross bedded, black chert bearing conglomeratic quartz arenite derived from uplifted Paleozoic strata to the west. He noted that it is of local extent and found in outcrops at the northern end of the Bighorn Basin adjacent to the Pryor Mountains where it attains a thickness of more than 10m (Moberly, 1956). He also noted that the Pryor thickens to the west and thin to the south and southeast where it forms thinner lenticular conglomeratic sandstone sequences. This member, according to Moberly, disconformably overlies the Morrison Formation to the north in the Pryor Mountain region but is not present much south of that area (Moberly, 1956). This unit is very important to Moberly's distinction of the contact between the Morrison and Cloverly. He (1960, p.1145) defines the basal contact of the Cloverly Formation at the "base of the lowest beds which either show evidence of significant additions of volcanic debris or contain pebbles or granules of black chert". The usefulness of designating the Pryor Conglomerate as a formal unit is questioned. First of all, dark chert-bearing conglomerates occur throughout the lower Cloverly and its equivalents in Wyoming and the western interior and are not necessarily confined to the basal contact. These units also do not form a continuous sheet deposit but appear to be mostly isolated patches whose paleocurrent indicators suggest multiple source areas (discussed more fully under the Regional Stratigraphy and partitioning sections). Secondly, fine grained dark chert fragments are found throughout the Morrison Formation in the northeastern part of the Bighorn Basin. In the southern end of the basin, conglomeratic sandstones have been reported from middle Morrison exposures by Mirsky (1962) beneath his thick lowest Cretaceous black chert sequence which he terms the Otter Creek member of the Cloverly. Pebble sized dark chert is found at an even lower stratigraphic position in the coquina of the upper part of the Sundance Formation. This demonstrates the difficulty of differentiating between the Morrison and Cloverly on the basis of the first occurrence of black chert even within the Bighorn Basin. The Morrison-Cloverly contact in the basin can be approximated however, as Moberly suggests, on the basis of the lowest beds which contain significant amounts of montmorillonite or silica and barite nodules. The color change from the red-green banding of the Morrison to the variegated purples and grays of the Cloverly is also useful in differentiating between the two formations where floodplain deposits from major Cloverly channels are not present.

Overlying the Pryor (used in the sense of Moberly, 1960), or directly overlying the Morrison where the Pryor is absent is the Little Sheep Mudstone Member. Briefly, it is composed of massive variegated mudstones which are generally noncalcareous and bentonitic. Interbedded with the bentonitic mudstones are devitrified tuffs including a prominent unit in the upper part of the Little Sheep Mudstone. The member weathers into low rounded hills which are laced with abundant chalcedony, calcite and barite concretions and veinlets. Lenticular channel sandstones and conglomeratic sandstones, some quite thick, composed of quartz arenites and quartz wackes are interbedded with the mudstones.

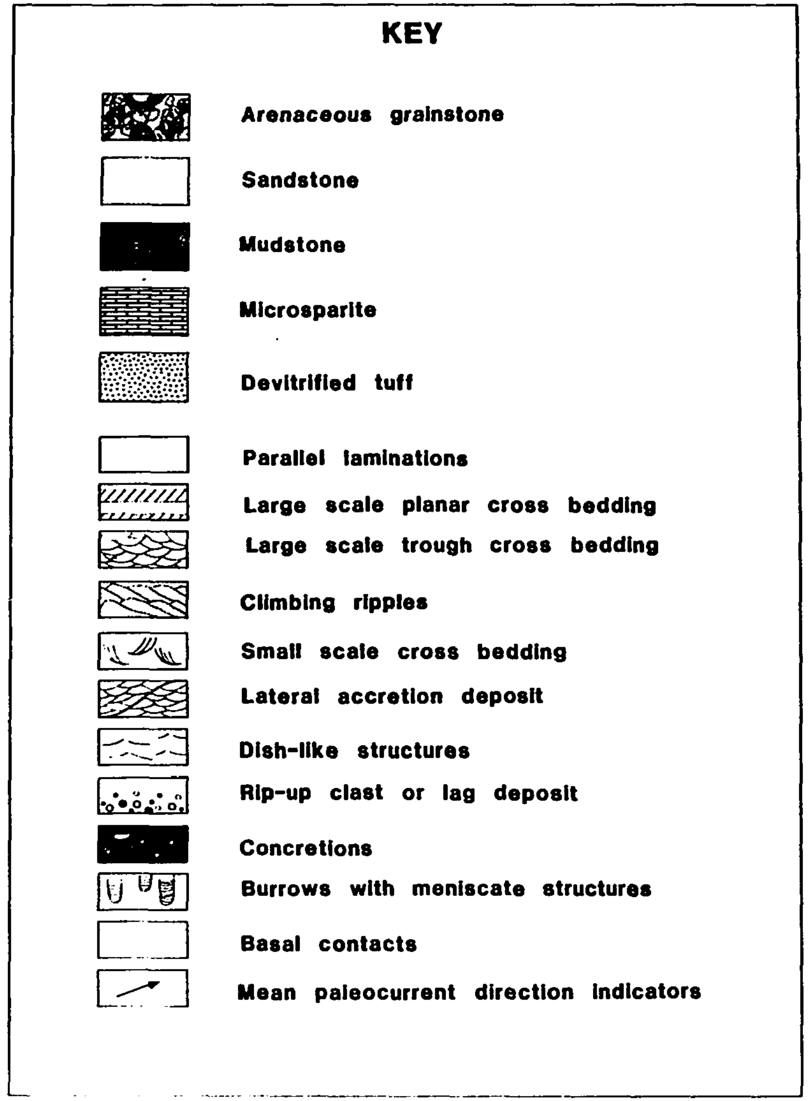
The Himes Member directly overlies the Little Sheep Mudstone. The lower part of the Himes is composed principally of resistant, trough cross bedded, lithic and feldspathic wackes and grayish olive to reddish brown massive silty claystones which locally interfinger with the upper part of the Little Sheep Mudstone. This lower unit appears to be confined to the northern reaches of the basin. The upper part of the Himes is a complex channel and fine grained mudstone sequence which includes a thick, lenticular, planar and trough crossbedded, quartz arenite informally referred to as the Greybull sandstone.

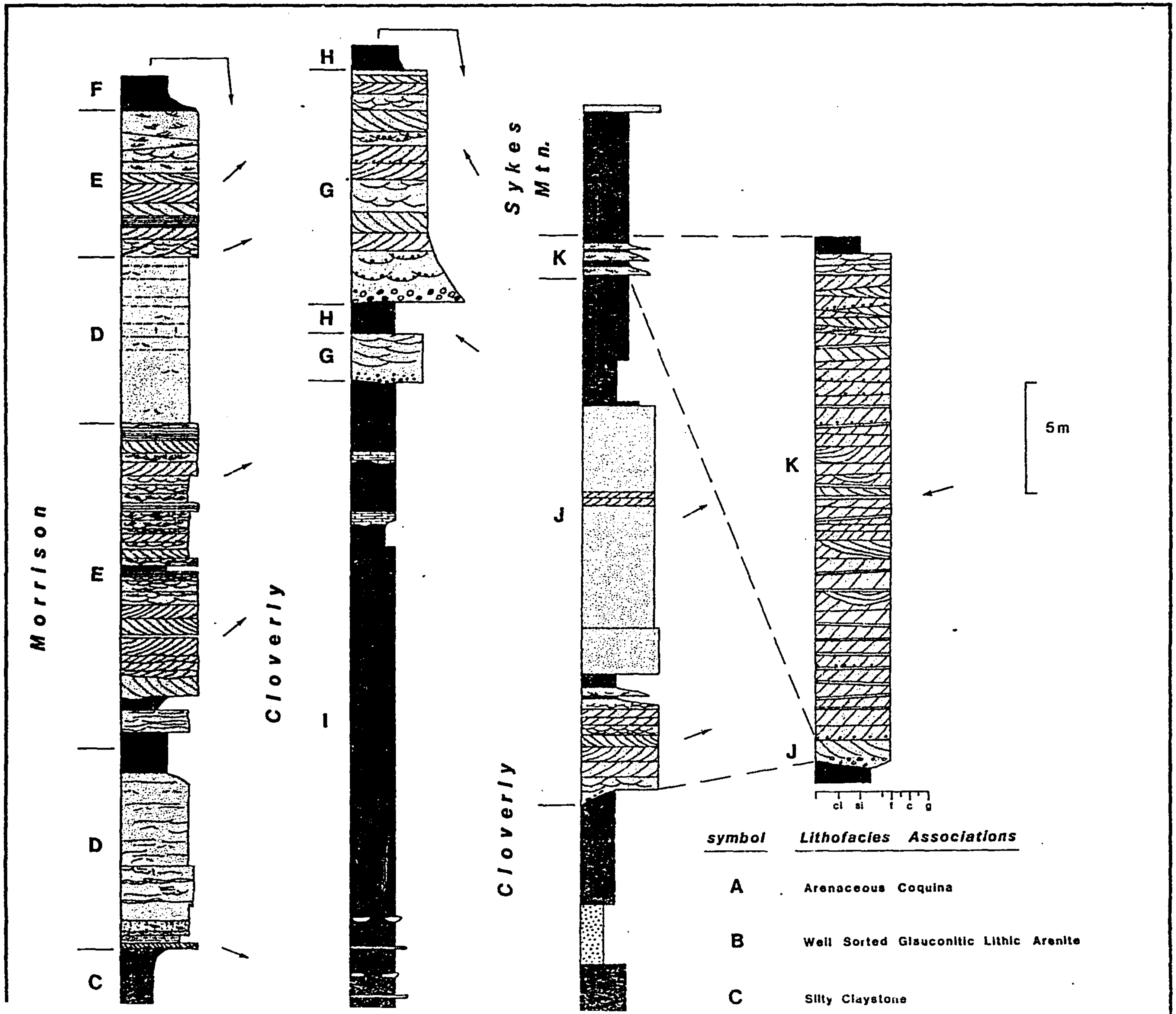
Upper Sundance and Lower Morrison

Description

A summary of the physical characteristics of the basal Morrison and upper Sundance allows for the identification of four distinct lithofacies associations (Figure 3). The lowest is defined as the arenaceous coquina lithofacies association. This is overlain by a well-sorted, glauconitic lithic arenite lithofacies association which is characterized by beautifully developed herringbone structures. Capping this is a thick silty claystone lithofacies association which contains local accumulations of discontinuous lenses of poorly sorted arenaceouus bioclastic siltstone. The silty claystone association is overlain by the laminated, very fine grained sandstone and siltstone lithofacies association (Figure 4a).

Figure 3.
Composite stratigraphic section of upper Sundance, Morrison, Cloverly and lower Sykes Mountain formations, Beaver Creek; Erratum Lithofacies Association E should read: Sandstone




Morrison
F
E
D
E
D
C
Cloverly
H
G
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Sykes Mtn.
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Cloverly
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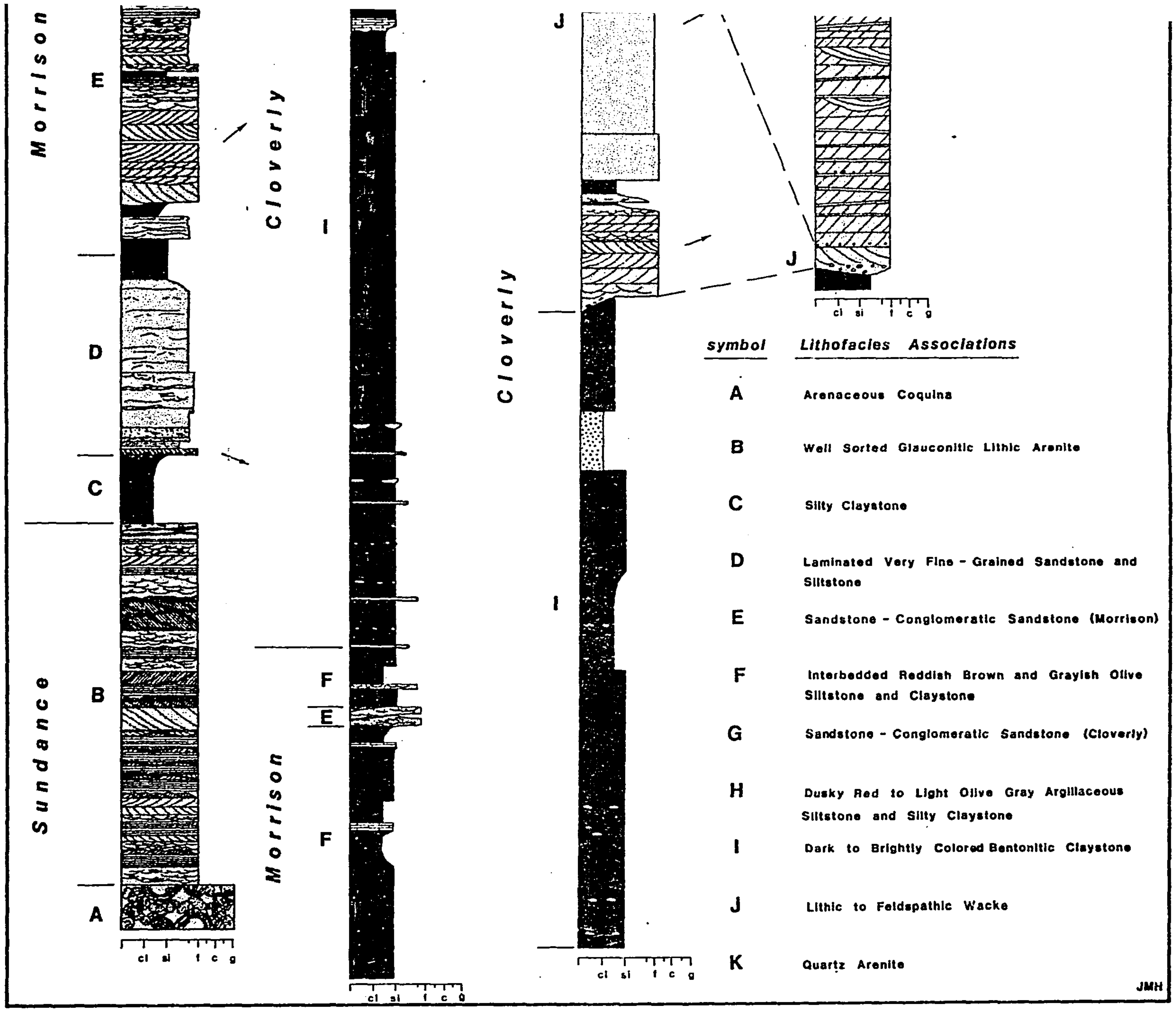
cl si f c g
5m
symbol
Lithofacies Associations
A
Arenaceous Coquina
B
Well Sorted Glauconitic Lithic Arenite
C
Silty Claystone



Morrison
E
Cloverly
I
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C
B
Sundance
A
cl si f c g
F
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Morrison
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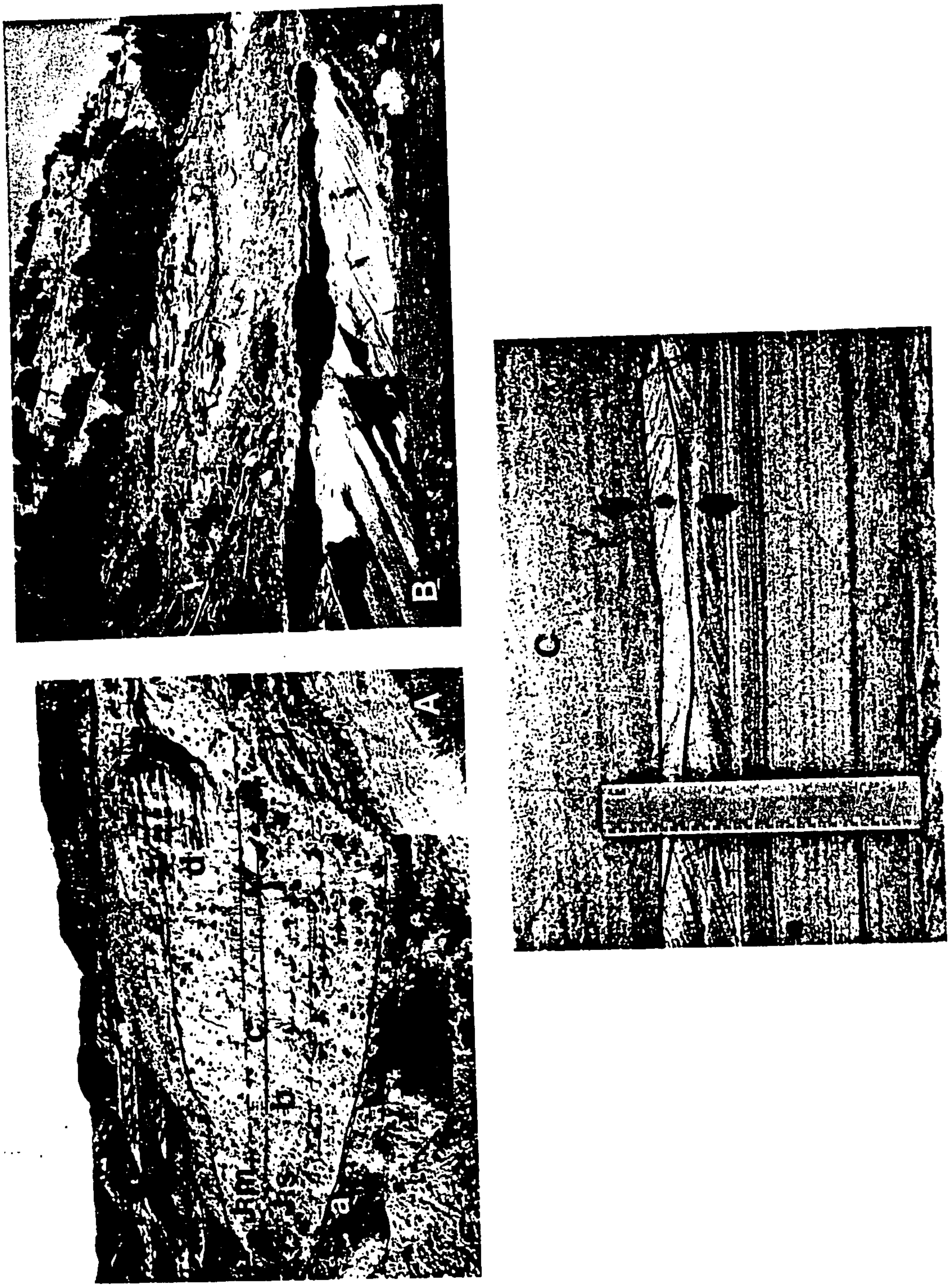
Cloverly
I
cl si f c g
J
cl si f c g
symbol
Lithofacies Associations
A Arenaceous Coquina
B Well Sorted Glauconitic Lithic Arenite
C Silty Claystone
D Laminated Very Fine - Grained Sandstone and Siltstone
E Sandstone - Conglomeratic Sandstone (Morrison)
F Interbedded Reddish Brown and Grayish Olive Siltstone and Claystone
G Sandstone - Conglomeratic Sandstone (Cloverly)
H Dusky Red to Light Olive Gray Argillaceous Siltstone and Silty Claystone
I Dark to Brightly Colored Bentonitic Claystone
J Lithic to Feldspathic Wacke
K Quartz Arenite
JMH

Figure 4a. Lithofacies association of upper Sundance and lower Morrison, Beaver Creek; small case letters refer to lithofacies associations of Figure 3

Figure 4b. Tidal bundles of upper Sundance, Horse Creek; arrows point to examples of reactivation surfaces; photo by J. R. Kvale, Greybull, Wyoming

Figure 4c. Tidal bundles of upper Sundance, Beaver Creek; large arrows signify relative flow direction of dominant flow structures, small arrow denotes relative flow direction of subordinant flow structure


B
A
C

The arenaceous coquina lithofacies association occurs in the upper Sundance and consists of a 2m thick, large scale trough cross bedded, black chert bearing, arenaceous, and glauconitic coquina of packed bivalves. No burrows were found within this unit. It is capped by the second lithofacies association consisting of 16m of light olive, glauconitic, well-sorted, fine to very fine grained calcareous lithic arenites. Primary structures include well-developed herringbone cross bedding consisting of normally dipping large scale foresets (30-100cm thick) which are truncated by reactivation surfaces lined with clay draped, reversely dipping ripple sets (Figure 4b). These are interpreted to be tidal bundles (Elliott, 1984; Uhlir and Vondra, 1984). Another type of tidal bundle present consists of parallel bedded sandstones exhibiting well-developed primary current lineations on bedding plane surfaces (Figure 4c). These are also truncated by ripple sets, but the capping clay drape is only occasionally present. Other structures not associated with reactivation surfaces include ripples, some of which are climbing, as well as parallel bedded, and large scale trough cross bedded sets. Burrows are poorly preserved and very rare.

A marked lithologic break occurs at the top of the glauconitic lithic arenite lithofacies association. The upper 1m of the lithic arenite sequence consists of flaser to wavey bedded, fine grained, well-sorted, glauconitic sandstone with thin stringers of silty claystone which are laterally continuous drape deposits (Figure 5a). This is capped by a 2.5m thick olive gray silty claystone of the third lithofacies association which is both massive and lenticular bedded with minor amounts of sand occurring as isolated starved ripples. The thick claystone is very similar in color and mineralogy to the claystone stringers in the lithic arenite lithofacies association. Thin bladed concretions of specular pyrite occur in the lithic arenite immediately below the thick silty claystone. The claystone contains organic remains including coaly material and rare, poorly preserved casts of gastropods. Microfaunal remains include dinoflagellates indicative of brackish water conditions (identified by E. Robertson, Phillips Pet. Co., pers. comm., 1985). The entire unit is burrowed and contains small (1cm) euhedral to subhedral selenite crystals. North of Sheep Mountain this unit contains discontinuous lenses of poorly sorted arenaceous bioclastic siltstone. The bioclasts consist of highly fractured bivalves.

The basal 13m of Morrison, which directly overlie the silty claystone lithofacies association, consist of interbedded yellowish gray to yellowish green, very fine-grained calcareous quartz arenites, quartz wackes, and

Figure 5a. Wavey bedded upper Sundance

Figure 5b. Dish structures in lower Morrison


A


B

siltstones of the fourth lithofacies association. The sandstones and coarse siltstones are mostly massive, but undulatory varve-like laminae of regularly and irregularly spaced finer grained siltstones and silty claystones are locally preserved. The laminae are often obscured by burrows but may be laterally continuous over many meters. Dish structures and small scale slump structures often cut the laminae (Figure 5b), indicating dewatering of rapidly deposited sediments (Lowe and Lopiccolo, 1974; Lowe, 1975). This suggests that the lack of observed primary structures may be related in part to the rate of sedimentation rather than the intensity of biologic activity. Bed thicknesses are generally less than or equal to 3 cm.

Thicker channel sandstone bodies are incised into the thin bedded sandstone and siltstone interval. These sandstones are up to 60cm thick, fine grained calcareous quartz arenites. They form flat bottomed lenticular channel deposits that pinch out over several 10's of meters. The sandstones consist of two or three erosive based sets of faint ripple bedded (some climbing) and large scale, unidirectional, trough cross stratified sands. A lag of rip-up clasts of silty claystone and occasional reptilian bone fragments commonly line the base of the sets. Volumetrically minor siltstones which are more oxidized (rust colored) in color than surrounding units are present within the basal Morrison lithofacies association. The siltstones are 10-15 cm thick and laterally continuous over several meters. The upper parts of the siltstone units are commonly brecciated with fragments 1 cm or larger surrounded by infilling of overlying siltstones. The horizons are only slightly calcareous, are more resistant than the surrounding materials and are highly burrowed with burrows uniform in size (1mm diameter) and nontapering. Most burrows are vertical with a few branching out downwards and may be the remnants of rootlets. The brecciation and infilling by the overlying material, the more oxidized color, and the presence of possible rootlets suggests subaerial exposure and possible pedogenic activity. Also of significance are 1cm thick units of grayish purple montmorillonitic claystones (see section on petrology). The claystones are locally silty, uniform in thickness, and are laterally continuous over many 10's of meters. These possibly represent altered volcanic ash. This lithofacies association is replaced to the south in the Thermopolis area by a distinctive eolian deposit of very large scale trough cross bedded lithic arenites noted by Mirsky (1962) and presently being studied in detail by David Uhlir of Iowa State University.

Depositional Environments

The transition from the uppermost Sundance lithofacies associations to those of the lower Morrison document a late Jurassic marine regression in the Bighorn Basin. Primary structures and lithologies of the coquina lithofacies and glauconitic lithic arenite lithofacies associations suggests that they represent subtidal to intertidal basal channel lag and channel bank deposits, respectively, produced by the lateral migration of tidal inlet channels along a prograding mesotidal barrier coastline (Uhlir *et al.*, 1986). The tidal inlet fill sequence is capped by the silty claystone facies association which is interpreted to be a lagoon or uppermost intertidal or supratidal mudflat deposit. This interpretation is based on: 1) the microfaunal content suggesting brackish water conditions, 2) the low energy depositional environment indicated by the dominance of mud sized material and general paucity of wave or current generated ripples, 3) the high organic content including the occurrence of storm deposited shell fragments, and 4) its relationship to the underlying tidal deposits. A lagoonal environment is perhaps more likely considering the barrier coastline interpretation of Uhlir and others (1986) for the underlying deposits. This fill sequence is similar to modern North Sea examples described by Boersma and Terwindt (1981), Terwindt (1971), and Visser (1980).

The primary structures of the interbedded very fine grained sandstone siltstone lithofacies association of the lower Morrison plus its stratigraphic proximity to underlying tidal deposits indicates deposition within an intertidal to supratidal coastal plain setting, possibly a marginal lacustrine environment or tidal flat with interfingering fluvial and/or tidal creek deposits. This interpretation is supported by the dominance of thinly interbedded to laminated and massive very fine grained sandstones and mudstones which show evidence of burrowing and possible pedogenic development. Sedimentation was periodically rapid as evidenced by the occurrence of dish structures and small scale slumps (Lowe and Lopiccolo, 1974; Lowe, 1975). Their presence may be indicative of sheet flooding, driven by fluvial or tidal processes, onto a tidal flat. The channel sandstone bodies which interfinger with and incise into the interbedded sandstones and mudstones represent fluvial channels or tidal creeks draining off the tidal flat or coastal plain into a lagoonal setting. The general lack of primary depositional structures, fossil remains and poor exposures of this interval makes a more precise depositional interpretation difficult. A study of the microfauna and flora coupled with a utilization of depositional structure enhancement techniques such as X-ray radiographs and polishing of large blocks collected from this interval may reveal additional data which would clarify the environmental setting of this lithofacies association. To the south, in the Thermopolis area, this depositional environment is replaced by a coastal dune setting developed directly on the tidal inlet channel fill sequence (D. Uhlir, Dept. of Earth Sciences, Iowa State Univ., pers. comm., 1985).

Upper Morrison

Description

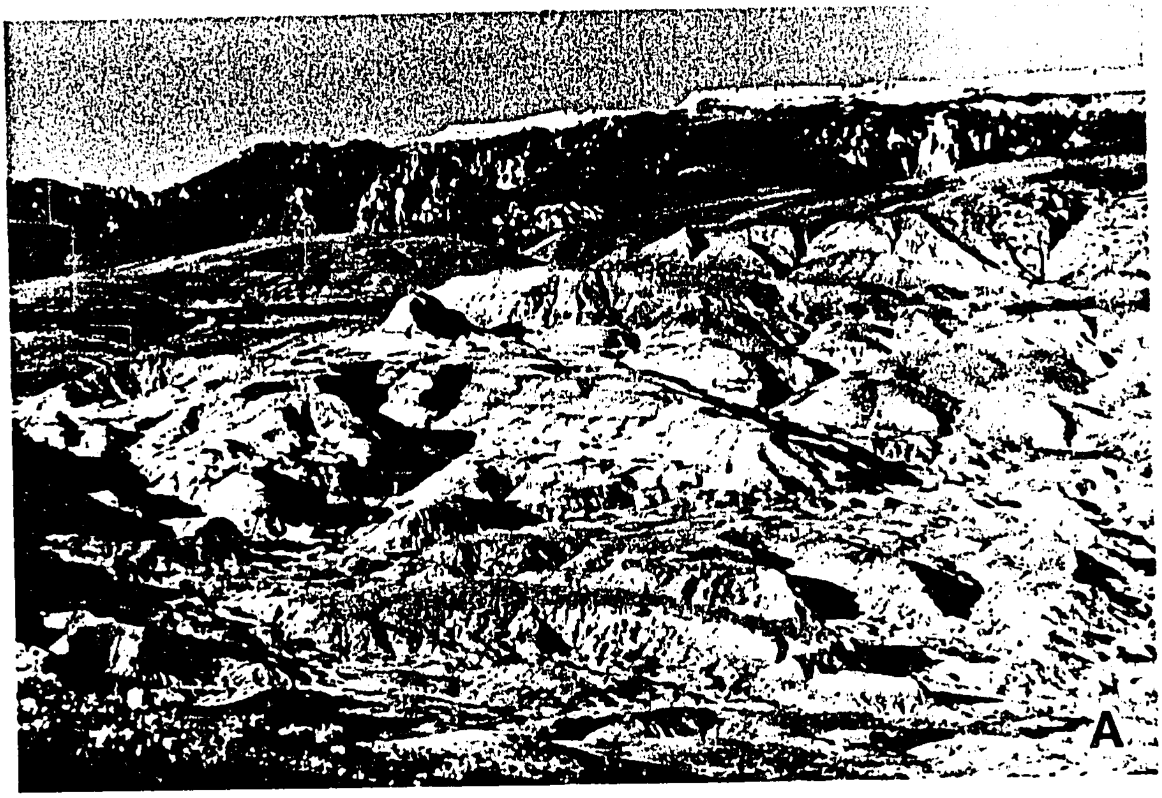
The upper part of the Morrison Formation can be divided into two lithofacies associations: 1) an interbedded reddish brown and grayish olive siltstone and claystone lithofacies association and 2) a fine grained channel sandstone lithofacies association (Figure 3). These lithofacies associations record deposition of fluvial channel sands and associated overbank deposits. The stratigraphically lowest major channel deposits observed in the Morrison in northeastern Bighorn Basin occurs at Beaver Creek (Figure 4a). There the channel sequences are typically broad, erosive based, 4 to 6 m thick, lenticular, flat bottomed, fine grained, multichannel deposits which fine upwards to siltstones. These are incised into, and interfinger with, the laminated very fine grained sandstone and siltstone lithofacies association. The sandstones are quartz arenites which occasionally contain grains of dark chert. At the thickest point of the channel, the lower sandstone sets are typically 40-100cm thick and are commonly lined with a basal lag of large goethite coated mudstone or wood fragments. Bone fragments are also present. Set thicknesses decrease upwards and laterally towards the channel margins. The interior of the channels are characterized by large scale trough cross bedded sets deposited in 3-dimensionnal large scale sand dunes. Some minor channel incision with oblique divergence of flow from the mean flow direction is present suggesting some longitudinal midchannel bar development within the interior channel units. Channel margin deposits are characterized by thin units (usually <30cm) which are generally ripple bedded or parallel laminated sets with primary current lineations (PCL) preserved on bedding plane surfaces indicating deposition as sinuous to discontinuous crested, current generated ripples and streaming lineations, respectively. Many of the thin units are extensively burrowed and interbedded with organic rich siltstones. Paleocurrent measurements indicate a northeasterly flow for the channel. Of the three channel complexes present in the lower fluvial facies of the Morrison at Beaver Creek, one shows evidence of limited development of lateral accretion surfaces indicating incremental growth of a point bar complex. The dominant bar forms, however, appear to be midchannel or side bars more typical of low sinuosity rivers.

The uppermost part of the Morrison is dominated by an interbedded reddish brown and grayish olive siltstones and silty claystones lithofacies association, similar in appearance to the red banded Eocene Willwood Formation in the basin (Figure 6). This association interfingers with and overlies the uppermost sandstones of the lower channel sandstones just discussed. The fine grained nature of the units is reflected by a badlands type topography that results from weathering and erosion. Of all the sediments of the Morrison and Cloverly formations in the Bighorn Basin, this lithofacies association contains the most abundant faunal remains. The famed Howe Dinosaur Quarry (Brown, 1935; 1937), located just east of the Beaver Creek/Cedar Creek confluence, occurs at this interval. It is from this quarry that large Jurassic sauropods consisting of mostly juvenile *Diplodocus* were collected (Dale Rusell, 1986, Canadian Museum of Natural History, pers. comm.).

Most of the siltstones and claystones are calcareous. The reddish color of the mudstones is probably the result of *in situ* weathering of nonred sediments which contained iron-bearing minerals such as hornblende or biotite (Walker, 1967). In very fresh exposures, many of the reddish units exhibit pale olive mottles which are localized around small

Figure 6. Morrison / Cloverly contact

Figure 7. Rootlets (arrows) in Morrison mudstone; stratigraphic up is to the left of the photo


A


B

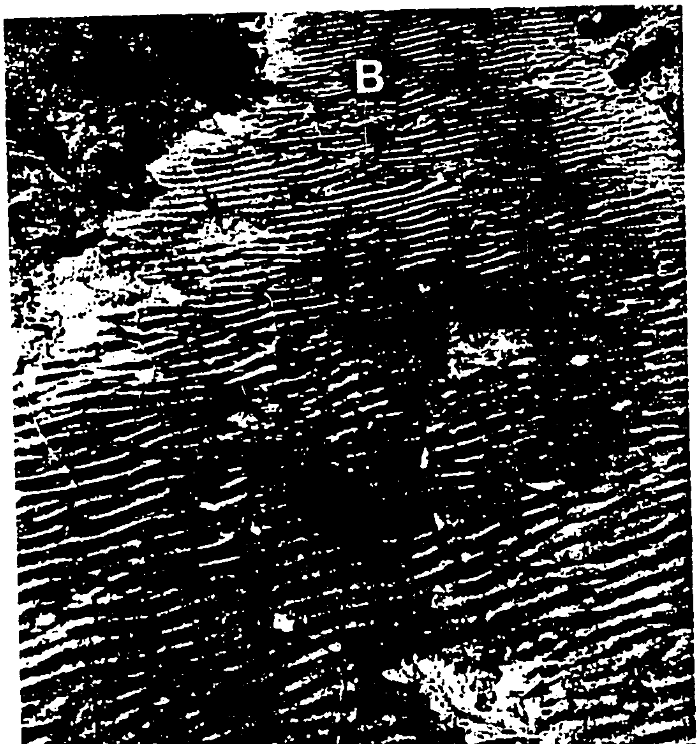
carbonate glaebules (<3cm diameter). In fresh samples, carbonate content is much higher in the pale olive mottles than in the red matrix. The remnants of downward branching rootlets were noted in the red horizons (Figure 7).

While the red colored horizons show little relief, the greenish mudstones may locally thicken and thin laterally as much as a meter or more, and in such cases appear to be fill deposits formed in small channels or depressions. Interbedded with the greenish mudstones are thin sandstones which are generally only a few centimeters thick, laterally discontinuous, and very fine grained. Sand-filled burrows several centimeters long occur within these horizons. In some areas, a greenish mudstone can be traced laterally to where it grades into a red mudstone.

Channel sandstones in the uppermost Morrison in the northern Bighorn Basin are thinner than those just described for the lower part of the upper Morrison. A good example of an uppermost channel sandstone is at Beaver Creek. This is a 1.5m thick, broad, flat bottomed sheet deposit that is encased in siltstones and is characterized by lateral accretionary surfaces which dip about 10 degrees (Figure 8a). The dominant primary structures include ripple laminations, sometimes climbing, parallel laminations with primary current lineations (PCL) on bedding plane surfaces, and also solitary large scale troughs. Burrows are present

Figure 8a. Lateral accretion surfaces in upper Morrison channel

Figure 8b. Dinosaur tracks (arrows) in upper Morrison


B

throughout the unit but are concentrated on upper and lower coset boundaries. An oscillatory rippled splay of sand off of this highly sinuous channel has faint impressions of dinosaur tracks preserved in it (Figure 8b). The geometry of the channel sandstones and associated overbank sands is that of a winged sandstone similar to the winged channels described by Friend and others (1979) in the Ebro Basin of Spain.

Depositional Environments

A summary of lithofacies for the uppermost Morrison Formation includes an interbedded reddish brown and grayish olive siltstone and claystone lithofacies that reflects floodplain deposition and a channel sandstone lithofacies association which originated in a fluvial channel environment. The coloration, presence of carbonate glaebules (caliche) and downward branching rootlets suggests that the reddish siltstones and claystones are the remains of better drained soils formed on a more distal part of the floodplain, whereas the grayish olive siltstones represent a more proximal floodplain setting where oxidizing conditions were not as intense perhaps because of a higher ground water table (Bown, 1985). Red beds have been traditionally associated with hot dry climates (Turner, 1980) although Walker (1974) discussed red beds forming in moist climates. Caliche horizons have been used as paleoclimatic indicators of semiarid to arid environments where evaporation rates exceed rates of precipitation (Allen, 1974; Blatt et al., 1972; Leeder, 1975). Semeniuk and Searle (1985) have shown however that a dry evaporitic climate in a coastal plain environment is not always conducive to the development of caliches, but that vegetation types and water table depths may be more important. They report maximum caliche development in humid areas where precipitation exceeds evaporation. Weak caliche development (i.e., the formation of small isolated glaebules and lack of crudely laminated nonglaebular porcelanous carbonate rock) has been suggested by some authors to signify relatively rapid rates of deposition on geomorphic surfaces that do not undergo prolonged pedogenesis (Allen, 1974; Leeder, 1975). However, the Morrison glaebule development may be reflecting conditions other than relatively high sedimentation rates such as aridity or vegetation. Semeniuk and Searle (1985) have noted the need for a thorough documentation of the variability of calcretes in Holocene sediments and question the validity of a direct correlation of caliche in the rock record to arid or semiarid environments.

The sinuous channel complexes which are associated with the overbank deposits are dominated by climbing ripples laminations and horizontal bedding indicating high depositional rates and high flow in flashy streams (Miall, 1978; Picard and High, 1973; Tunbridge, 1981). They are also characterized by epsilon cross stratification indicating that lateral accretion of point bars was important. Streams characterized by flashy discharge are typically ephemeral and are best known from arid to semiarid environments (Miall, 1978; Picard and High, 1973; Reineck and Singh, 1980), however, the presence of point bars suggests a sinuous, relatively mature stream with steady discharge that was not short lived (Collinson, 1978; Reineck and Singh, 1980; Smith, 1985). Sinuous streams are typically from areas characterized by low gradient conditions coupled with cohesive and vegetated overbanks (Collinson, 1978; Reineck and Singh, 1980). The abundance of large sauripod remains at this level suggests that vegetation was at least seasonally abundant, but the lack of coal or coaly beds indicates that extensive bogs or swamps did not exist. Flooding out of the channels presumably deposited overbank muds, but the overwhelming volumetric dominance of overbank mudstones to channel sandstones suggests that additional processes such as wind or sheetwash were responsible for mud deposition as well (Wells, 1983). In summary, the Morrison Formation in the northern Bighorn Basin was deposited in a coastal plain setting and represents a transition from marine tidal deposits (upper Sundance Fm.) to fluvial deposits (upper Morrison). The nature of upper Morrison channel sandstones and mudstones suggests that the Late Jurassic climate was seasonal and somewhat arid, but the evidence for this is not conclusive.

Little Sheep Mudstone and Lower Himes

Description

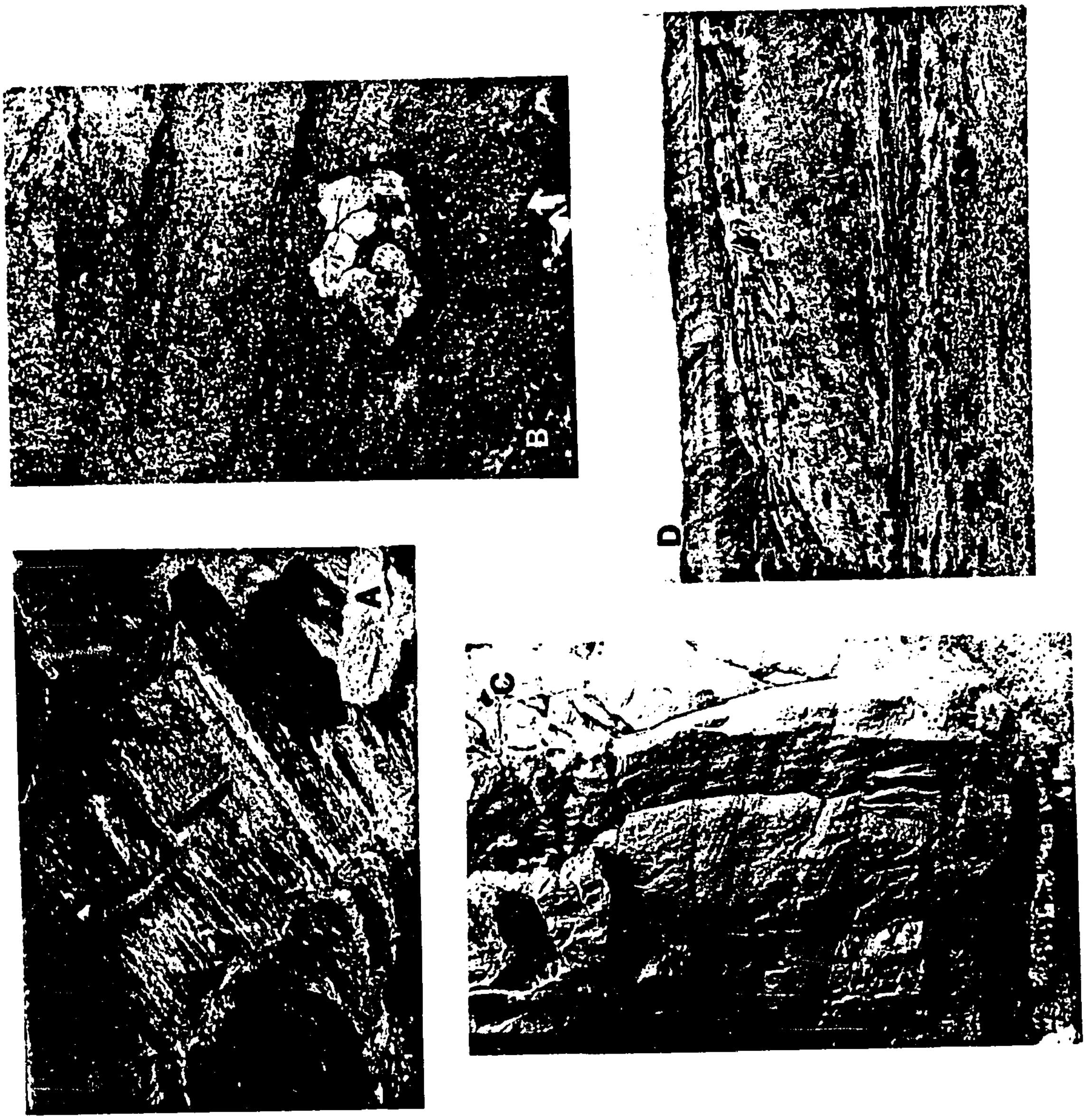
In the northeastern and eastern margins of the Bighorn Basin, the Little Sheep Mudstone Member of the Cloverly Formation can be divided into three lithofacies associations (Figure 3). These are: 1) the sandstone-conglomeratic sandstone lithofacies association, 2) the dusky red to light olive gray argillaceous siltstone and silty claystone lithofacies association and 3) the dark to brightly colored bentonitic claystone lithofacies association. Volumetrically, the sandstone-conglomeratic sandstone lithofacies association is the least important along the east flank of the basin but locally dominates the sequence south of Pryor Mountain. Most of the nonconglomeratic sandstones bodies in the Little Sheep Mudstone Member are thin, very calcareous, fine to very fine grained quartz arenites and quartz wackes. They form sheet deposits which exhibit a variety of primary structures ranging from climbing ripple laminations to parallel bedding with solitary trough sets (Figure 9a). These sandstones have little overbank material associated with them and are highly

Figure 9a. Ephemeral stream deposits in Little Sheep Mudstone

Figure 9b. Large irregularly shaped sandy chert clast in basal Cloverly channel conglomerate at Beaver Creek

Figure 9c. Medial Little Sheep Mudstone conglomeratic sandstone channel at Beaver Creek; arrow points to calcrete in overbank mudstone

Figure 9d. Medial Little Sheep Mudstone conglomeratic sandstone channel sequence north of Greybull


B
D
A
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burrowed on upper coset boundaries. Their morphology is that of broad, lenticular, multilateral, flat bottomed channels that generally show no evidence of lateral accretion, although a channel with well developed lateral accretion surfaces does occurs along the west flank of Rose Dome north of Sheep Mountain. They are usually encased in darkly colored claystones belonging to the third lithofacies association. Based on primary structures, channel morphology, and general absence of overbank material, these sandstones are interpreted to be deposited by sand dominated, flashy ephemeral-type, low sinuosity streams (see depositional environments section). Excellent examples of these sandstones are exposed along the west flank of Sheep Mountain north of the town of Greybull. Most of the thin sandstones indicate a westerly or southerly source area.

A series of channels which occur in the lower half of the Little Sheep Mudstone is a significant exception to the dominant ephemeral-type of stream system. Such channels are present though out the Bighorn Basin where they form sheet or lenticular deposits of thick, stacked, multistorey sequences of black chert bearing conglomeratic quartz arenites and/or fine to medium grained quartz arenites. The dominant primary structures include both large and small scale trough cross stratification indicating deposition within highly sinuous to discontinuous crested bedforms

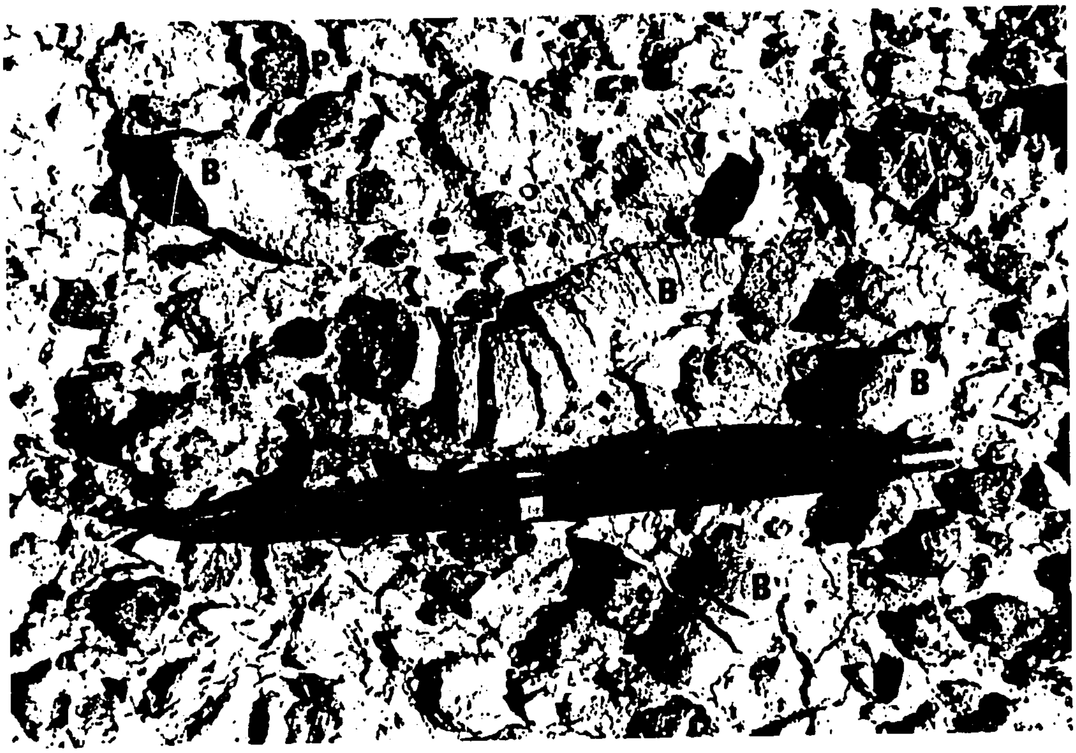
(Figures 9b, c, d). Gravel-sized clasts (<10cm) are dominated by well-rounded dark chert with lesser amounts of quartzite, bone fragments, and opalized wood fragments. Large irregular shaped boulder sized (up to 20cm) clasts of light colored sandy chert are also present (Figure 9b). This chert is very similar to the light colored Cloverly cherts which are believed to be devitrified tuffs and as such are interpreted to be intraformational in origin. Oblique minor channel incision of bedforms coupled with a lack of lateral accretion surfaces indicates deposition of these units within low sinuosity braided channels.

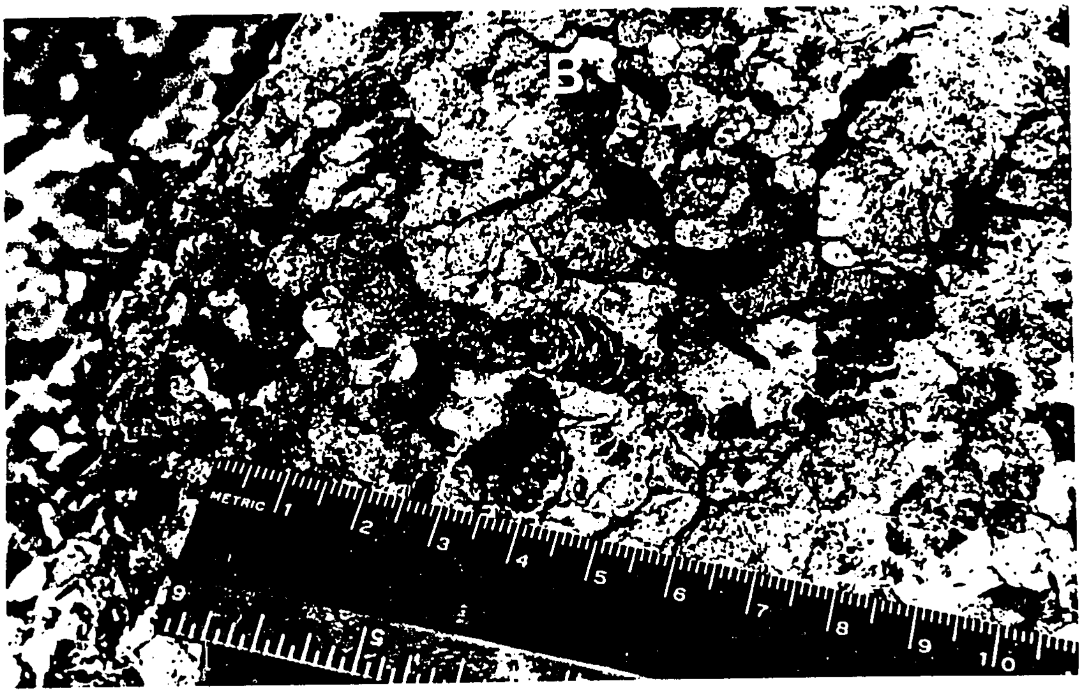
These conglomeratic sandstone channels are significant because, unlike the other Little Sheep Mudstone channels, they have relatively thick, well developed overbank deposits associated with them indicating that these are perennial stream deposits. The overbank deposits are grouped into the second lithofacies association of importance in the Little Sheep Mudstone Member; the dusky red to light olive gray argillaceous siltstone and silty claystone lithofacies. This facies is laterally persistent and very similar to the overbank lithofacies association of the uppermost Morrison Formation at Beaver Creek and Sheep Mountain. It is characterized by mottled reddish and olive colored silty claystones which are either calcareous or contain distinct horizons of dense massive carbonate nodules and layers which are generally unsilicified. These are interpreted to be caliche horizons (see depositional environments section). The more olive colored units contain a slightly higher silt fraction. Large euhedral selenite crystals with dissolution features are often present as surface lag on these units. Dominant clays within this unit include mixed layered illite-montmorillonite and kaolinite.

The siltstones are occasionally interbedded with thin, poorly sorted, calcareous sandy siltstones or very fine grained silty sandstones which contain abundant pedotubules. The tubules are generally nontapered and straight to slightly twisted with no branching. Their contact with the enclosing matrix is sharp. The beds are indurated but upon weathering disaggregate and form a lag of individual tubules and small angular fragments (roughly 1cm across) of similar sizes (Figure 10a). The pedotubules resemble those described by Bown and Kraus (1981) from the Eocene Willwood Formation which they believe are the remains of filled burrows of insects and/or other invertebrates. The angular fragments may be the remains of pedons because of their similar size, poorly sorted lithology, and association with the burrows. Occasionally in the olive gray siltstones, subhedral to euhedral cubes of goethite are found filling voids in the carbonate nodules and carbonate cemented

Figure 10a. Pedotubules (B) and pedons (?) (P) from Little Sheep Mudstone

Figure 10b. Meniscated burrow from Little Sheep Mudstone devitrified tuff


P
B
B
B
B
P


B
METRIC
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10

burrows. The cubes are often concentrated on ant mounds built on the siltstones.

The third lithofacies association is the most characteristic of the Little Sheep Mudstone Member. This is the dark to brightly colored bentonitic claystone lithofacies. This lithofacies is unlike anything in the Morrison Formation. Typically, the claystones are greenish black to grayish red, weathering to various hues of black and purple. They are siliceous and noncalcareous. They may be silty and usually have well-developed surfaces of slickenslides preserved in fresh samples. Diagenetic processes have so altered the claystones that few primary structures reflecting the original depositional environment remain. However, rare burrows, occasionally with meniscated back-filled structure, are present. Commonly the claystones contain beds of silica nodules with minor amounts of void filling calcite and barite and carbonate nodules, the larger ones exhibiting calcite, silica and barite filled septarian fractures. These concretions were utilized by the early American Indian as a source of stone to be fashioned into tools and weapons. The nodular horizons are localized but occur throughout the unit.

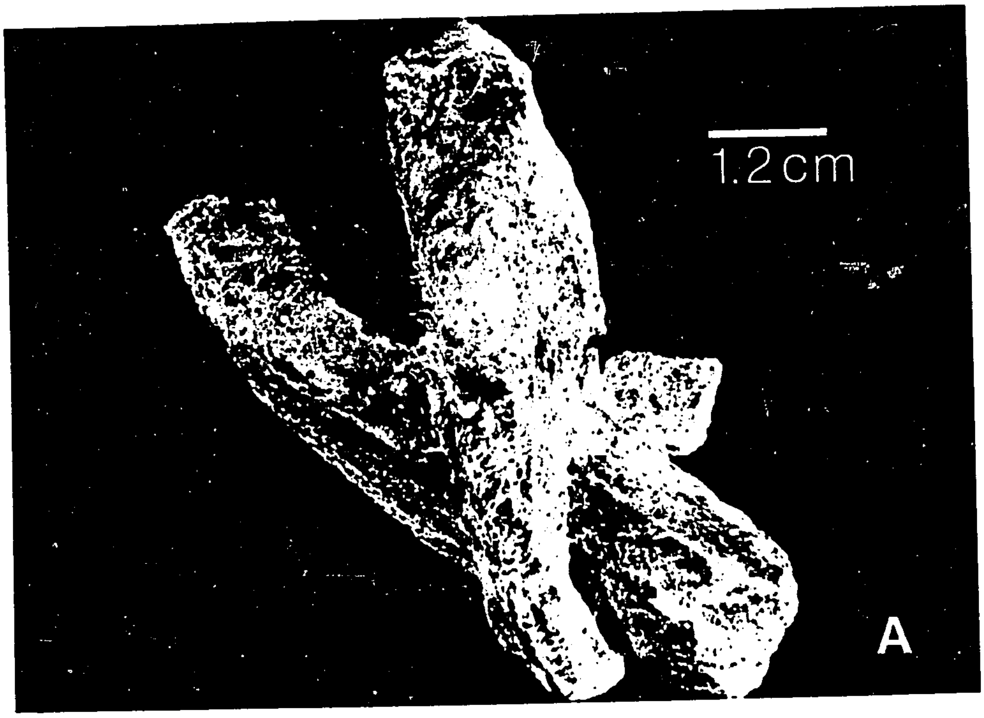
Of paleontological interest is the discovery in the Little Sheep Mudstone of a theropod tooth, possibly a <u>Deinonychus</u> in the midst of the remains of a <u>Tenontosaurus</u>.

Deinonychus is thought to have been the "wolf" of the Early Cretaceous, commonly hunting in packs, and his remains have been reported in association with Tenontosaurus skeletons elsewhere (Colbert, 1980).

One of the most significant features of this lithofacies association is the presence of white, very fine grained, mostly massive chert, up to a meter thick, which is locally arenaceous. These beds are the remains of devitrified tuffs and can be found throughout this interval in the northern reaches of the basin. The most conspicuous of these cherts occurs near the top of the Little Sheep Mudstone Member and is present throughout the northeast and eastern margins of the Bighorn Basin. The devitrified tuffs are lenticular, locally burrowed (Figure 10b) and rippled, and sometimes occur in sequences interbedded with siliceous bentonitic claystones. Lateral to the devitrified tuffs may be nodular horizons of silica or carbonate. The lenticular nature of the deposits and the presence of ripple bedding and sand size material suggests that the ash was deposited (or redeposited) in shallow depressions such as a pond and reworked by fluvial and/or biologic processes. Chalcedony and calcite pseudomorphs after displacively grown lenticular selenite occur locally (Figure 11a, b). At Douglas Draw, just west of Shell, silica pseudomorphs of selenite occur lateral to a pond-like filling of devitrified tuff.

Figure 11a. Chalcedony pseudomorph after selenite; identified by Dr. D. Biggs, Iowa State University

Figure 11b. Calcite pseudomorphs after selenite, identified by Dr. D. Biggs, Iowa State University


1.2 cm
A

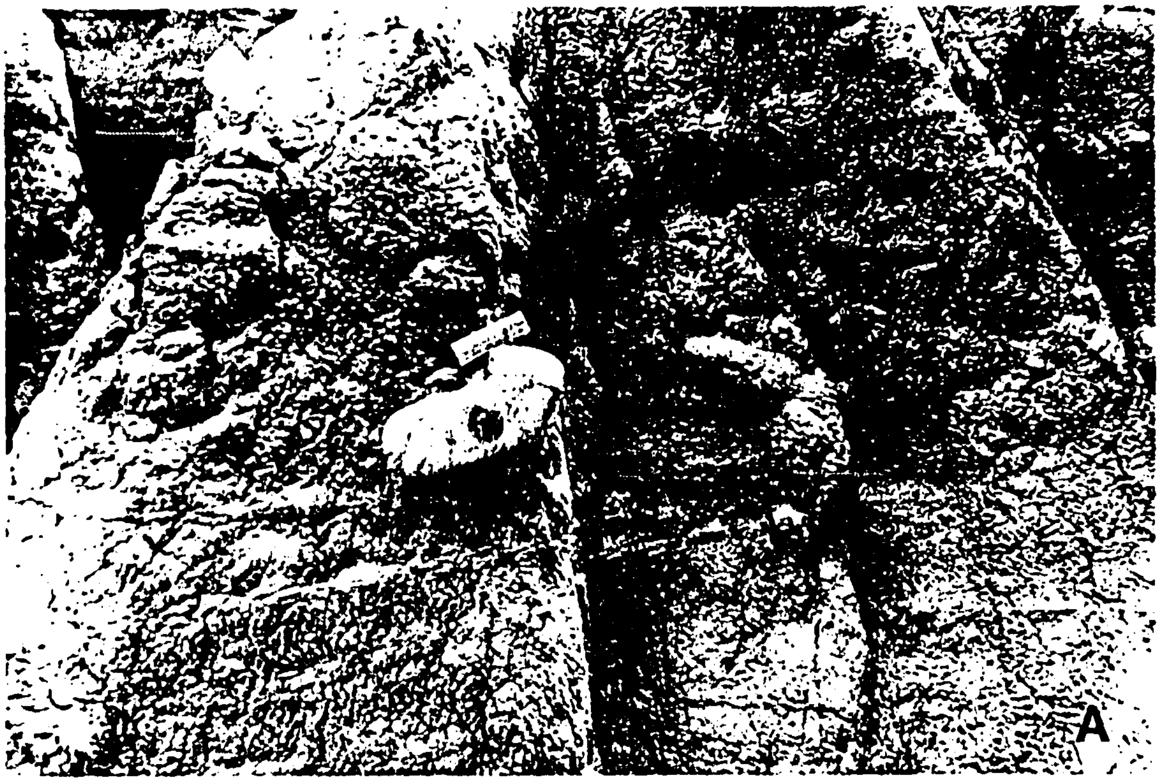

1.2 mm

Possible Megalosaurus teeth, crocodilian and possible fish bone fragments were found associated with a devitrified tuffaceous sequence giving additional support to a pond-like depositional setting. This relationship will be discussed more fully in the depositional environments section. At least three levels of devitrified tuffs are present in the Little Sheep Mudstone Member in the eastern part of the basin.

The Himes Member is the uppermost member of the Cloverly Formation and overlies the Little Sheep Mudstone Member. It can be subdivided into two lithofacies associations: a lower lithic to feldspathic wacke lithofacies association and an upper quartz arenite lithofacies association (Figure 3). In the northern part of the basin, the basal contact of the Himes is typically a disconformity overlain by a stacked sequence of Himes channel sandstones, composed of lithic to feldspathic wackes that contain abundant clasts of zoned plagioclase, polycrystalline quartz, and volcanic rock fragments (see the petrology section). However, locally it does interfinger with the underlying Little Sheep Mudstone Member at the south flank of Rose Dome north of Sheep Mountain. A boulder sized clast (long axis of 25cm) of andesite was found insitu in the Beaver Creek area (Figure 12a) with comparable sized clasts of basalt found as float at the base of weathered

Figure 12a. Andesite clast in lower Himes channel; ruler is 15cm in length

Figure 12b. Geomorphic expression of lower Himes channels (green); lenticular tan-colored sandstone is type III upper Himes channel


A


B

lower Himes channels along the west flank of Little Sheep Mountain southeast of Lovell. Diagenetic clay (montmorillonite) is an important component in the lower lithofacies and upon weathering forms a popcorn textured surface on outcrops which obscures primary structures. The sandstones of this lithofacies association are colored various hues of green and gray and exhibit large scale trough cross bedding that indicates a general paleoflow to the east. The sandstone sequence fines upwards to grayish green to moderate brown clayey siltstones which locally contain organic remains, are bentonitic and often mottled. The lower lithofacies is resistant and forms prominent cliffs above the low rolling hills of the Little Sheep Mudstone Member (Figure 12b).

Depositional Environments

The lithofacies associations of the Little Sheep Mudstone Member and lithic wacke of the Himes Member in the Bighorn Basin are part of an extensive clay playa complex which extended as far east as the Powder River Basin (Fuson Shale) and as far west as the modern Absaroka range, perhaps farther. The dark to brightly colored bentonitic claystone lithofacies association, which includes the low sinuosity sandstone bodies dominated by parallel bedding and climbing ripple laminations, represents deposition in an ephemeral saline lake (mudflat) fed by flashy ephemeral streams.

These deposits are recognized by: 1) the occurrence of silica and calcite pseudomorphs of lenticular gypsum which grew displacively in the bentonitic mudstones, 2) occurrence of flashy ephemeral streams interbedded with the mudstones and 3) presence of abundant nodules and veinlets of silica, calcite and barite within the mudstones. Interfingering with the mudflat deposits are the sediments of perennial streams represented by the dark chert-bearing, multistorey, conglomeratic channel sequences as well as the lower Himes channel deposits. These features are quite common in the modern and ancient examples of clay playa complexes described by Handford (1982) Hardie and others (1978), and Tunbridge (1984). Similar deposits on the Indogangetic plain have been referred to as "terminal fans" by Friend (1978).

The Bighorn Basin of Little Sheep Mudstone times offered a suitable setting for the development of ephemeral saline lakes. It is quite possible that the fold-thrust belt to the west resulted in northern Wyoming developing into a rain shadow tectonic basin (see tectonics section). The lack of channel incision in the ephemeral channels and the lateral persistence of the mudstone units indicates that relief was subdued and it is possible that segments of the northern Wyoming foreland basin experienced hydrologic closure or restriction.

Inflow into the Little Sheep Mudstone playa was by ephemeral and perennial streams although unconfined sheet flow and ground water may also have supplied water to the basin. During wet seasons, water depth within the ephemeral lakes was never very deep perhaps only exceeding a few centimeters. This is suggested by the lack of remains of identifiable lacustrine organisms and shoreline features such as deltas, beaches, etc. Very shallow lakes lack strong currents capable of reworking sediments into shoreline deposits (Hardie et al., 1978). The occurrence of crocodilian and other reptilian remains in the lenticular devitrified tuffaceous sequences indicates that ponds periodically persisted through time accumulating volcanic ash and maintained salinites tolerated by these organisms, suggesting that the ponds may have been fed by fresh water springs.

The chemical constituents of brines which deposit mineral assemblages of clay playas were determined by Hardie and Eugster (1970) to be ultimately controlled by the bedrock lithology which supply sediment to the basins. Authigenic minerals identified as having existed in the Little Sheep Mudstone playa include gypsum, calcite, barite, silica and clay. Volcanic ash undoubtedly supplied Ba^{+2} and Ca^{+2} plus SiO_2 and montmorillonite. Paleozoic strata exposed in the fold-thrust belt and possibly in intrabasin uplifts could have supplied SO_4^{-2}, Ca^{+2}, and CO_3^{-2}.

A saline mudflat (mudflat saturated by brine) as opposed to a dry mudflat (terminology from Hardie et al., 1978) is proposed for the Little Sheep Mudstone in the Bighorn Basin because of the lack of primary structures and the occurrence of silica and calcite pseudomorphs after euhedral gypsum in the dark mudstones. Cody (1979) has shown that lenticular gypsum indicates the occurence of dissolved organic matter in the presence of alkaline conditions. Saline mudflats are characterized by muds full of displacively grown salt crystals (Hardie et al., 1978). The result of the growth of the salts is the hyaloturbation of the muds and the subsequent destruction of the preexisting primary structures (Hardie et al., 1978; Handford, 1982). While gypsum crystals are preserved in the overbank deposits of the perennial streams (age uncertain), evidence of gypsum crystal growth in the mudflats is only preserved by their pseudomorphic replacement by silica and calcite. The occurrence of other salts can only be speculated. The soluble minerals appear to have been selectively removed, perhaps by late diagenetic processes, leaving behind only the less soluble minerals such as silica, calcite and barite.

The silica, barite and calcite nodules exhibit either a mottled texture, indicating a coalescence of growing patches of carbonate or silica, or a massive texture, suggesting they were deposited from saturated groundwater (Semeniuk and Searle, 1985). The large carbonate nodules are cut by septarian fractures filled with silica, barite and calcite which indicate dessication of the carbonate nodules, perhaps during dry seasons, and subsequent infilling of secondary minerals during later phases of mineralization (see Petrology section).

The primary structures and lenticular nature of the nonnodular beds of sandy cherts (devitrified tuffs), many of which form lateral to nodular horizons but are encased by the same dark siliceous siltstones, suggests that these cherts were formed by processes somewhat different from those that deposited the silica nodules. James (1977) offered two possible models to explain the Kootenai "upper calcareous member" cherts of southwest Montana. Either model helps to explain the presence of not only the cherty and carbonate nodules in the Little Sheep Mudstone but also the chalcedony, barite and calcite veinlets found through out the member as well as the devitrified cherty tuffs.

The unifying concept of both models is the increasing solubility of silica and decreasing solubility of calcite with increasing pH (Krauskopf, 1979; Davis, 1983). The pH values of lakes with normal calcium concentrations generally vary from 7 to 9 (Krauskopf, 1979). Under calcium-rich lacustrine conditions, pH can increase to 8.5 and above with changes in physical, or more importantly, biogenic conditions (Hardie et al., 1978). For example high photosynthetic rates will dramatically increase CO_2 assimilation and reduce the amount of free CO_2. The result is an increase in pH which will continue during the summer when considerable amounts of silica, could be taken up into the system and calcite precipitated. High evaporation rates such as in an ephemeral lake or pond would concentrate not only the silica and calcite but also barite and gypsum enhancing the deposition of all four minerals (Hardie et al., 1978; Krauskopf, 1979).

The first model utilizes a thermally stratified lake divided into an epilimnion (warm, circulating, and turbulent), metalimnion (steep temperature gradient), and the hypolimnion (cooler, deep, undisturbed portion of the lake). The lake contains broad shallow shelves which are totally or nearly contained within the epilimnion. With an increase in photosynthesis, CO_2 values decrease and pH increases. The result is an increase in the dissolution of detrital silica (probably volcanic ash). The dissolved silica would be restricted to the epilimnion. Dead algae and other organic material would settle through the epilimnion and metalimnion and collect in the hypolimnion. The decomposition of the organics would increase the CO_2 content of the hypolimnion and lower the pH. A partial or complete overturn or mixing of the hypolimnion and the epilimnion would result in a decrease of pH in the epilimnion and the precipitation of inorganic gelatinous silica on the lake bottom. If mobile enough, these masses of silica could move along permeable boundaries such as bedding planes or fractures in underlying sediments forming veinlets and nodules.

The second model involves a weakly to nonstratified lake characterized by shallow depths in windy or hot regions. The lake, as a result, is continually mixed with uniformly distributed dissolved silica. In this case, silica enriched waters would continually move down through the sediments precipitating silica around decaying organic matter. The lack of evidence for the development of deep lakes in the Little Sheep Mudstone makes this second model more likely. In an ephemeral lake, the silica would be concentrated even more through evaporation, perhaps eventually killing the organics, suddenly lowering pH and causing the precipitation of silica. Another possibility is that flooding into an ephemeral lake would cause the necessary decrease in pH and subsequent inorganic precipitation of silica. The presence of sand and silt sized particles within the Little Sheep Mudstone devitrified tuffs and the presence of very thin sandstones and siliceous siltstones interbedded with the cherty nodular horizons suggests that periodic flooding may have caused the precipitation of silica.

The nodules and veinlets were formed by the mobilization of silica, calcite and barite rich waters or gelatinous masses into fractures or bedding plane surfaces along banks or bottoms of ponds or lakes. These minerals sometimes replaced other minerals which had already been deposited such as gypsum or preexisting carbonates. The devitrified tuffs, however, show evidence of reworking both by fluvial processes and by organisms indicating it was altered insitu. Its lenticular shape, primary structures and bedding characteristics suggests that it was carried in by winds or water, deposited in shallow ponds and reworked by fluvial processes rather than being deposited as a gelatinous mass eventually forming a silica nodule. The biologic and fluvial processes all occurred before the tuff was totally devitrified and the silica crystallized to form a cherty material. The chalcedony veinlets which both underlie and cut the tuffaceous sequence indicate that at least some silica became mobile enough to fill fractures and move along permeable boundaries.

Two possible reasons why chalcedony veinlets and silica cement are not as important in the Morrison Formation as in the Cloverly Formation are: 1) there was not an abundant source of silica (ash) available during the Late Jurassic and 2) less arid conditions prevented the formation of ephemeral ponds during Morrison times.

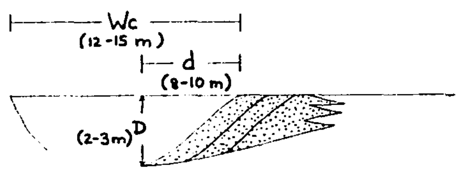
Upper Himes and Sykes Mountain

Description

The uppermost unit of the Himes Member of the Cloverly Formation consists of an interval of three distinct types of channel complexes deposited under mostly fluvial conditions, which are characterized by westward directed paleocurrent indicators (Figure 13). In the southern part of the basin, this interval rests disconformably on the Little Sheep Mudstone. The youngest of the channel types, (type I) and most significant economically, is a thick (usually more than 15 m), elongate channel complex which was deposited within a generally westward flowing, low sinuosity, fluvial dominated system (see Appendix A for legal locations and paleoflow trends of major type I deposits in Bighorn Basin). This unit is informally referred to as the Greybull sandstone. The terminology originated as a drillers term (Greybull sand) in the Greybull Dome, south of the town of Greybull, for an upper Cloverly sandstone which yielded hydrocarbons

Figure 13. Summary of major channel types in Greybull interval

Channel Charac-teristics	High Sinuosity, Sand-Filled Greybull Channel	High Sinuosity, Mud and Sand-Filled Greybull Channel	Low Sinuosity, Sand-Filled Greybull Channel
Channel Width ($W_c = 1.5d$)	12 - 15 m	38 m	350 - 1100 m
Channel Depth (D)	2 - 3 m	4 m	0.6 - 1.5 m
Sinuosity ($P = 3.5F^{-0.27}$)	2 - 2.3	3.5 m	0.8 - 0.5
Channel Behavior	Some lateral migration between switching	Some lateral migration between switching	Fixed channel
Descrip-tion	- Ripple laminated v.f. quartz arenites mostly confined to channel point bars. - Overbank massive mudstones	- Multistorey channel sequence: massive to ripple laminated mudstones and v.f. quartz arenites which pass into thicker large-scale trough cross-bedded sandstones near center of channel.	- Channel-fill sequence of fine-grained quartz arenites with thin overbank deposits - Large-scale tabular planar foresets
Interpre-tation	Meandering channel deposit reflecting lower gradient and/or lower discharge.	Meander channel deposit reflecting low gradient and higher discharge.	Sandy low sinuosity channel deposit reflecting high gradient and highest discharge. Youngest channel deposit.



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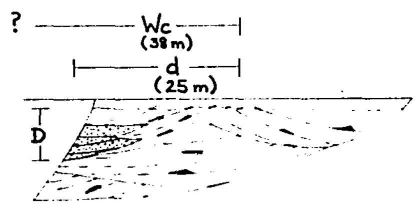
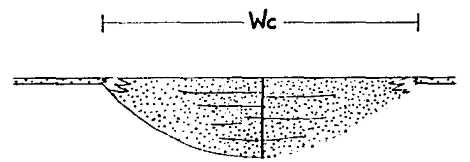


L.A.S. = 18°



15 m (+)

and water (Hintze, 1915). Hewett and Lupton (1917) attempted to designate the Greybull sandstone as a formal member of the Cloverly but failed to name a formal type section. The unit has also been referred to as the Dakota sandstone (Hintze, 1915), and the "B" unit of the Greybull sandstone (Stone, 1983). All of these terms are still informally applied to the uppermost sandstones of the Cloverly Formation and lowermost sandstones of the Sykes Mountain Formation by geologists and drillers in the basin (Steve Hollis, Marathon Oil Company, pers. comm., 1984).

Associated with this channel type are thin (usually less than 1 m) laterally continuous sandstones which are interbedded with nonresistant argillaceous siltstones and represents, in part, overbank deposits. The second and older channel body (type II) is an upward fining, interbedded mud and sand dominated, multilateral, multistorey channel complex with well-developed lateral accretion surfaces indicative of point bar migration in a high sinuosity meandering channel complex (see Appendix B for legal locations of some of these deposits). Individual channel thicknesses are on the order of 4 to 5m. The third type of channel body (type III) is a 1 to 3 m thick, upward fining, ripple laminated sandstone which possesses lateral accretion surfaces indicative of point bar migration in a high sinuosity fluvial channel complex (see Appendix B for legal locations of some of these deposits). This may be the oldest of the three channel body types. Thin, laterally restricted mudstones characterize the overbank deposits associated with the type II and III channels.

Type I channel bodies are well-exposed in cliff faces along the flanks of Laramide uplifts in the northern reaches of the Bighorn Basin. Individual channels range in width from 350 to 1100m. They form elongate sand bodies that are deeply incised into underlying strata up to 16m. Maximum thickness of these channel bodies is 25m in the Beaver Creek area (north of Shell) and the Sykes Mountain area, south of Pryor Mountain, but thins to 15 m farther out into the basin. The sandstone bodies appear to be fixed channel systems that filled continuously after initial incision into the bedrock. This interpretation is supported by the lack of multistorey fill sequences typical of the other two channel types.

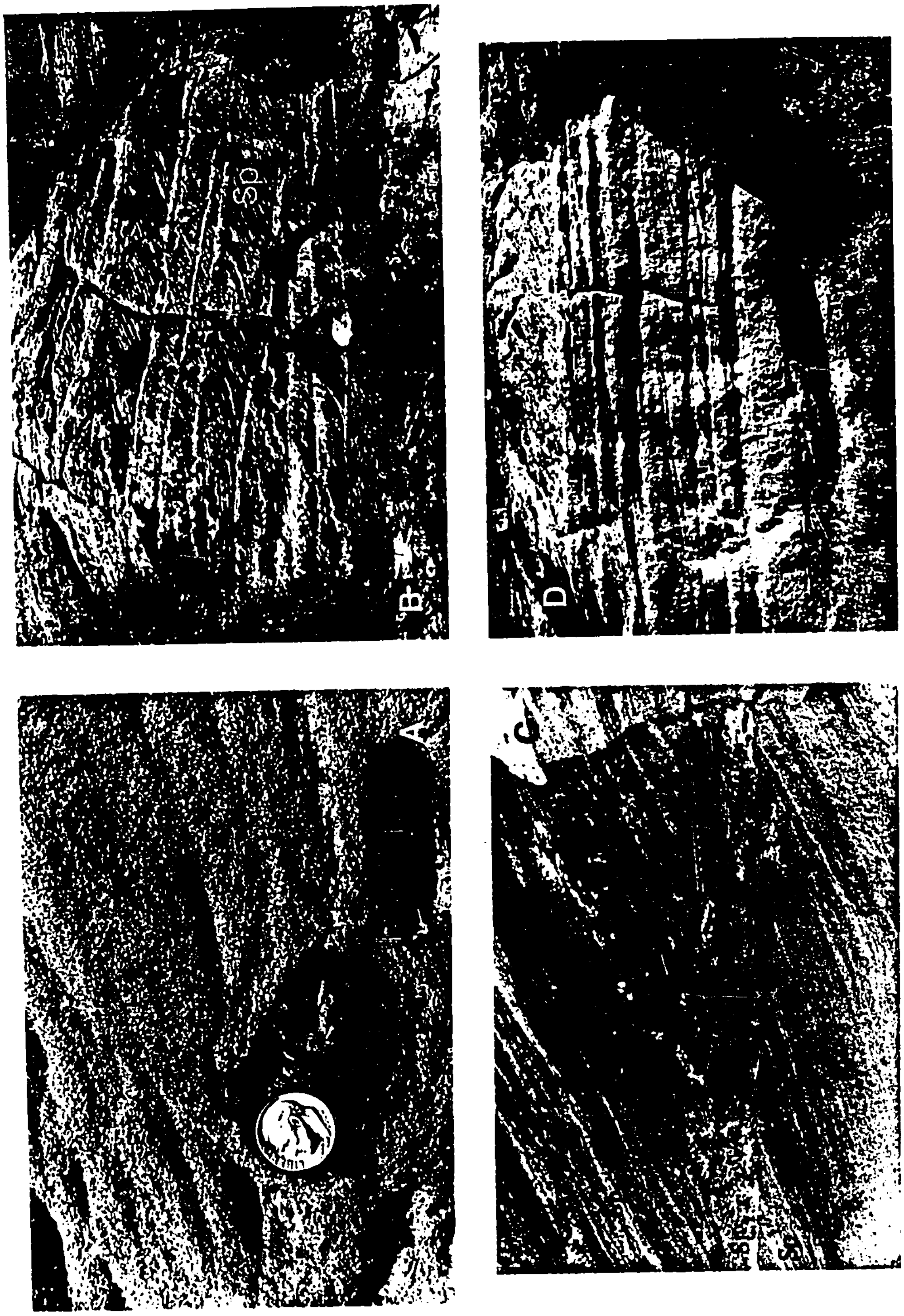
Two major facies associations have been identified in the type I channel: 1) the overbank facies association and 2) the channel facies association. The base of the channel facies association is defined by an erosional surface that is nearly horizontal but may show relief of several tens of centimeters. The surface is lined with a concentration of large goethite coated rip-up clasts (up to 10 cm in diameter) of argillaceous siltstones and/or argillaceous

Figure 14a. Goethite coated rip-up clasts

Figure 14b. Large scale planar foresets of type I channel; Sp = large scale planar foresets, Sr = ripple laminated sets; thickest Sp set is approximately 1m thick

Figure 14c. Counter current ripples and alternating coarse-fine laminations in planar foresets

Figure 14d. Reactivation surfaces (arrows point to examples) in type I channel


B
Sp
D
A
C

fine grained sandstone and rare pebble sized quartzite and chert clasts (Figure 14a). The nearly horizontal surface suggests a broad or laterally shifting channel. The mid-channel part of the channel facies association is characterized by a stacked sequence of simple foreset bars composed of quartz arenites and dominated by wedge to tabular sets of large scale planar foresets (Sp lithofacies of Miall, 1978) (Figure 14b). The foresets are straight with dips ranging from 26 to 33^{o} and exhibit a distinct grain separation resulting in alternating coarse and fine laminations (Figure 14c). Hunter (1985) interprets these foreset laminations to be formed by laterally extensive avalanching over a large slipface. Oversteepening or overturning in the down flow direction is common in the planar foresets. Multiple reactivation surfaces within individual large scale sets also occur (Figure 14d). Large scale trough cross beds (St lithofacies) are distinctly subordinate to the Sp lithofacies in the lower sets but dominate in the upper few meters of the channel sands and nearer the channel margins.

Underlying the planar cross bedded sets are 1 to 10 cm thick counter current ripple laminated units (Sr lithofacies) in which ripple sets have climbed the toes of individual foresets, producing an intertonguing between foreset and bottomset deposits (Figure 14c). Counter current ripples are a product of relatively strong separation eddies formed at the toe of an avalanche face of a migrating sand wave or dune (Reineck and Singh, 1980). There is a slight upwards coarsening from the Sr facies to the St/Sp facies. Counter current ripples (sometimes referred to as backflow ripples) have been reported from both tidal and fluvial deposits (Boersma et al., 1968; DeRaaf and Boersma, 1971; Hunter, 1985). Paleocurrent indicators are strongly unidirectional with deviation from the mean not exceeding 15 degrees in some midchannel exposures. Individual cross-bed sets range from 20 to 60 cm in thickness and attain lengths of 30 to 40 m indicating uniform water depths. Occasionally, the foresets are marked by uniform clay drapes (Fm lithofacies) a few cm thick which are often traceable onto the bar top (Figure 15a). Several clay-draped foresets may occur within one bedform and are often replaced with goethite (Figure 15b). The clay drapes represent the deposition of suspended load material during slack water conditions on essentially nonerosive reactivation surfaces. These features are restricted to the channel margin.

Evidence of nonperiodic compound bar development is present near the channel margin. Diagnostic features of compound bar development include broadly convex-up surfaces which forms on top of, and often cut, a series of simple

Figure 15a. Clay draped foreset; St = large scale trough cross bed, Fm = massive mudstone

Figure 15b. Goethite replaced mudstone (arrows) draping bar form; sitting figure in the shadow to the right for scale

Figure 15c. Compound bar; Fl = laminated mudstone, Sp = large scale planar foresets, Sl = low angle trough crossbedded sandstone, Sr = ripple laminated sandstone, St = large scale trough crossbedded sandstone


A
St
Fm
St
C

foreset bars (Figure 15c). The upward curved surfaces flatten out downstream and are usually lined by goethite-cemented rip-up clasts and have ripple reworked upper surfaces. Low angle trough sets with oblique paleocurent directions (Sl lithofacies) fill scours which are occasionally incised into the compound bar.

Tongue shaped sand flow structures of subaqueous origin, but typical of eolian deposits (Hunter, 1977), are locally present on 20 cm high foresets of bedforms along the west margin of a channel at Red Pryor anticline, south of Pryor Mountain (Figure 16a). The only known reference to subaqueous tongue shaped sand flows is by Buck (1985) from tidal sands of the lower Greensand (Early Cretaceous) of southern England. Buck (1985) suggests that subaqueous tongues shaped sand flows form when sedimentation rates are high but bedform migration is slow.

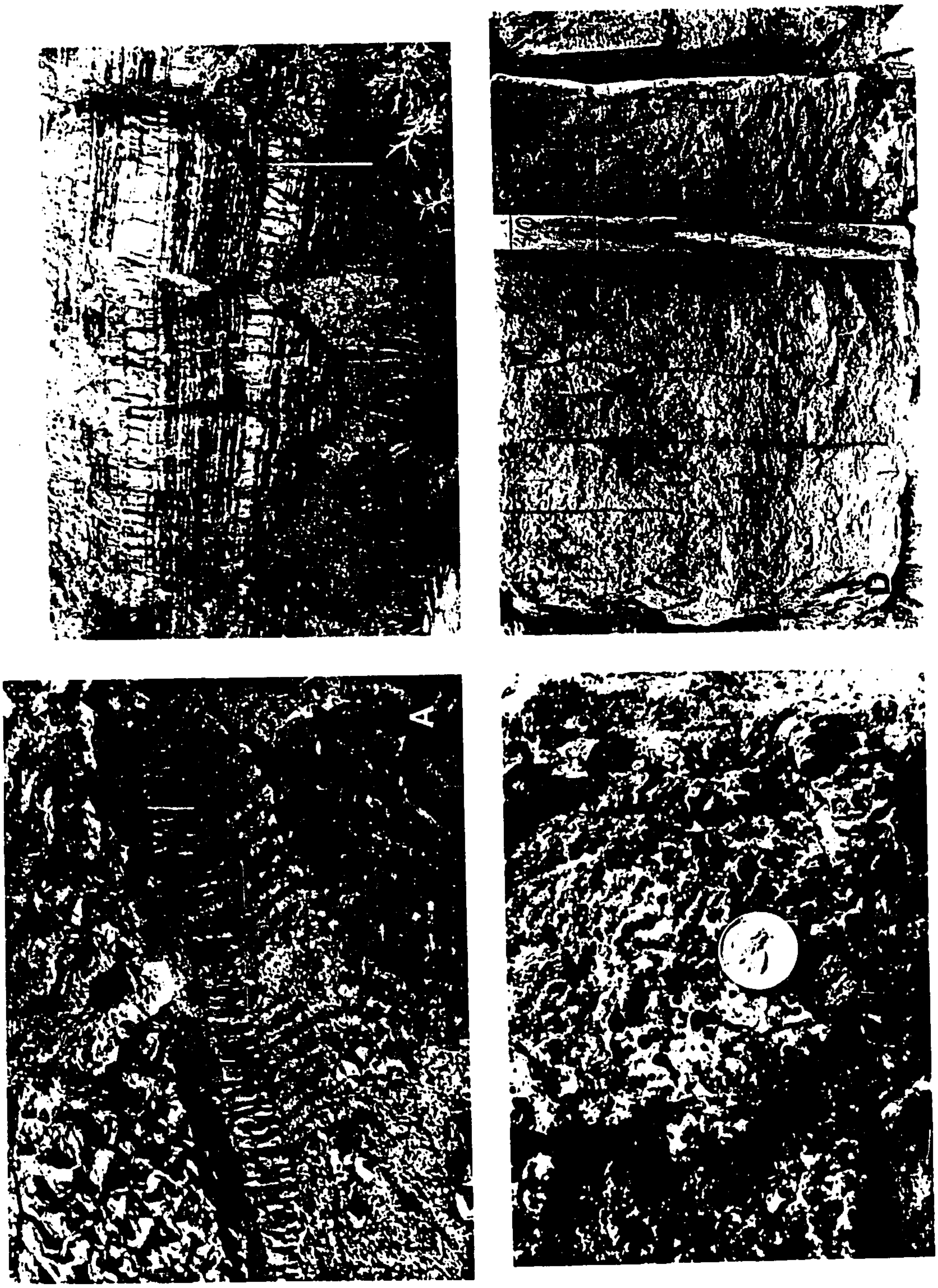
The channel facies association overlies, and is lateral to, the overbank facies association. This association consists of interbedded sandstones and nonresistant argillaceous siltstones. The sandstones are quartz arenites moderately well-sorted to moderately sorted, fine to very fine grained that commonly fine upwards to argillaceous siltstones or arenaceous siltstones (Figure 16b). The siltstones are locally organic-rich, fissile, and, in places, totally replaced by geothite and are generally quite

Figure 16a. Tongue shaped sand flows; ruler is 15cm long

Figure 16b. Interbedded sandstone and siltstone of overbank facies; Jacob staff is 1.5m long

Figure 16c. Burrows in goethite cemented siltstone

Figure 16d. Vertical escape burrows (fugichnia)


D
A

thin (1 to 5 cm) compared to the interbedded sandstones (2 to 45 cm). At one locality, small sand-filled mud cracks were observed in the siltstone indicating subaerial exposure. The dominant primary structures in both the sandstones and siltstones are ripple laminations. The corresponding lithofacies are Sr and Fl (laminated mudstones), respectively. Climbing ripples indicating high sedimentation rates are also preserved. The upper part of the sandstones commonly display flaser bedding that gives way to lenticular bedding in the siltstone which is typical of a waning flow sequence. The sandstone bodies are flat-bottomed with sharp basal contacts which are only slightly irregular. This coupled with fining upwards sequences and the uniform thickness and lateral persistence of the sand bodies over several meters suggests deposition of bed load material on very flat surfaces.

Burrows are very common in this facies association and provide evidence of rapid deposition of the sandstones. The siltstones and upper few millimeters of the sandstones are locally burrowed by both vertical and horizontal burrows of uniform diameter (2 mm) (Figure 16c). The density of burrowing in the main portion of the sandstone body, however, is much lower and the burrows are all vertical, and many extend completely through the sandstones essentially connecting older colonized surfaces (preexisting siltstones and upper few millimeters of the sandstones) with a newer stabilized surface (Figure 16d). This suggests that the vertical burrows represent upward directed paths of escape of some species of annelid or other types of organism through rapidly deposited sands (fugichnia). Across the overbank facies, the character of the lithofacies remains essentially the same from the proximal to distal positions except that the Fl facies increases in thickness distally so that the Fl:Sr ratio becomes greater than 1. The percentage of sandstone within overbank deposits proximal to individual type I channel complexes varies between channels even though channel facies associations remain constant. This may be indicative of variations in seasonal discharge between channels. Crocodilian dermal scutes and teeth plus fragments of turtle carapace were found associated with type I overbank mudstones. Also present within these sediments along the south flank of Rose Dome are dinoflagellates indicating brackish water conditions (E. Robertson, 1986, Phillips Pet. Co., pers. comm.).

Channel sandstones are often separated from overbank sandstones by arenaceous siltstones which occur just at the channel pinch outs and seem to interfinger with channel deposits yet rest unconformably on overbank sediments (Figure 17a). These sandy siltstones are poorly sorted, structureless and lack the bedding characteristics of either

Figure 17a. Type I channel pinch-out; arrow points to truncated overbank sandstone

Figure 17b. Multistorey type II channels at Sheep Mountain

Figure 17c. Multistorey type II channels at Alkali Creek


Ksm
B


C

the overbank or channel facies. They define the lateral limits of the channels and appear to be slump deposits of overbank material which had fallen into the channel and had not been reworked.

The second type of channel body (type II) forms multistorey, multilateral channel complexes which are characterized by well-developed lateral accretion surfaces (Figures 17b, c). These lateral accretion surfaces differ from those of the type III channel bodies in that they are marked by thinnly interbedded mudstones and sandstone rather than thick sandstones which fine upwards to thin siltstones. The type II lateral accretion surfaces also dip at 18 degrees, a higher angle than the 10 degree dip of the type III surfaces.

The sandstones are typically lenticular to wedge shaped bodies a few centimeters thick along the channel margins but grade into sheet type bodies towards the channel interior. The sandstones are mostly massive but faint ripple laminations were noted in some sets. In some areas, channel interiors are dominated by fine grained large scale trough cross bedded quartz arenites with set thicknesses on the order of 20-40 cm indicating deposition in large, sinuous crested to discontinuous crested sand dunes. The tops of these sets are often marked by dispersed fragments of coalified plant remains. The mudstones are likewise mostly massive, but faint ripple laminations are present indicating that at least some of the mudstones were transported or reworked as bedload material. Small scale slump features were also observed along the channel margin. The entire interval is burrowed.

Overlying the type II channel belt at Alkali Creek is a reddish brown, argillaceous, green mottled mudstone. This interval contains dinoflagulates indicative of brackish water conditions (E. Robertson, 1986, Phillips Pet. Co., pers. comm.). While individual channel fill deposits are 3-5m thick, the multistorey channel complex may locally be incised 20m(+) through the lower Himes channel complex (if present) and into the underlying Little Sheep Mudstone. Little overbank deposits are associated with these channel complexes. Excellent examples of these types of channels are found near Alkali Creek, west of Sheep Mountain (T54N R94W S12 and S13).

The third type of channel body (type III), like the type II channel, occurs in outcrops of erosive based, multistorey, multilateral channel complexes (Figure 18a). Channel bodies fine upwards from fine grained sandstones to siltstones, are typically 1-3m thick and characterized by lateral accretion surfaces dipping at 10 degrees signifying deposition in point bars. Along the west flank of Little Sheep Mountain (T55N R95W S22), north of Sheep Mountain,

Figure 18a. Multilateral type III channels at Little Sheep Mountain; arrows point to examples of lateral accretion surfaces; sandstone approximately 2m thick; photo by David Uhlir, Iowa State University

Figure 18b. Scroll bar development in type III channel; bedding plane view of Figure 18a; photo by David Uhlir

A

B

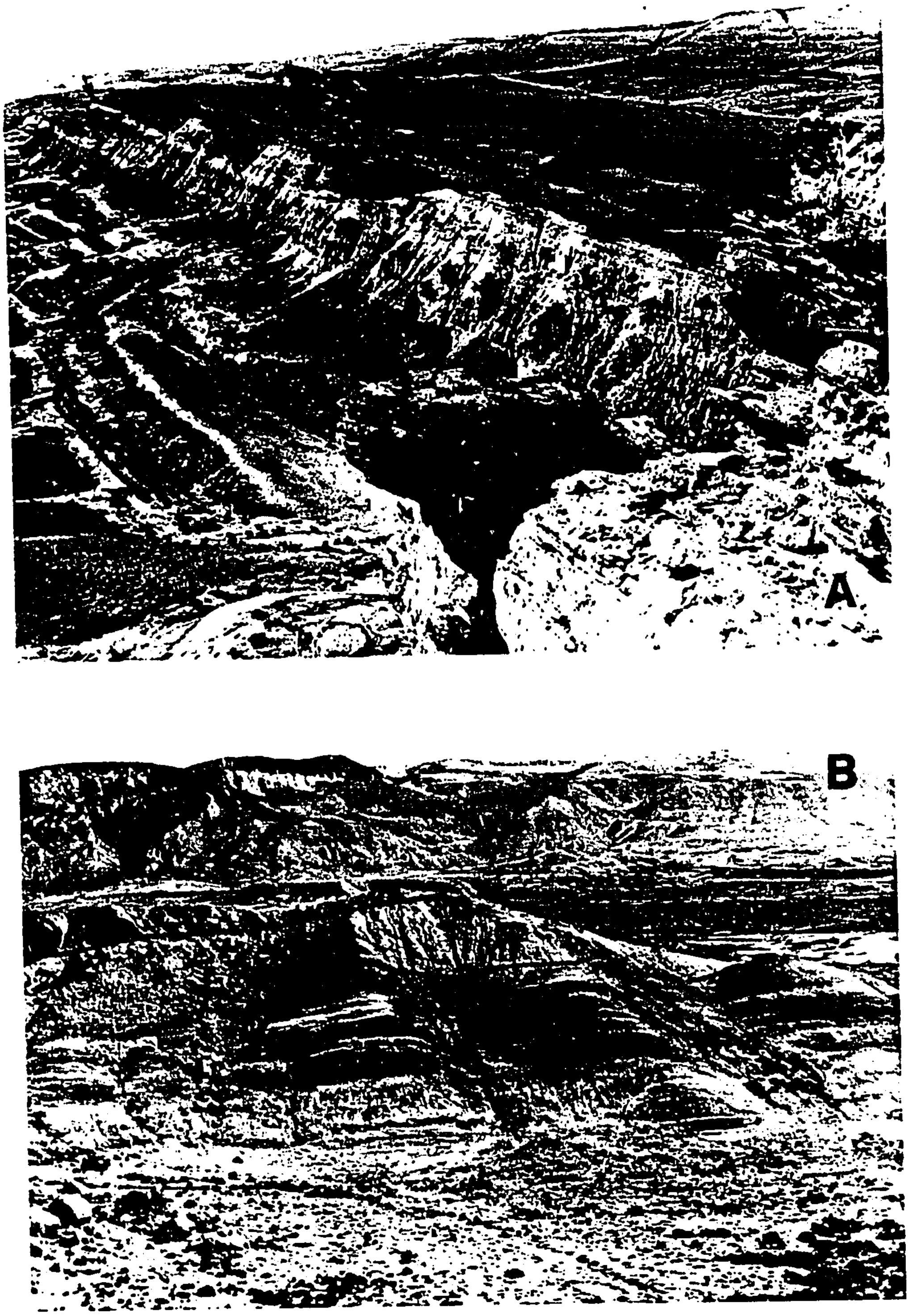
scroll bars are preserved on the upper surface of sandstone body (Figure 18b). The sandstones are mostly ripple laminated and confined to the point bar complex. Massive clay plugs are common and are sometimes capped by organic rich mudstones suggesting channel cutoff and oxbow lake development. Overbank deposits are generally not laterally extensive and are composed of argillaceous siltstones.

The relative ages of the three channel types of the Greybull interval are not clear. However, along the south flank of Rose Dome and the west flank of Little Sheep Mountain overbank sandstones and mudstones which are lateral to type I channels overlie a meandering channel deposit of a type III channel complex indicating that at least in these instances the high sinuosity type III systems predates the low sinuosity type I systems (Figure 19a). Relative ages of type I and type II channel complexes can be established at the Alkali Creek and west flank of Sheep Mountain exposures. In these areas type II channel complexes are incised by type I channels indicating that type II channels were deposited prior to type I channels (Figure 19b).

Conformably overlying the upper part of the Himes Member are the siltstones and interbedded darkly stained ferruginuous fine to very fine grained sandstones of the transitional marine Sykes Mountain Formation. The sandstones are flat bottomed, slightly erosive and laterally

Figure 19a. Type I overbank deposits (small arrows) overlying type III channel (large arrow); type III channel is one shown in Figures 18a and b

Figure 19b. Type I channel incised into type II channel

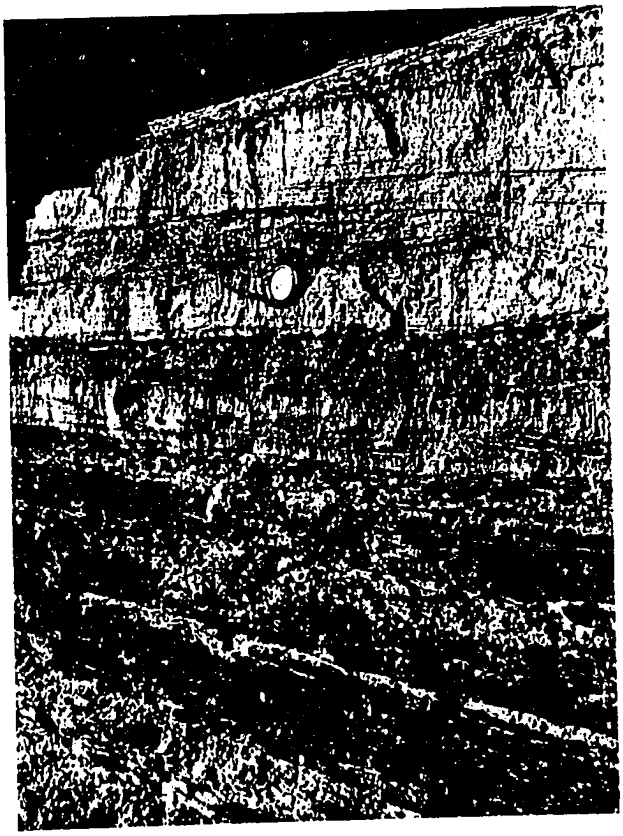

A
B

extensive for literally 100s of meters yet rarely exceed 1m in thickness. Dominant primary structures include parallel bedding (Figure 20a) with primary current lineations (PCL) on bedding plane surfaces, and low angle large scale trough cross bedding. Occasional herringbone cross stratification is also present (H. Soliman and N. Kandaker, Dept. of Earth Sciences, Iowa State Univ., pers. comm., 1986). Burrowing is intense in the upper few centimeters of the sandstones and on upper bedding plane surfaces resulting in the distruction of primary structures in the upper parts of the sands (Figure 20b). Upper surfaces of the sandstone often show evidence of reworking by wind or wave generated currents resulting in the formation of oscillation ripples, interference ripples or straight crested asymmetrical ripples which may bifurcate. Interbedded with the sandstones are sequences of mudstones which are typically lenticular to wavey bedded with fine grained sandstones which are dominated by hummocky cross stratification (Figure 21a) and often cut by large sand filled burrows up to 2cm across and several centimeters long. Some of the large burrows may have been generated by the pelecypods *Unio* (*Lampsilis*) *farri* or *Unio* (*Elliptio*) *douglassi* which are commonly found as shell lags at the base of the sandstones.

Coarse to medium grained channel sandstones are commonly found incised into the interbedded sandstones and

Figure 20a. Stacked sets of parallel bedded sandstone

Figure 20b. Close up view of Figure 20a showing burrowing



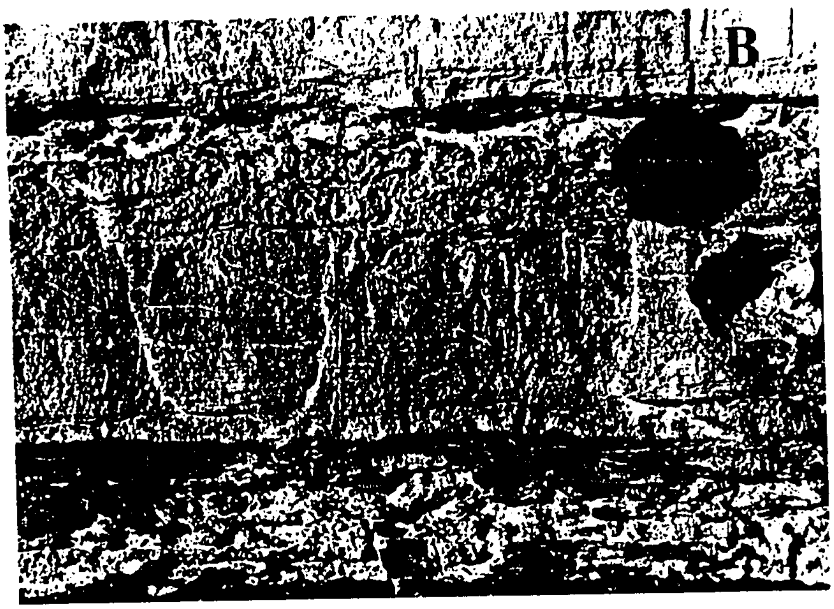
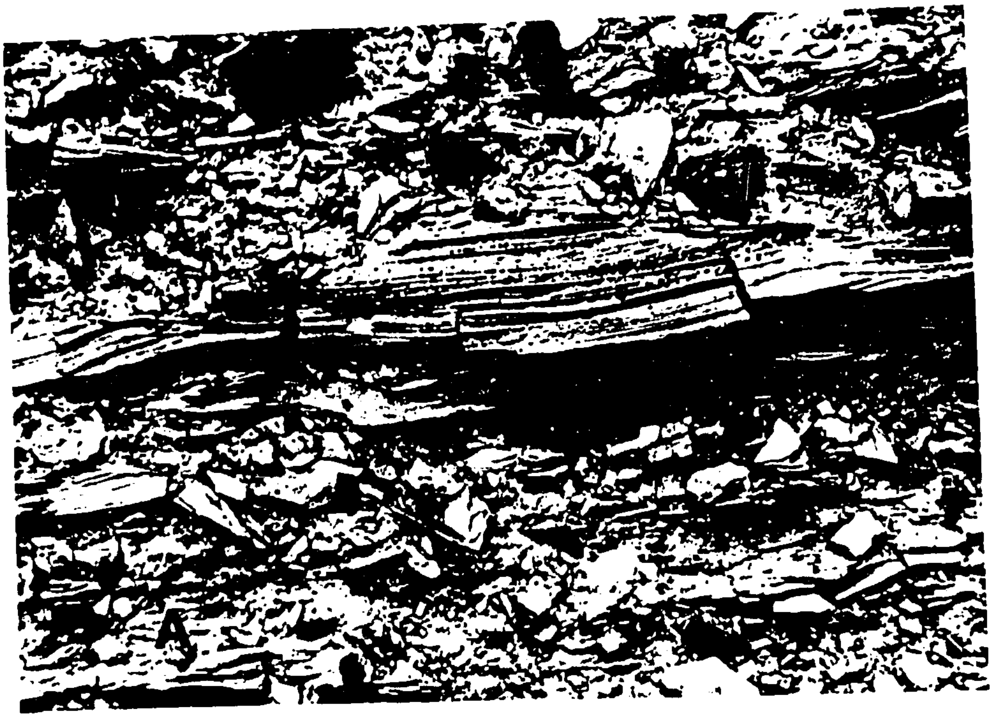

B

Figure 21a. Hummocky cross stratification in Sykes Mountain Formation

Figure 21b. Sykes Mountain Formation tidal flood channel overlying type I channel; arrows denote relative paleoflow directions of respective channels



B

mudstones. Maximum channel dimensions are not known, but several channel complexes 10's of meters across and 2m thick were observed just above the Sykes Mountain/Cloverly contact in the Sykes Mountain area just northeast of Lovell. The channels are dominated by large scale trough cross stratification which indicates dominant paleoflow was in a general easterly direction (Figure 21b). Soliman and Khandaker (pers. comm., 1986) also report herringbone cross stratification in these channels. Interestingly, paleoflows are opposite of those of the type I Greybull channels. The overall trend of the Sykes Mountain Formation is a fining upwards, coupled with a decrease in size of associated channels. The Sykes Mountain as a formation becomes much finer grained laterally from the Sheep Mountain area (west) to the Beaver Creek area (east).

Depositional Environment

The Greybull interval (upper Himes) is a complex sequence considered by previous workers to be totally nonmarine (Moberly, 1960; Shelton, 1972; Stone, 1983; Kvale and Vondra, 1985b). Kvale and Vondra (1985b) interpreted the type I channels within the Greybull interval to be Platte-type braided river deposits based on the similarity of their primary structures to those described in the modern Platte River (Smith, 1970; 1971; 1972; Miall, 1977; 1978; 1981; Blodgett and Stanley, 1980; Crowley, 1983). However, a totally fluvial interpretation does not completely explain many of the features and relationships observed within this interval. A more accurate interpretation of the Greybull interval is that of an estuary complex.

Estuaries form the boundaries between fluvial channel systems and tidal marine settings. They are essentially sediment traps with sedimentation occurring at the landward limit of a saltwater wedge which forms at the interface between fresh water and sea water (Davis, 1983; Emery, 1967). The transition from fluvial channels to estuarine channels is often marked by a decrease in channel sinuosity and an increase in channel depth and width (Clifton, 1982; Elliott, 1978). In a fluvial dominated estuary system this could be caused, in part, by the formation of a saltwater wedge which moves up an estuary during flood tides and which restricts the normal fluvial discharge (Caspers, 1967; Guilcher, 1967). The backed up flow is released during ebb flow which results in an increase in flow energy beyond that generated by normal fluvial discharge. The result is channel straightening and scouring coupled with bank erosion.

Land and Hoyt (1966) noted the similarity of depositional processes and products of fluvial and estuary systems. Davis (1983) states that by definition an estuary is subject to tidal current activity and most papers which document ancient estuary deposits in the rock record have stressed the importance of tidal processes in their interpretations.

While the evidence of tidal currents is not well-documented in the Greybull interval, certain inferences about their importance within this sequence can be made. The Greybull interval is interpreted to have been deposited in a coastal plain environment because of its stratigraphic proximity to, and conformable relationship with, the overlying Sykes Mountain Formation. The Sykes is a shallow marine shelf deposit laid down during the Early Cretaceous marine transgression into northern Wyoming. The repetitious sequences of wavey to lenticular bedded mudstones (in which the sandstones are dominated by hummocky cross stratification) which coarsen upwards to thin, laterally continuous, flat bottomed sandstones characterized by low angle large scale trough cross bedding, primary current lineations (PCL), and intense burrowing on sandstone body tops indicates deposition in a storm dominated setting, possibly as off shore bars that were reworked during the marine transgression (Reineck and Singh, 1980). The occurrence of herringbone structure in channel bodies which cut the laterally continuous sandstones shows that tidal processes were also important in transporting sediments. Reconnaissance studies of the large scale trough cross bedding of the Sykes channels shows that the dominant paleocurrent was directed towards the Early Cretaceous shore indicating that these are tidal flood channels.

An examination of the channels within the Greybull interval suggests that tidal currents operated inland as well. Specifically, the differences in bedforms, overall grain size and general character of the lateral accretion surfaces (LAS) between the type II and type III channels indicates that these channels formed in different environments. Steeply dipping ($>10^{\circ}$) LAS with mudstones of relatively uniform thicknesses are more typical of point bars formed in tidal channels (Elliott, 1978; De Mowbray, 1983; Smith, 1985). LAS facies which lack mudstones or have mudstones of irregular thickness deposited on shallowly sloping surfaces ($<10^{\circ}$) are generally recognized in point bars formed in fluvial or upper delta plain deposits (Smith, 1985). This suggests that type II channels are tidal creek deposits and type III channels are fluvial meander channel deposits. Maceral studies completed on type III overbank sediments support a totally fluvial interpretation for this channel (E. Robertson, 1986, Phillips Pet. Co., pers. comm.) however these studies have yet to be performed on the type II channel deposits.

The type I channels which were originally interpreted to be sandy braided river deposits (Kvale and Vondra, 1985b) are now considered to be deposited in an upper estuary setting. The braided stream model was deemed inappropriate for three reasons. Firstly, various authors studying modern Platte-type braided channel deposits have noted the high divergence of flow indicators within the braid channel bedforms and deposits (Smith, 1972; Cant and Walker, 1976; Cant, 1978; Walker, 1981). This contrasts to the type I Greybull channels which have a low paleocurrent divergence. This is especially pronounced within the channel interior. The type I sandstone bodies are essentially low sinuosity channel systems filled with sands deposited by straight crested transverse bars with a general absence of braid channel development. Secondly, it is difficult to generate a model which would explain the formation of a low sinuosity sandy braided channel underlain by high sinuosity fluvial and tidal creek deposits and overlain by a marine unit without the occurrence of a regional unconformity. Thirdly, the most important bit of evidence to support a marine influence in the Greybull interval is the occurrence of dinoflagellates in the mostly massive mudstones which are lateral to the type I channel sandstones, indicating brackish water conditions.

The primary structures described from the type I channel deposits can be explained in terms of deposition within an upper estuary setting and the lateral mudstones and sandstones within a tidal flat marsh environment. Strongly unidirectional, large scale planar foresets are commonly reported from modern estuaries where ebb currents dominate (Land and Hoyt, 1965; DeRaaf and Boersma, 1971). The large scale planar foresets are the internal structural remains of simple transverse bars (Allen, 1980). Features associated with these large scale planar foresets include counter current ripples, ripple reworked bar tops, tongue shaped sand flows, reactivation surfaces and clay draped reactivation surfaces. While not ubiquitous, these structures are very common to tidal estuaries where, with the exception of the counter current ripples, they form as a result of reversing or slackening currents caused by tidal currents. The latter four are more common nearer the channel margin either because they formed more readily there or else their preservation potential was higher nearer channel margins.

An upper estuary setting is invoked for the type I channels because of the lack of bimodality to the paleocurrent trends. With the exception of the clay draped reactivation surfaces, tidal flood currents were of insufficient strength and duration to produce features which escaped ebb flow erosion. Most reactivation surfaces are not mud draped and the absence of a drape probably caused many reactivation surfaces present in outcrop to be missed.

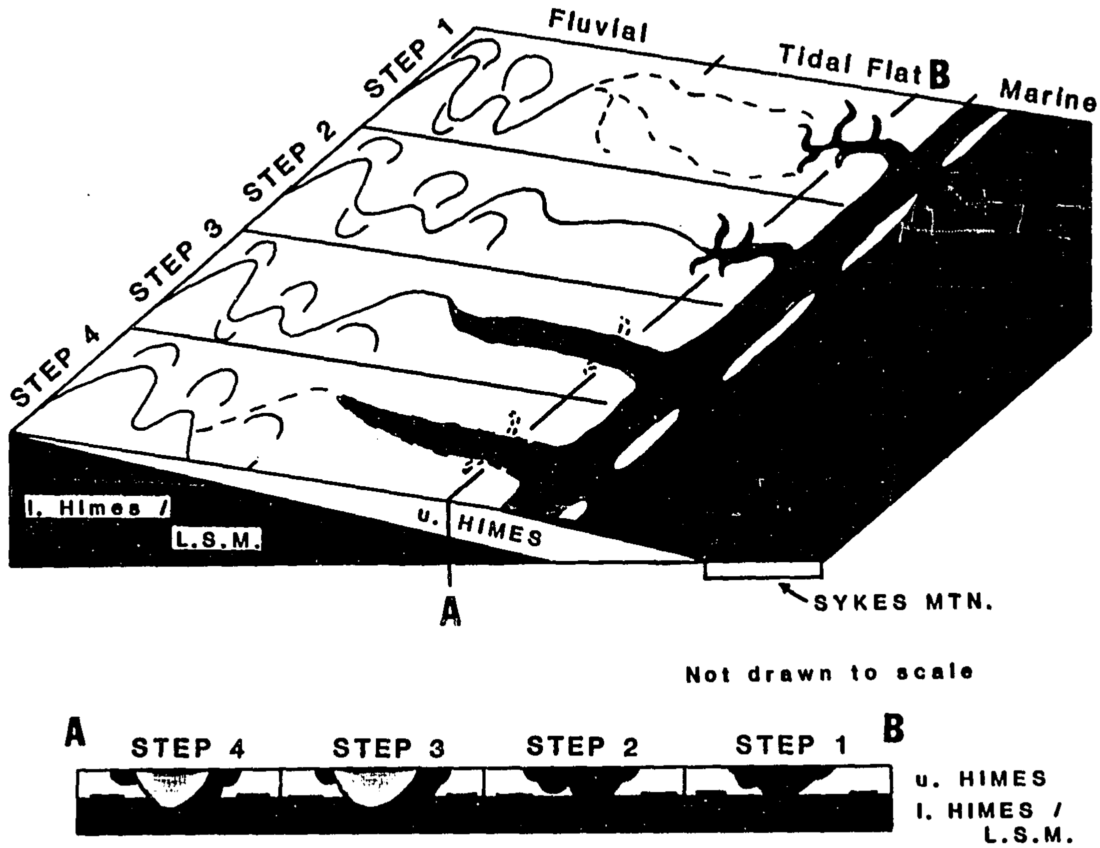
The reactivation surfaces should be reexamined and measured utilizing the technique of Allen (1981) to determine if any periodicity exists which would document phases of neap-spring tidal currents.

Several points are worth emphasizing before a reconstruction of the Greybull estuary complex is possible. First of all relative ages of the three channel types are important. Type I channels were shown to be the youngest and are interpreted to be upper estuary deposits influenced by weak tidal currents. Type II channels, whenever present, are always cut by, and in some instances almost completely removed by, type I channels suggesting that the positioning of type II channels somehow controlled the localization of type I channels. Type II channels are very similar to tidal creek deposits. Type III channels are demonstratably older than type I channels, are fluvial in origin and are the smallest of the major channels types. Secondly, dinoflagellates indicating brackish water conditions are present in mudstones overlying types I and II channels and in mudstones lateral to type I sandstone bodies.

The sedimentary structures and channel bedforms of the Greybull interval are very similar to those described by Campbell and Oaks (1973) for estuary deposits of the Lower Cretaceous Fall River Formation of the Powder River Basin. The type I channels which resemble the upper estuary deposits of Campbell and Oaks are, in part, the stratigraphic equivalents. The Fall River lower estuary deposits strongly resemble the shallow shelf deposits of the lower Sykes Mountain Formation.

Campbell and Oaks (1973) describe the formation of the Fall River estuaries as being initiated by erosion of tidal flats and landward marsh deposits by small streams. Tidal currents widened the mouth of the estuary forming a fan shaped scour which had its narrow end towards land. Sediment infilling was then accomplished by fluvial current processes landward which gave way to tidal current processes down the estuary.

The growth of Greybull estuaries was somewhat similar (Figure 22). The formation was probably initiated by an avulsion event of a meander channel (type III) on the coastal alluvial plain which then sought a new course across the tidal flats and marsh deposits. Initially, flow would have been unconfined on the tidal flats, depositing sands and muds. Eventually, erosion and incision, perhaps in part controlled by preexisting tidal creek channels (type II), confined the flow and established an estuary. Channel widening occurred, with erosion of the estuary probably occurring fairly rapidly with infilling perhaps not taking place until the mouth was somewhat restricted by wave or tide generated shoals. Infilling then occurred with upper



Step 1: Avulsion of fluvial channel (type III) with sheet flood onto tidal flat.

Step 2: Establishment of estuary channel (type I) on tidal flat; path possibly controlled by preexisting tidal creek (type II).

Step 3: Estuary channel (type I) widened and deepened by interaction of fluvial and tidal currents.

Step 4: Abandonment of estuary channel caused by avulsion of fluvial channel (type III) upstream.

Figure 22. Postulated paleoreconstruction of Greybull estuary development

estuary sediments (type I) being deposited by fluvial dominated processes and lower estuary sediments being brought in by tidal currents. The trough cross bedding at the top of the type I channels is probably related to channel shallowing and the formation of large scale 3 dimensional dunes.

The dominance of mudstones in the tidal flat areas and the occurrence of tidal flood channels, which are incised into storm deposited, laterally continuous sandstones suggests that the Greybull shoreline was barred with offshore bars cut by tidal inlets. By virtue of the marine transgression the preservation potential of the bars and channel inlets was low. Only that part of the inlet channel which had scoured below the erosion base was preserved from being reworked with the bars into sand sheets. Additional work on the Sykes Mountain Formation needs to be done to confirm this interpretation, but a barred coastline with a microtidal range (<2m) could explain why flood tides were so weak in the upper estuary (Davis, 1983).

The dominant seaward directed foresets of the type I channels coupled with their grain size similarity to type III channels indicates that the principal source source of sand sized sediment was to the east (landward). However, at least some, if not most, of the Sykes Mountain sediment was derived from the west and transported by longshore currents.

Along the south flank of Rose Dome, a Sykes Mountain flood dominated channel contains coarse to very coarse sand and granule sized clasts. The coarsest sand that any of the Greybull channels could have supplied is medium sand. A western source also explains the eastward fining of the Sykes Mountain Formation across the northern Bighorn Basin.

PETROLOGY

Clay Mineralogy

Mineralogic studies of the clay sized fraction of the mudstones from the Morrison and Cloverly formations were conducted on samples collected from the Sheep Mountain and Beaver Creek areas. X-ray diffraction patterns were generated for 12 Morrison and 22 Cloverly samples from oriented clays mounted on glass slides. In addition, 3 samples from the Sykes Mountain Formation and 2 samples from the Sundance Formation and were also analyzed. The results are summarized in Table 2. The Sykes Mountain samples are mudstones collected by Mr. Hosni Soliman, Iowa State University, from lower, medial, and upper positions within the Sykes Mountain Formation. The basal Sundance sample was collected by Mr. David Uhlir also of Iowa State University.

In general, the formations consist of varying amounts of kaolinite, mixed layer montmorillonite-illite, illite and montmorillonite. The monomineralic montmorillonite is restricted to the dark mudstones of the Little Sheep Mudstone Member, the clays associated with the lower Himes channel sandstones, a unit interpreted to be an altered tuff in the lower Morrison and an altered tuff collected by D. Uhlir from the lower Sundance. The kaolinites show maximum XRD peak development in channel and channel overbank deposits associated with the Greybull interval but are not

Table 2. Summation of clay mineralogy

SAMPLE	FM.	DOM. CLAY MINERAL	DEPOS. INTERP.
BCAS 1c	uJrs	mo/il, il	clay drape (tidal)
BCAS2c	lJrm	mo/il, il, ka	lagoonal muds
BCAS2 4	lJrm	il	tidal flat (?)
BCAS2 9	lJrm	mo/il, il	tidal flat (?)
BCAS2 11a	lJrm	mo/il, il	tidal flat (?)
BCAS2 11c	lJrm	mo/il, il	tidal flat (?)
BCAM 1	uJrm	mo/il, il	fluv. ch. overbank
BCAM2 2a	uJrm	mo/il, il, ka	fluv. ch. overbank
BCAM 11	uJrm	il	fluv. ch. overbank
SCC 1	uJrm	mo/il, il	fluv. ch. overbank
SCC 2a	uJrm	il	fluv. ch. overbank
SCC 3	uJrm	il	fluv. ch. overbank
BCA1 11	LKcl	mo/il, il, ka	chert cgl. ch.ovbk.
NN3 1	LKcl	mo/il, ka	playa mdflat proximal to ephem. ch.
NN3 7a	LKcl	il, ka	playa mdflat (red-purple in color)
WB1 1	LKcl	mo/il, il	chert cgl. ch.ovbk.
WB1 10	LKcl	mo/il, il	ephemeral channel
NN1 7b	LKcl	mo/il, il	chert cgl. ch.ovbk.
BCA2 7	UKcl	mo/il, il, ka	lower Himes ch.
BCA2 10	UKcl	mo/il, ka	lower Himes ch.

SAMPLE	FM.	DOM. CLAY MINERAL	DEPOS. INTERP.
NN1 11b	UKcl	mo/il, il, ka	paleosol sep. u. Himes and l. Himes
PFS B	UKcl	ka, il	type III ch. ovbk.
L.S. Mtn. Recon.	UKcl	mo/il, il, ka	type I ch. pinchout
Bull I 1a	UKcl	mo/il, il, ka	type I ch. overbank
WB1 10.1	UKcl	mo/il, il, ka	type I ch. overbank
NN1 11a	UKcl	mo/il, il, ka	type I ch. overbank
BCAS2 12	UKcl	mo/il, il, ka	type I ch. overbank
Sykes 1	lKsm	il, ka	marine shallow shelf
Sykes 2	mKsm	mo/il, il, ka	marine shallow shelf
Sykes 3	uKsm	mo/il, il, ka	marine shallow shelf
BHR1 4	lJrs	m, mo/il, il	altered tuff
BCAS2 6	lJrm	m, mo/il, il	altered tuff on tidal flat
BCA2 3	LKcl	mo/il, il	playa mudflat (lt. grey in color)
BCAM 1.1	LKcl	m, mo/il,il, ka	playa mdflat (black)
WB1 6	LKcl	m, mo/il, il	playa mdflat (grey)
WB1 7	LKcl	m, mo/il, il	playa mdflat (black)
WB1 8	LKcl	m, mo/il, il	playa mdflat (black)
WB1 11	LKcl	m, il	playa mdflat (grey)
BCA2 8	UKcl	m, mo/il, il, ka	lower Himes ch.

restricted to those settings. The illite and mixed layered clays consisting of montmorillonite-illite are found throughout the Morrison and in association with overbank deposits of the channel sandstones in the Cloverly. In addition, the 3 Sykes Mountain samples plus the uppermost Sundance sample also contain these clays. The mixed layered montmorillonite-illite superlattice is both randomly interstratified and "half ordered" (Reynolds and Hower, 1970, p. 34). Utilizing techniques described by MacEwen and others (1972) and comparing observed XRD patterns to XRD patterns calculated by Reynolds and Hower (1970), it appears that montmorillonite layers constitute more than 50% of the superlattice with some samples exceeding 80% montmorillonite.

Clay mineral assemblages in sedimentary rock units are controlled principally by source rock compositions, post burial diagenetic processes, and the original sedimentary depositional environment which was influenced by climate, topography and regional or local tectonics (Blatt et al., 1972; Singer, 1980; 1984). While it is tempting to interpret the clay mineral assemblages in the Morrison and Cloverly as a function of paleoclimatic factors, this is impossible to do until the impact that nonclimatic factors may have had on the generation of the various clay species are better understood. At present, the paleoclimatic significance of the Morrison and Cloverly mudstones can only be discussed in general terms. However, Mr. Christos Mantzios, Iowa State University, is conducting research on the geochemistry and petrography of these mudstones in an attempt to better understand the significance of these units.

The most important aspect of the monomineralic montmorillonites in the Little Sheep Mudstone and lower Himes members of the Cloverly Formation is its suggestion of a significant volcanic input into the Bighorn Basin during the earliest Cretaceous. Moberly (1960) suggested that volcanic activity to the west increased from the Late Jurassic to The Early Cretaceous. He considered the bulk of the Cloverly Formation to consist of altered volcanic debris. Evidence of the deposition of volcanic material includes not only the montmorillonite clays but also the occurrence of abundant silica nodules, devitrified siliceous tuffs and volcanic rock fragments found throughout the Little Sheep Mudstone and lower Himes units.

It is suggested that Moberly's belief that volcanic activity increased from Late Jurassic to Early Cretaceous is perhaps not correct, but rather the preservation potential of volcanic material increased over this interval of time. Volcanic activity during Late Jurassic is supported by the occurrence of thin monomineralic montmorillonite layers in the Morrison and Sundance formations. The occurrence of the mixed layered montmorillonite-illite in the Morrison may also be the remnants of volcanic ash. The presence of illite could simply be a function of increased leaching of montmorillonite on an alluvial floodplain in a wetter climate than existed during Cloverly times (Singer, 1984). This is further substantiated by the occurrence of a similar clay mineral association in channel overbank deposits of the Cloverly, suggesting that the association is related to depositional setting. The formation of illite, however, could also be a function of post depositional diagenetic processes related to burial (Blatt et al., 1972).

Encased in the dark argillaceous mudstones of the Little Sheep Mudstone Member of the Cloverly are lenticular to equant nodules and veinlets composed of barite, calcite and silica. The larger nodules are composed predominantly of carbonate although smaller nodules consisting of barite or silica were noted. Maximum size of the nodules is up to 3m long and 1m thick. Veinlets can be several cm's across. Most of the larger nodules have mineralized septarian fractures, while the smaller nodules may have surface structures consisting of small spheroids a few mm's across of fibroradiating calcite crystals. Although these nodules were not examined in thin section they appear to have grown displacively.

Formational sequences of the minerals which comprise the larger nodules were deduced from field observations and is as follows:

1. formation of finely crystalline calcite nodule
2. formation of fractures
3. formation of pore lining fibrous calcite with crystal orientation perpendicular to pore walls
4. formation of reddish-brown stained, noncalcareous polycrystalline mineral (barite?)
5. deposition of amorphous silica
6. formation of coarsely crystalline calcite (sometimes dog tooth spar)
7. deposition of euhedral prismatic quartz crystals

The entire diagenetic sequence is not always present in every fracture, but a general formation of calcite prior to silica with a final phase of carbonate deposition is generally present. Veinlets not associated with the nodules lack the initial carbonate phase but consist of amorphous silica lining the outer walls of the veinlets with large subhedral calcite crystals forming the core of the veinlet.

Sandstone Mineralogy

Limited macroscopic and microscopic examinations of sandstone samples of the Cloverly Formation were conducted. Point counts were performed on sandstones collected from the four major Cloverly channel complexes in the Beaver Creek area. Counts ranged from 300 to 100 points per slide and minerals were described utilizing the nomenclature of Dickinson (1970). Each sample was described in terms of its percentage of monocrystalline quartz (M); polycrystalline quartz which includes cherts, polycrystalline quartz of metamorphic origin, chalcedony, etc. (C); plagioclase (P); orthoclase (O); total unstable aphanitic rock fragments (L); volcanic rock fragments (V); interstitial constituents which include cementing agents such as calcite, clays, etc. (I); and porosity (PS) (Table 3).

The dark chert bearing channels (first 5 samples) are characterized by abundant chert fragments. Many of the chert fragments contain euhedral iron oxide-stained dolomite rhombs (Figure 23a) and probably formed as a replacement of limestones. Sponge spicules were also observed in chert fragments (Figure 23b). Detrital chalcedony is also present including the zebraic variety which has been described as forming in association with evaporite minerals (Scholle, 1979). Detrital grains of monocrystalline quartz which show quartz overgrowths also occur (Figure 23c). All of these siliceous resistates suggest a Paleozoic sedimentary source. Pebble to boulder sized clasts of dark chert were found which contained bryozoan, coral and brachiopod remains which further substantiate a Paleozoic sedimentary source for the dark Clovery cherts. Moberly (1960) and Hooper (1962) also interpreted a Paleozoic source for the dark chert clasts of the Cloverly.

The dominant cementing agents for the dark chert bearing units investigated are calcite and pore lining

Table 3. Mineralogical composition of selected Cloverly Formation sandstones

SAMPLE	M[a]	C[a]	P[a]	O[a]	L[a]	V[a]	I[a]	PS[a]	TC[a]	UNIT
BCA-4	23[b]	39	<1	0	<1	0	21	16	300	chert cgl.
BCA-6m	45	23	2	0	0	0	4	27	300	chert cgl.
BCA-7p	26	46	<1	0	0	0	6	21	300	chert cgl.
BCA-8t	34	45	0	0	0	0	6	15	300	chert cgl.
S2-2	26	8	1	1	<1	33[c]	29	2	300	chert cgl.
S2-7	10	27	22	6	5	4	23	2	300	l. Himes
S2-25GM	8	29	28	2	4	3	24	2	300	l. Himes
S5-18	49	16	2	0	0	0	7	26	100	type I ch.
1B-8	65	18	0	1	0	1[d]	3	12	100	type I ch.
1C-5	45	15	0	3	4	2	10	20	100	type I ch.
1C-1	52	17	1	0	0	1[e]	6	20	100	type I ch.

[a]M = monocrystalline quartz, C = crypto- or polycrystalline quartz, P = plagioclase, O = orthoclase, L = polycrystalline aphanitic rock fragments (clastic rock fragments, tectonites, microgranular rock fragments), V = volcanic rock fragments, I = interstitial constituents, PS = porosity, TC = total number of grains counted per sample

[b]Expressed as number percent.

[c]percentage of sedimentary rock fragments

[d]percentage of heavy minerals and muscovite

[e]percentage of heavy minerals

Figure 23a. Photomicrograph of Fe-stained, silica replaced dolomite rhombs in chert

Figure 23b. Photomicrograph of sponge spicules in chert

Figure 23c. Photomicrograph of quartz overgrowths


0.2 mm
B
A
25 µm
C
100 µm

clays. The calcite formed during the weathering process which concentrated it at the surface of the outcrop. Minus cement porosity (PS + I) is approximately 30 to 35% for these units. Feldspars are uncommon to rare and include plagioclase, some of which shows albite twinning, microcline and orthoclase.

The lower Himes sandstones are characterized by abundant plagioclase with the Ca-rich varieties sericitized (Figure 24a). Dissolution features of the feldspars are fairly common (Figure 24b). Volcanic rock fragments are also present (Figure 24c). Diagenetic pore lining and pore filling clays are abundant. Porosity is very low.

The sandstones associated with the Greybull interval are characterized by abundant quartz. Polycrystalline quartz of possible metamorphic origin, whose individual subcrystal boundaries are relatively straight between elongate to equant grains, is common (Figure 24d). Monocrystalline quartz, some of which exhibits quartz overgrowths are also common. Chert grains are rare. Microcline, orthoclase and albite twinned plagioclase are present. Authigenic kaolinite lines pores and appears to be the dominant cementing agent in fresh samples (Figure 24d). Porosities are high.

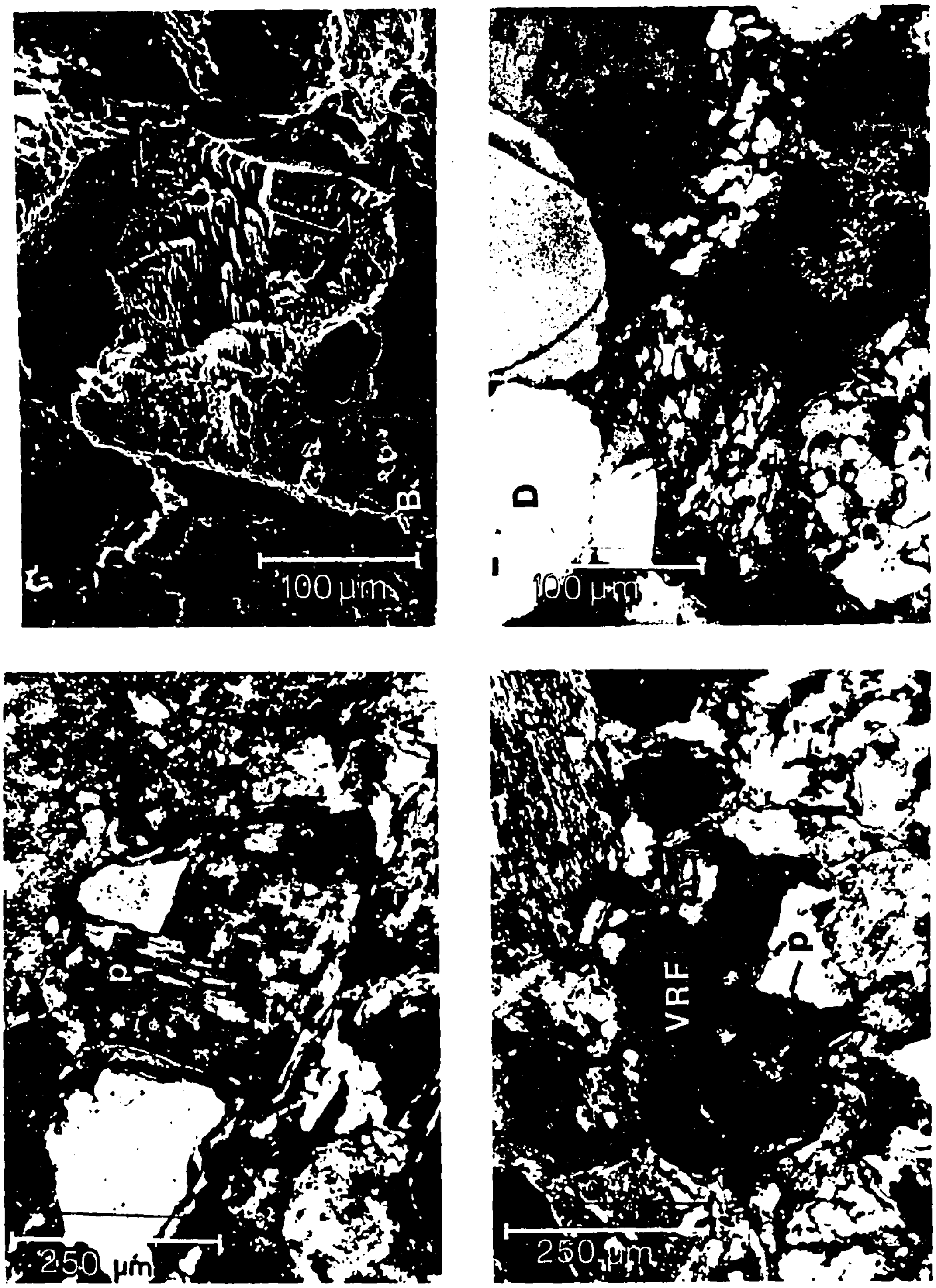
Although limited in their scope, the analyses performed on the Cloverly sandstones tend to support the sandstone

Figure 24a. Photomicrograph of sericitized plagioclase (p)

Figure 24b. SEM photomicrograph of dissolution features in feldspar

Figure 24c. Photomicrograph of volcanic rock fragment (VRF) with plagioclase (p)

Figure 24d. Photomicrograph of quartz grain of possible metamorphic origin and pore lining authigenic kaolinite



B
100 µm
D
100 µm
p
250 µm
VRF
p
250 µm

petrologic studies of Moberly (1960). Heavy minerals were observed in every sandstone sample examined in thin section. Previous heavy mineral studies on the Morrison and Cloverly were done by Moberly (1960), Hooper (1962), MacKenzie and Poole (1962), MacKenzie and Ryan (1962), Mirsky (1962) and Chisholm (1963). Hooper (1962) discusses the heavy mineral assemblages of the Cloverly dark chert bearing units along the Casper arch. Mirsky (1962) and Moberly (1960) discuss heavy minerals associated with the Morrison and Cloverly in the Bighorn Basin. The others are broad regional studies. None of the previous workers recognized the importance of intraforeland sediment sources and their studies were based on the misconception of the importance of the western fold-thrust belt as a source area. The petrography of the Morrison and Cloverly as a result needs to be reexamined. In general, however, it appears that both the Morrison and Cloverly contain (among others) zircon and tourmaline with the Morrison having an additional component of garnet.

REGIONAL STRATIGRAPHY

The foreland basin that existed during the Late Jurassic and Early Cretaceous in northern Wyoming was part of a much larger basin which extended from Mexico northward into Canada (Dickinson, 1976). The eastern margin of the basin, which is now occupied by the Black Hills, was then a regional syncline receiving sediments from the Sioux Uplift in southwest Minnesota, southeastern South Dakota, and northwestern Iowa (Bolyard and McGregor, 1966; Witzke *et al.*, 1983). Isopach maps of the Lower Cretaceous Inyan Kara Group indicate a westward thinning of sediments away from the Black Hills towards the central Powder River Basin (Bolyard and McGregor, 1966).

The western margin of the foreland basin was receiving sediments derived from highlands to the west (Eyer, 1969; Furer, 1970) which were the result of uplift along the Sevier Idaho-Montana-Wyoming fold-thrust belt. Flexure bending caused by the tectonic stacking of imbricate thrust sheets associated with the orogeny and the loading of sediment derived from the erosion of the advancing thrust sheets produced an asymmetric foreland basin (Jordan, 1981; Wiltschko and Dorr, 1983). This was filled with an eastward thinning sequence of Upper Jurassic-Lower Cretaceous sediments.

The middle part of the northern Wyoming, Late Jurassic-Early Cretaceous foreland basin, in what is now the Bighorn Basin, received sediments derived from multiple source areas including the fold-thrust belt, the craton (Sioux Uplift) and intrabasin uplifts. Volcanic vents located in the northern Absaroka-Yellowstone region combined with possible arc associated vents, located west of the fold-thrust belt, in central Idaho, to supply volcanic ash to the area. This conclusion is based on paleocurrent measurements, comparative lithologic analyses of sandstone bodies, and on the correlation of units of the Bighorn Basin Upper Jurassic-Lower Cretaceous nonmarine sequence with those in eastern and western Wyoming and adjacent areas. The importance of, and evidence for, these source areas is discussed in greater detail below.

Several problems become apparent when comparing coeval deposits across the northern Wyoming foreland basin. Despite volcanic vents supplying ash to the foreland basin, probably via eolian processes, streams that drained the fold-thrust belt, intrabasin uplifts and the Sioux Uplift carried detritus derived from Mesozoic sedimentary units, Paleozoic carbonates and/or Precambrian crystalline rocks. After transport distances of several hundred kilometers, detritus consisting of mostly siliceous resistates and clay minerals were deposited. Even though source areas were separated by many hundreds of kilometers and represent very different tectonic settings, the resulting lithologies were similar. The result was a complicated interfingering relationship between eastern, western and intrabasin derived, mostly fluvial, sequences within the region of the Bighorn Basin. In addition, the usual rapid lateral and vertical variability typical of fluvial deposits, the general lack of fossils, and the misuses of stratigraphic nomenclature (both formal and informal) have further complicated correlations across the foreland basin. Certain formational correlations are possible, however, and Figure 25 and Table 4 summarizes them.

A comparison of lithologies of the Morrison and Cloverly formations of the eastern Bighorn Basin with those of the Lakota Formation (after Bolyard and McGregor, 1966) of eastern Wyoming and western South Dakota suggests four significant correlations. Most workers consider the Morrison in the Powder River Basin to be equivalent to the Morrison in the Bighorn Basin (see Previous Work). The second correlation involves a series of dark chert bearing fluvial channel systems in the lower to middle Lakota and Cloverly formations. Dark chert bearing conglomeratic sequences occur at this stratigraphic position through out the early Sevier foreland basin, but are enigmatic as they do not appear to represent a continuous regional sheet

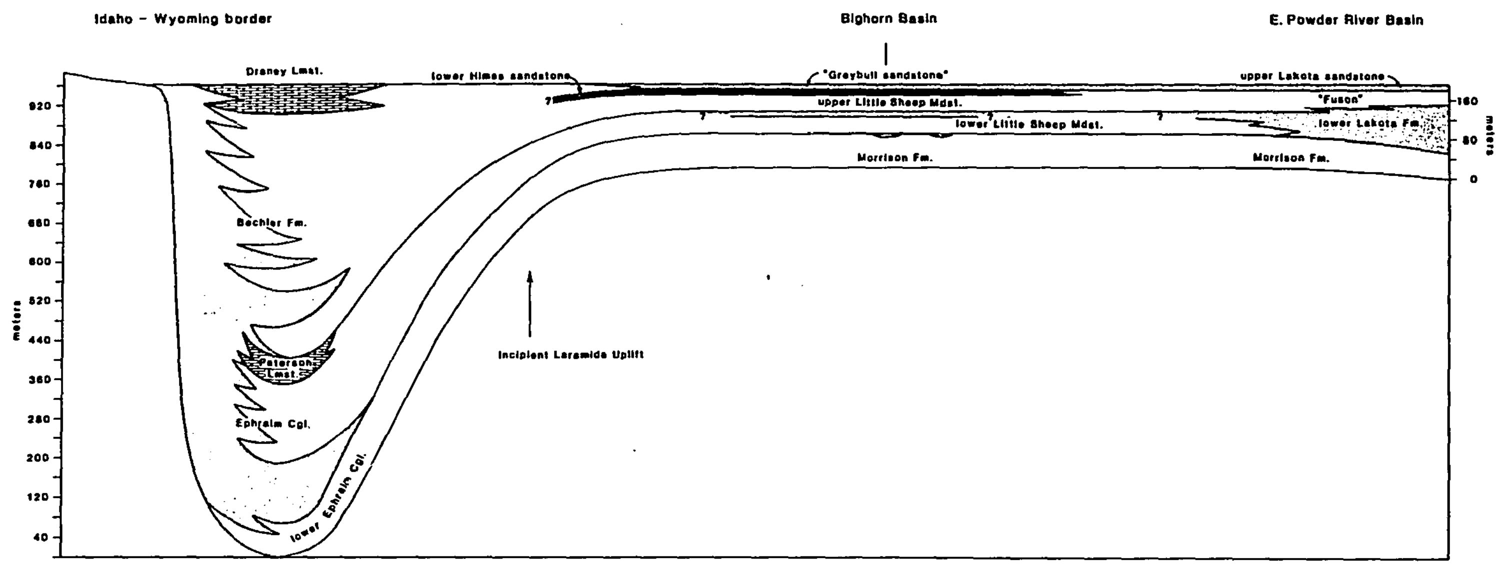
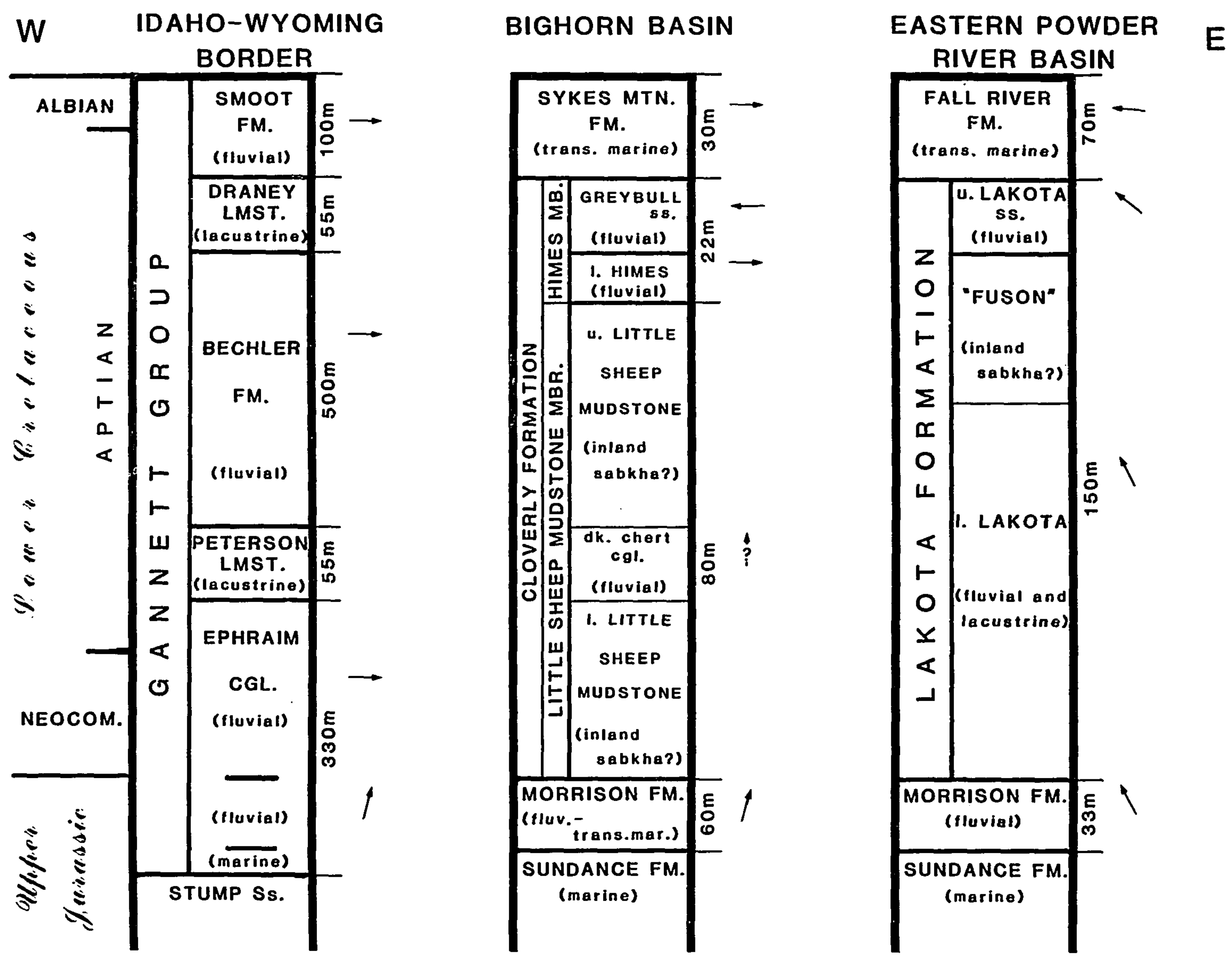


Figure 25. Generalized cross section of upper Jurassic / lower Cretaceous nonmarine sequence across northern Wyoming

Table 4. Stratigraphic correlation chart of northern Wyoming


W
IDAHO-WYOMING BORDER
BIGHORN BASIN
EASTERN POWDER RIVER BASIN
E
ALBIAN
APTIAN
NEOCOM.
Lower Cretaceous
Upper Jurassic
GANNETT GROUP
SMOOT FM. (fluvial)
100m
DRANEY LMST. (lacustrine)
55m
BECHLER FM. (fluvial)
500m
PETERSON LMST. (lacustrine)
55m
EPHRAIM CGL. (fluvial)
(fluvial)
(marine)
330m
STUMP Ss.
SYKES MTN. FM. (trans. marine)
30m
CLOVERLY FORMATION
HIMES MB.
GREYBULL ss. (fluvial)
l. HIMES (fluvial)
22m
LITTLE SHEEP MUDSTONE MBR.
u. LITTLE SHEEP MUDSTONE (inland sabkha?)
dk. chert cgl. (fluvial)
l. LITTLE SHEEP MUDSTONE (inland sabkha?)
80m
?
MORRISON FM. (fluv.-trans.mar.)
60m
SUNDANCE FM. (marine)
FALL RIVER FM. (trans. marine)
70m
LAKOTA FORMATION
u. LAKOTA ss. (fluvial)
"FUSON" (inland sabkha?)
l. LAKOTA (fluvial and lacustrine)
150m
MORRISON FM. (fluvial)
33m
SUNDANCE FM. (marine)

deposit. Rather it appears, on the basis of paleocurrent data, that they have multiple source areas perhaps reflecting syntectonic partitioning of the foreland basin (discussed below).

A third lithologic correlation can be made between the varicolored claystones and siltstones which occur in the upper quarter of Bolyard and McGregor's (1966) Lakota Formation in southwestern South Dakota (Fuson Shale of Darton, 1901) and the varicolored mudstone sequence of the Little Sheep Mudstone Member in the Bighorn Basin. This mudstone interfingers with the underlying Cloverly dark chert bearing conglomerates locally within the Bighorn Basin. A fourth correlation is possible between the westward directed sand dominated type I Greybull channel sandstones (Kvale and Vondra, 1985a) in the Bighorn Basin and the thick sand dominated westward directed channel sandstones of the uppermost Lakota in the Powder River Basin which are incised into the upper Lakota varicolored mudstones. Gott and others (1974) report that these sandstone complexes have downcut into the Lakota by as much as 50m and form resistant vertical cliff face exposures. The sandstones are quartz arenites and are characterized by mid-channel, large-scale planar foresets, some of them overturned and separated by thin 1cm thick sets very similar to the thick Greybull channels of the Bighorn Basin (Kvale and Vondra, 1985b).

Certain correlations can also be extended to the west from the Bighorn Basin. Much of the Gannett Group of western Wyoming and southeastern Idaho has been considered by many workers to be correlative with the Morrison and Cloverly Formations in Wyoming and with the Morrison and Kootenai Formations in Montana (Eyer, 1969; Furer, 1970; Holm et al., 1977). On the basis of lithologic similarity, stratigraphic position, marine molluscan fauna and dinosaur remains, the lower part of the Ephraim Conglomerate is considered to be Late Jurassic in age and equivalent to the Morrison and uppermost Sundance formations to the east (Eyer, 1969; Furer, 1970; Holm et al., 1977; Wanless et al., 1955). Microfaunal evidence indicates that the upper Ephraim is at least in part Aptian (Eyer, 1969). The Peterson and Draney Limestones are also Aptian, and the Smoot Formation spans the Aptian/Albian boundary (Eyer, 1969). Eyer (1969) and Furer (1970) have equated the sequence encompassing the upper part of the Ephraim Conglomerate through the Draney Limestone with the Cloverly and Kootenai of Wyoming and Montana. The Smoot Formation of western Wyoming, although considered nonmarine by Eyer (1969) and Durkee (1980), is considered the time stratigraphic equivalent of the transitional marine Sykes Mountain Formation in the Bighorn Basin (Holm et al., 1977).

The two major clastic sequences in the Gannett Group are the Ephraim Conglomerate and the Bechler Formation. Dark chert clast conglomerates occur within both of these units but are not convincingly traced beyond the early Sevier foredeep region of the fold-thrust belt. Both formations have been interpreted to indicate pulses of uplift along the Paris Thrust (Furer, 1970; Jordan, 1981; Wiltschko and Dorr, 1983). The Peterson and Draney Limestones were suggested to have been deposited during periods of tectonic quiescence (Furer, 1970; Jordan, 1981; Wiltschko and Dorr, 1983). Holm and others (1977), however, have interpreted these mostly freshwater carbonates to be indicative of the repeated uplift of Paleozoic carbonates and their dissolution and subsequent redeposition. Recent comparisons of the early Sevier foredeep sedimentary facies with Late Cretaceous-Early Paleogene tectonic controlled sedimentary sequences in Laramide basins, discussed below, also suggests that the Peterson and Draney limestones are syntectonic deposits (Kvale and Beck, 1985).

TECTONIC CONTROL ON EARLY CRETACEOUS SEDIMENTARY FACIES

Partitioning of Early Cretaceous Foreland Basin

Recent field studies of the Lower Cretaceous nonmarine sequence indicate that incipient Laramide structures and provinces in southwest Montana and northern Wyoming segmented the early Sevier foreland basin and controlled or influenced sedimentation patterns and periodically may have been important intraforeland basin source areas (DeCelles, 1984; Kvale and Beck, 1985; Schwartz, 1982; Suttner et al., 1981). Evidence of partitioning within northern Wyoming and southern Montana is suggested by the enigmatic dark chert bearing conglomerates which occur at the base of the Lakota and Cloverly formations in the Powder River and Bighorn basins. These deposits are not convincingly traceable to similar dark chert-bearing horizons (Ephraim Conglomerate) within the fold-thrust belt of western Wyoming. Rather the Ephraim appears to be restricted to just the fold-thrust belt region generally west of the Prospect Thrust (Furer, 1970; Sippel, 1982). The regional paleodrainage is not clear at this time, but paleocurrent data from conglomeratic units in the Bighorn and Powder River basins (Figure 26) (Gott et al., 1974; MacKenzie and Ryan, 1962; Mirsky, 1962; Moberly, 1960) suggests that channels drained multiple source areas outside and within the foreland basin principally to the south, west and east and possibly

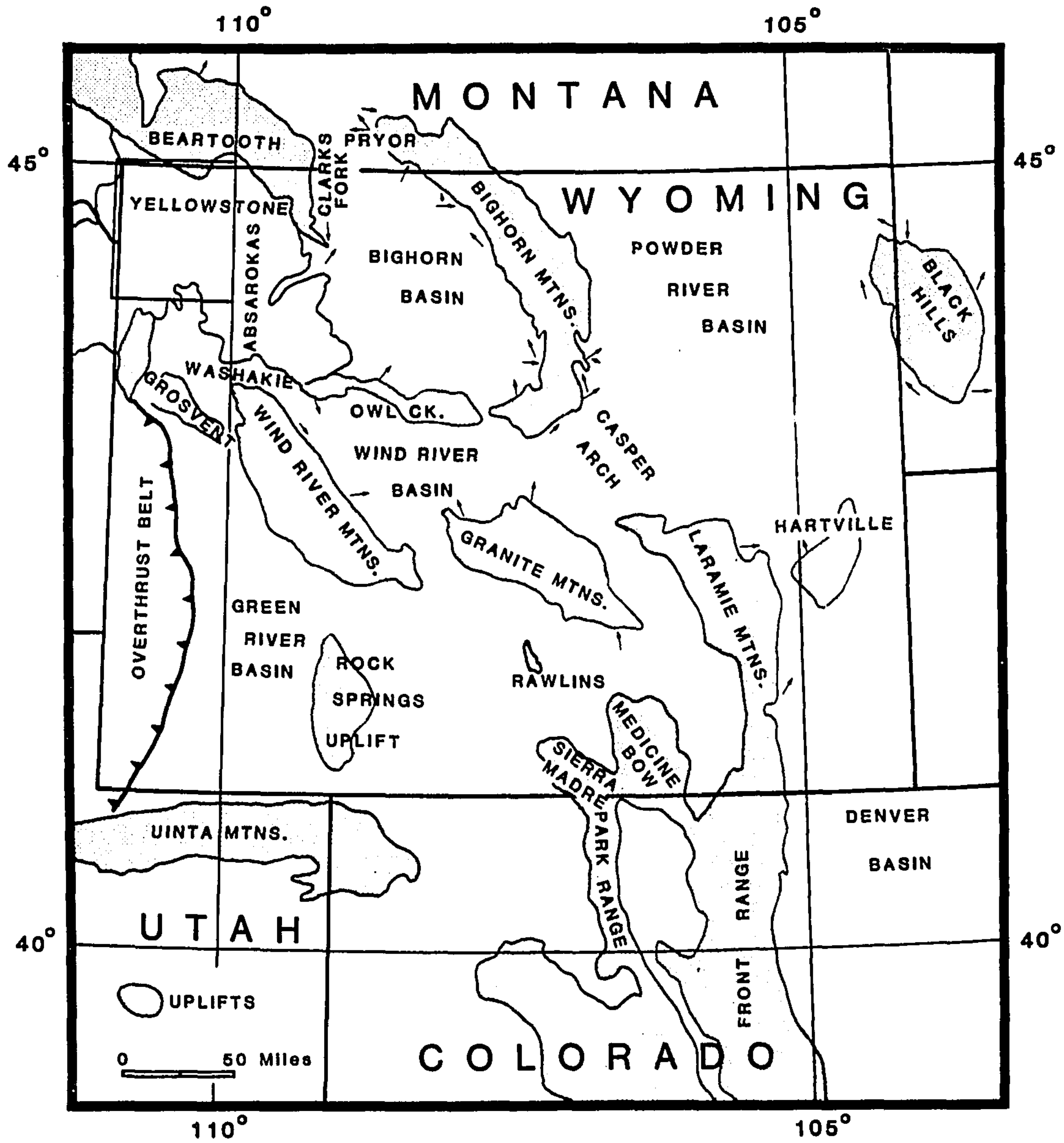


Figure 26. Generalized paleocurrent trends of Cloverly dark chert conglomerates; data compiled from this study, MacKenzie and Ryan (1962), Mirsky (1962), and Moberly (1960)

coalesced to form a trunk drainage system which flowed northward out of the Bighorn Basin, perhaps connecting with the large northward directed Cutbank channel system of the lower Kootenai described near Great Falls, Montana by Hopkins (1985). One intrabasin source area was the Beartooth uplift. DeCelles (1984) has shown that during the earliest Cretaceous the Beartooths were a positive feature that was stripped of lower Kootenai conglomerates which were then transported eastward and redeposited as a basal Cloverly dark chert clast conglomerate that Moberly (1960) termed the Pryor Conglomerate. More recent studies indicate that at least some of the conglomerates of the Pryor may have been derived from Permian aged carbonates exposed in the uplift (DeCelles, Dept. Geology, Univ. Indiana, Bloomington, pers. comm., 1985). There is no evidence that the Bighorn Mountains were an uplift at this time, but other nonvolcanic source areas within the Early Cretaceous foreland basin seem probable, but unidentifiable, until additional work is done.

Intrabasin volcanic activity was important during the Early Cretaceous in northern Wyoming. A volcanic highland supplied sediments to the Cloverly depositional basin during the Aptian. The sandstone unit of the lower Himes was deposited by a major volcaniclastic rich braided channel complex. The unit was found only in the northern third of the Bighorn Basin and paleocurrent data (Figure 27) suggests transport from the Yellowstone/northern Absaroka region. The channels have not been recognized east of the Bighorn Mountains although volcaniclastic horizons have been reported from the upper Lakota Formation (Gott et al., 1974; Waage, 1959b). The lower Himes contains abundant sand size fragments of volcanic rock and zoned plagioclase as well as andesite clasts over 20cm in diameter. Strati- graphic equivalents deposited within the foredeep in southwest Montana and eastern Idaho contain few of these components (Furer, 1970; Suttner et al., 1981). These data suggest that volcanic activity in the Yellowstone/ northern Absaroka region, dated to be only as old as Eocene in age (Smedes and Prostka, 1972), is at least 60 million years older than previously thought.

The first indication that structures today associated with the modern Bighorn Mountains controlled sedimentation during the Early Cretaceous is found within the Greybull interval. Facies and paleocurrent analyses have shown that the large, type I, sand dominated, straight channel systems of the Greybull interval were confined to regions north of what is now the area of maximum uplift in the modern Bighorn Mountains (Figure 28).

Along the west flank of the Bighorns, the Greybull channels are confined to areas delineated by the Tongue

Figure 27. Generalized paleocurrent trends of lower Himes channels; circles denote lower Himes deposits whose paleocurrent indicators were not measured

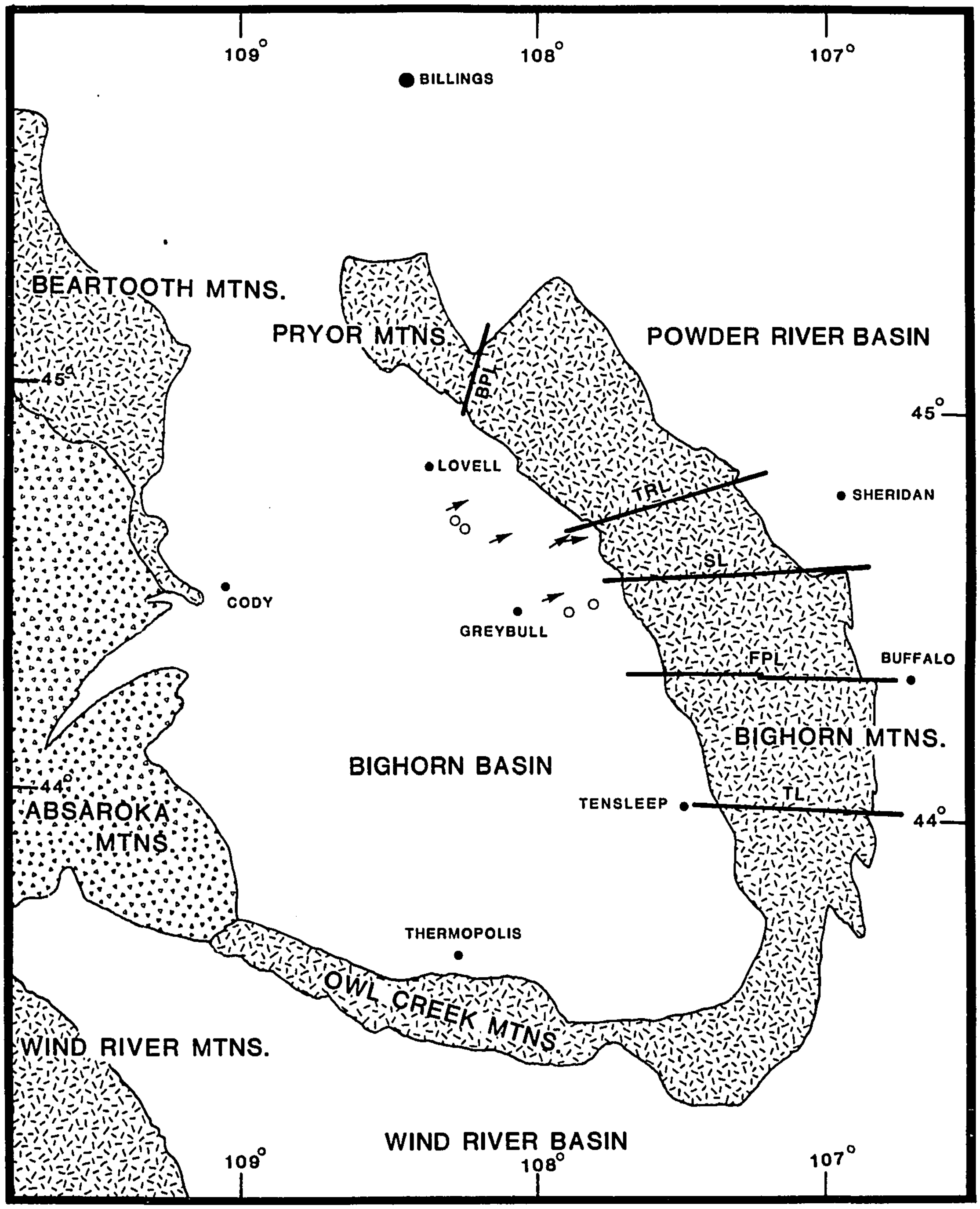
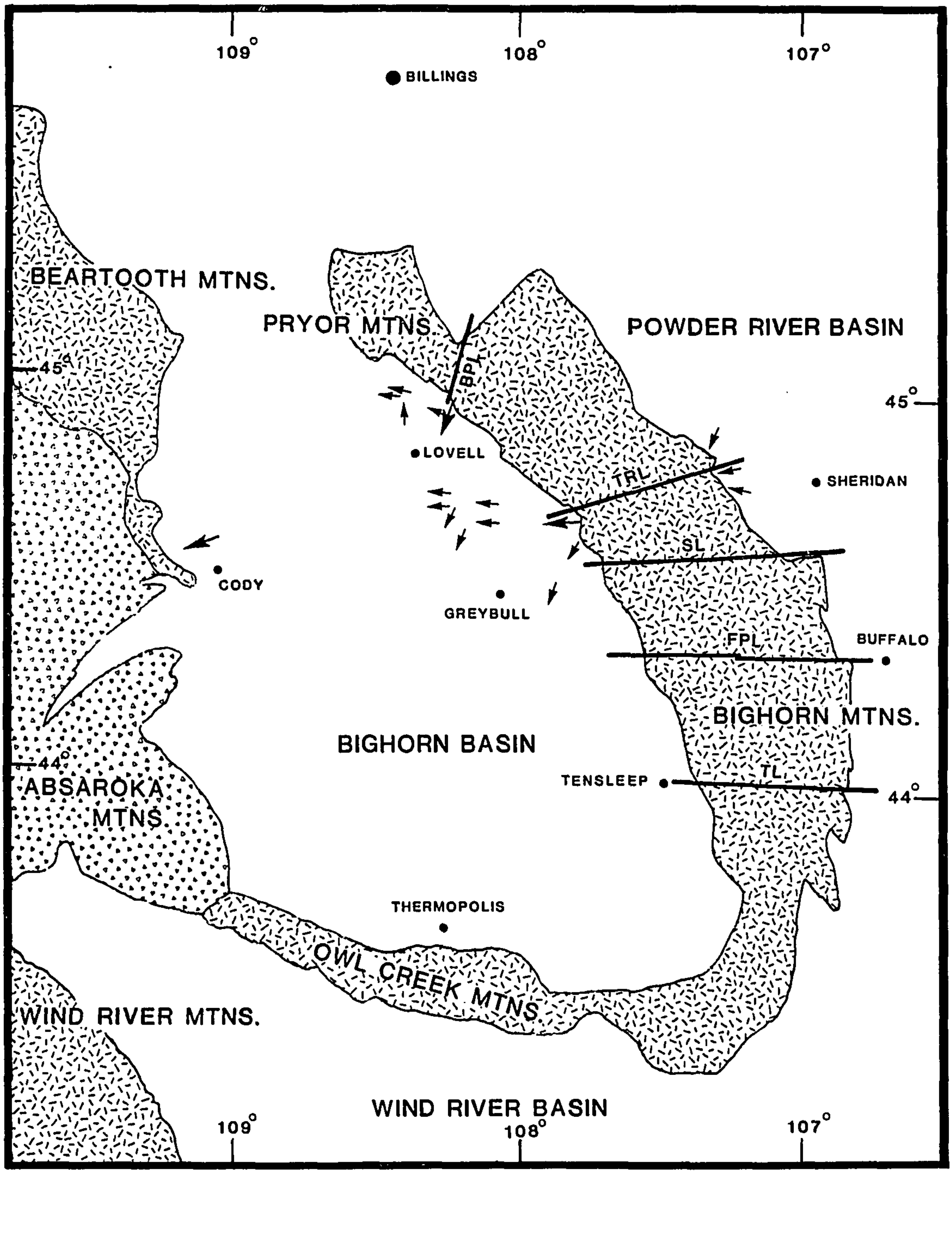

109°
108°
107°
BILLINGS
BEARTOOTH MTNS.
PRYOR MTNS.
BPL
POWDER RIVER BASIN
45°
45°
LOVELL
TRL
SHERIDAN
SL
CODY
GREYBULL
FPL
BUFFALO
BIGHORN MTNS.
BIGHORN BASIN
44°
ABSAROKA MTNS.
TENSLEEP
TL
44°
THERMOPOLIS
OWL CREEK MTNS.
WIND RIVER MTNS.
WIND RIVER BASIN
109°
108°
107°

Figure 28. Generalized paleocurrent trends of type I channels; paleocurrent trends in the Powder River Basin were established by David Huntsberger III, Dept. of Earth Sciences, Iowa State University


109°
108°
107°
BILLINGS
BEARTOOTH MTNS.
PRYOR MTNS.
POWDER RIVER BASIN
BPL
45°
LOVELL
TRL
SHERIDAN
SL
CODY
GREYBULL
FPL
BUFFALO
BIGHORN MTNS.
BIGHORN BASIN
44°
ABSAROKA MTNS.
TENSLEEP
TL
THERMOPOLIS
OWL CREEK MTNS.
WIND RIVER MTNS.
WIND RIVER BASIN

River lineament defined by Hoppin and Jennings (1971) and the structural break between the Bighorn Mountains and the Pryor uplift. In both cases, the Greybull channels form multilateral channel sandstone bodies at least 20m thick (Kvale and Vondra, 1985b). From these points the channels "splayed out", flowing west to southwest, and are preserved as mostly isolated channel bodies laterally separated from one another by several kilometers of relatively thin overbank material. It appears on the basis of paleocurrent data that all of the thick sand-dominated Greybull channels within the Bighorn Basin can be traced back to these two areas. Comparable channels do not occur elsewhere along the west flank of the Bighorns.

Reconaissance studies along the east flank of the Bighorns by D. V. Huntsberger III (Dept. of Earth Sciences, Iowa State University, pers. comm., 1985) indicates that large Greybull channels are likewise confined to areas north of the region of maximum uplift. Preliminary paleocurrent measurements of channels near the eastern reach of the Tongue River lineament reveals that flow was westward directed along the lineament. Paleocurrent data are not yet available for channels north of the Pryor uplift.

The occurrence of multilateral Greybull sandstone bodies in specific regions along the flanks of the Bighorn and Pryor mountains indicates that the channels were areally restricted in their avulsion patterns. The proximity of the outcrops to major lineations suggests that differential movement during the Early Cretaceous along the lineations produced a half graben-like setting, restricting the lateral migration of the channels. Farther out into the Bighorn Basin, structural controls were not as pronounced and the channels were free to migrate by avulsion laterally over great distances. It is also significant that in the two areas in the Bighorn Basin where the Greybull channels strongly coalesced, the thicknesses of the complete Cloverly sequence exceed those in the rest of the basin by 10 to 20%. In contrast, the Greybull interval exposed along the west flank of the Bighorn Mountains, immediately adjacent to the area of maximum uplift, lacks the thick channel facies and is characterized by thin overbank deposits and smaller channel development. This is likely indicative of a regional high over a broad arch whose elongation direction is suggested by the northeast trend of the Tongue River lineament. Additional work, however, is required along the east and south flank of the Bighorn Mountains to substantiate this trend.

The concept of pre-Laramide tectonics or structural trends controlling or influencing sedimentation patterns in northern Wyoming is poorly understood. Recent work has shown that pre-Laramide deformation within the eastern and central Powder River Basin has occurred intermittently during the Phanerozoic, influencing depositional patterns of the Permian Minnelusa Formation; the Lower Cretaceous Lakota (Cloverly) Formation, Fall River Sandstone and Muddy Formation and the Upper Cretaceous Turner, Sussex, Shannon and Parkman sandstones (Gott et al., 1974; Rasmussen and Bean, 1984; Slack, 1981; Weimer et al., 1982).

Much of the control on these units is said to be movement along subtle basement lineaments which reflect regional stresses developed during Precambrian times (Hoppin and Palmquist, 1965; Houston, 1971; Marrs and Raines, 1984; Rasmussen and Bean, 1984; Slack, 1981). Hoppin (1974) suggested that these lineaments are best observed in the Laramide uplifts immediately adjacent to the basins where they manifest themselves as faults, breccia zones and mineralized zones. However, Hoppin and Jennings (1971) point out that the problem of differentiating Laramide deformation along lineaments from earlier periods of deformation is very difficult even when basement rock is exposed. One approach to solving this problem is to look at pre-Laramide sedimentary sequences developed in basins adjacent to uplifts. If movement had taken place along a lineament (either horizontal or vertical) and was sustained over a period of time, then facies development such as multistorey channel sandstones or shifts in trends of channels or marine bars over a short distance would have resulted. Fluvial and transitional marine sediments are particularly susceptible because of their sensitivity to slight changes in base level or inhomogeneities of bedrock or alluvium.

Thrust Control on Early Cretaceous Sedimentary Facies Associations

Recent research on sedimentary facies associations in Laramide intermontane basins (Beck, 1985; Beck et al., 1986) of the Rocky Mountain foreland has revealed a consistent pattern of facies development which records episodes of displacement and quiescence on adjacent thrusts. Episodes of thick-skinned basement thrusting or thin-skinned thrusting coupled with tectonic stacking are correlated with: 1. tectonically induced subsidence adjacent to thrust tips, 2. the localization of low gradient depositional environments within each foredeep, 3. the formation of regional thrustward dipping paleoslopes, and 4. the restricted deposition of coarse clastics to thick narrow bands immediately adjacent to thrust tips.

However, periods of little or no thrust displacement are characterized by: 1. minimal tectonically induced subsidence, 2. expansion of low gradient depositional environments, 3. flattening of thrustward dipping paleoslopes, and 4. the advance of thin tongues of thrust-derived coarse clastics towards the distal margins of foreland basins.

Coarse clastics were derived from thrust tips during periods of displacement as well as during periods of relative or complete quiescence. During periods of thrust displacement, tectonically induced subsidence steadily moved the coarse erosional products of thrust tips beneath the local base levels of adjacent foredeeps preventing their transport towards the distal margins of foreland basins. Thick fan deposits of coarse clastics accumulated in narrow tracts adjacent to thrust tips. These consisted of vertically stacked sequences of conglomerates and conglomeratic sandstones which give way laterally to progressively finer grained deposits.

During periods of relative or complete quiescence of thrusted basin margins, rates of erosion of already uplifted thrust tips were probably similar to those near the end of episodes of rapid thrusting, however, rates of tectonically induced subsidence were very low. Coarse detritus eroded from thrust tips was no longer moved as quickly beneath the local base levels and prograded transversely across adjacent foredeeps. Only during times of relative or complete thrust quiescence would tongues of coarse thrust derived detritus reach far into foreland basins. These changes in the vertical and lateral distribution of coarse clastics were due to changes in local base level controlled by the rate and degree of tectonically induced subsidence. The sedimentary-tectonic model described above allows for the reconstruction of histories of thrust movement using the distribution of sedimentary facies associations occurring within adjacent foreland basins.

Comparison of sedimentary facies associations in Laramide intermontane basins with those of the classic Sevier Upper Jurassic/Lower Cretaceous nonmarine sequence of the Rocky Mountain Foreland yields a striking similarity of facies patterns (Beck et al., ca. 1987). Application of a sedimentary-tectonic model developed for the Laramide foreland basins to the Mesozoic nonmarine sequence allows a comprehensive event stratigraphic model to be developed for the northern Wyoming reach of the early Sevier foreland basin consistent with the accepted lithostratigraphic correlations and known ages of the units.

During early Aptian, rapid thrusting of the Paris Thrust resulted in the foredeep deposition of the Peterson Limestone within an extensive freshwater lake which paralleled the fold-thrust belt. Coarse clastic sediment derived from the fold-thrust belt was confined to the foredeep (excluding perhaps volcanic ash), west of the site of carbonate deposition. Sediment to the rest of the foreland was provided by streams draining intrabasin highlands such as an incipient Beartooth Uplift as well as perhaps the Sioux Uplift to the east. The origin of the incipient Laramide uplifts is not known but may be related to blind basement thrusting generated during times of regional compression or broad arching caused by differential movement along parallel basement involved lineaments as suggested by Slack (1981). The resulting deposits form the coarse chert-rich clastics found in the lower to middle part of the Lakota Formation in the Powder River Basin and the conglomeratic channels in the lower to middle Little Sheep Mudstone in the Bighorn Basin. The regional paleodrainage is not clear at this time, but preliminary paleocurrent data from the Little Sheep mudstone channels indicate that the drainage was to the north-northwest out of the Bighorn Basin suggesting that a subtle Washakie-Beartooth drainage divide separated the foredeep from the rest of the foreland basin. As movement of the Paris Thrust slowed, deposition of gypsiferous, siliceous and calcareous muds and volcanic ash of the upper Little Sheep Mudstone continued in ephemeral lacustrine and related environments. During the mid-Aptian, the Bechler Formation was deposited adjacent to the fold-thrust belt in eastern Idaho and western Wyoming while the feldspathic to lithicwacke channel sequence of the lower Himes was deposited in northcentral Wyoming. Preliminary paleocurrent and petrographic data indicate that the lower

Himes channel systems drained a volcanic source area in the Yellowstone/northern Absaroka region. Deposition of predominantly fine-grained mudstones of the upper Lakota (Fuson of Darton, 1901) continued in eastern Wyoming during this time. Rapid movement of the Paris Thrust in late Aptian to early Albian caused a regional west dipping gradient to form cratonward of the foredeep. This resulted in an easterly derived fluvial channel system which extended into northcentral Wyoming. The resulting deposits constitute the mostly fluvial Greybull interval of the Himes Member in the Bighorn Basin and the uppermost Lakota fluvial sandstones of Gott and others (1974) in southwestern South Dakota. The channels extend at least as far as the northwestern margin of the present Bighorn Basin where the Greybull interval has been recognized in outcrops near the town of Cody. Coincident with the development of this channel system was the development of another extensive freshwater lake within the foredeep, adjacent to the fold-thrust belt in which the Draney Limestone accumulated. By Aptian-Albian times, the Early Cretaceous seaway advanced into northcentral Wyoming forming the transitional environments in which the Sykes Mountain Formation and perhaps the uppermost part of the Greybull interval were deposited while fluvial conditions persisted along the western highlands as recorded by the Smoot Formation.

A summation of depositional events within the early Sevier foreland basin of northern Wyoming shows that extensive laustrine conditions existed within the foredeep (Peterson and Draney limestones) with well-developed westward directed fluvial systems entering the basin from the craton during Aptian-Albian times coincident with rapid movement on the Paris Thrust. Preliminary paleocurrent data suggest that incipient Larmide uplifts (Washakie and Beartooth) partitioned the foreland basin in northern Wyoming. The uplifts were probably very subtle and may have been low relief surface expressions of blind basement thrusts or differential movement along basement cored lineaments activated during major compressive events to the west. These uplifts may have been source areas of some of the conglomeratic sequences deposited within the foreland basin. During periods of quiescence or slower thrust movement, deposition of fine-grained sediments within low gradient ephemeral lacustrine environments exemplified by the variegated mudstones of the Little Sheep Mudstone in the Bighorn Basin and the upper Lakota (Fuson of Darton, 1901) within the Powder River Basin dominated sedimentation within the central foreland basin with westward flowing fluvial systems dominating only the easternmost edge of the foreland basin.

SUMMARY

Field and laboratory studies of the Upper Jurassic Morrison and Lower Cretaceous Cloverly formations in the Bighorn Basin have resulted in the following conclusions.

1. Lower Morrison sediments in the northern part of the Bighorn Basin are transitional with upper Sundance tidal deposits. They were deposited in a lagoonal and coastal tidal flat or lacustrine environments.

2. The upper part of the Morrison Formation is a coastal plain fluvial deposit dominated by mixed-layered montmorillonite/illite - rich, red banded mudstones. The mudstones were deposited as overbank material from flooding of flashy ephemeral, high sinuosity, fluvial channels. The dominance of mudstones over sandstones suggests that sheet wash and eolian processes were also important in transporting and depositing the mud sized material. The presence of caliches, flashy ephemeral streams and red beds in the upper Morrison sediments suggest a somewhat arid climate, but the abundant fossil reptilian remains present suggests that the aridity was seasonal.

3. The Cloverly Little Sheep Mudstone Member and lower part of the Himes Member were deposited in an extensive, montmorillonite-rich, clay playa complex. Biologically and fluvially reworked, lenticular, devitrified tuffs are present within the Little Sheep Mudstone. Extensive

interfingering of high and low sinuosity, flashy ephemeral stream sandstones and dark chert bearing, low sinuosity, perennial stream deposits occur throughout the sequence. The lower Himes channels are volcaniclastic-rich, perennial, low sinuosity stream deposits which overlie, and locally interfinger with, the upper playa claystones of the Little Sheep Mudstone. This channel system drained volcanic highlands in the Yellowstone/northern Absaroka region which may have also supplied volcanic ash to the Little Sheep Mudstone playas. Other possible source areas for Little Sheep Mudstone sediments include the Sevier fold-thrust belt, intrabasin uplifts such as an incipient Beartooth uplift, arc volcanic vents in Idaho, and the craton.

4. The upper part of the Cloverly Himes Member (informally referred to as the Greybull interval) is characterized by three distinct channel body types representing deposition in: 1) low sinuosity upper reach of an estuary (type I), 2) high sinuosity tidal creek (type II), and 3) high sinuosity fluvial channel (type III). The channel bodies are encased in mudstones, some of which are believed to have been deposited on a tidal mudflat/marsh. The clay fraction of the mudstones is dominated by mixed layered montmorillonite-illite and kaolinite. Flow in the upper estuaries was ebb dominant and, like the fluvial channels, in a general east to west direction. The estuary

complex formed as a result of the Early Cretaceous marine transgression. Storm reworked offshore bars and flood dominant tidal channels are preserved in the shallow shelf deposits of the Sykes Mountain Formation. The upper Himes rests disconformably on the lower Himes and Little Sheep Mudstone.

5. The Early Cretaceous foreland basin sedimentation patterns of northern Wyoming were controlled by movement along the Paris thrust in eastern Idaho and influenced by incipient Laramide structures within the foreland basin. Intrabasin source areas during the Early Cretaceous include a volcanic province in the Yellowstone/northern Absaroka region and an incipient Beartooth uplift.

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APPENDIX A

Locations and Paleocurrent Measurements of Type I upper Himes (Greybull interval) channels

n = 50
T58N R96W
S1/3 S34

n = 100
T58N R96W
W1/2 SW1/4 S34

n = 70
T58N R96W
SE1/4 S29

n = 34
T58N R96W
E1/2 SW1/4 S22

n = 16
T58N R96W
south of center S27

n = 29
T58N R96W
C S34

n = 23
T57N R94W
C1/3 E1/2 S19

n = 50
T57N R94W
W1/2 NW1/4 S19

n = 47
T57N R95W
N1/2 SE1/4 S13

n = 40
T54N R93W
N1/2 SW1/4 S4

n = 35
T54N R94W
C NE1/4 S29

n = 17
T58N R96W
NE1/4 SW1/4 S34

n = 10
T58N R96W
NE1/4 NW1/4 S33

n = 50
T58N R96W
NE1/4 S19

n = 25
T58N R96W
W1/2 NE1/4 S27

n = 15
T58N R96W
C S34

n = 13
T58N R96W
north of center S34

n = 77
T57N R94W
C1/3 S1/2 N1/2 S19

n = 53
T57N R95W
S1/2 SE1/4 S13

n = 50
T57N R95W
NE1/2 NE1/4 S14

n = 65
T54N R93W
intersec. of sections 4,5,8,9

n = 75
T54N R94W
SE1/2 S12

n = 75
T54N R95W
NE1/4 S1

n = 25
T55N R94W
C N1/3 S25

n = 50
T55N R95W
W1/2 NE1/4 S34

n = 60
T55N R95W
N1/2 S1/2 S15

n = 44
T52N R92W
SW1/4 S11

n = 100
T54N R91W
W1/2 S33

n = 59
T54N R91W
NE1/4 S6

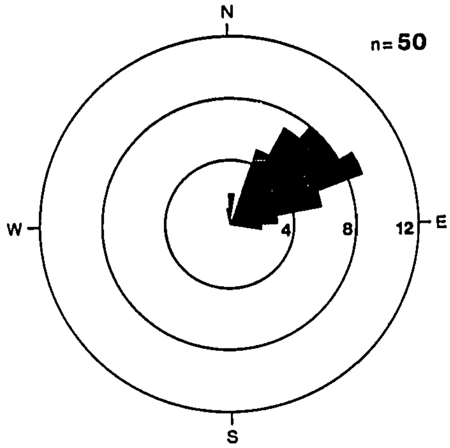
n = 49
T55N R95W
S1/2 NE1/4 S34

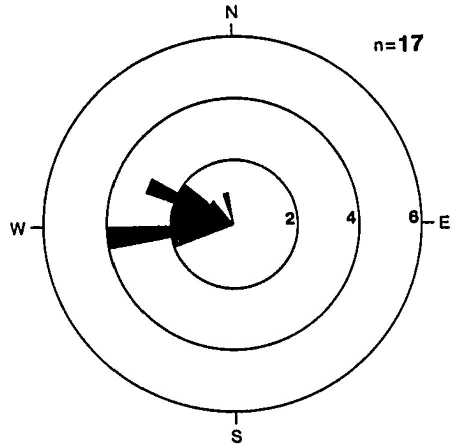
n = 30
T55N R95W
W1/2 SE1/4 S27

n = 70
T55N R95W
E1/2 W1/2 S15

n = 29
T52N R92W
intersec. of NW1/4 S14 and
SW1/4 S11

n = 135
T54N R91W
C1/3 S8


N
n=50
W
4
8
12
E
S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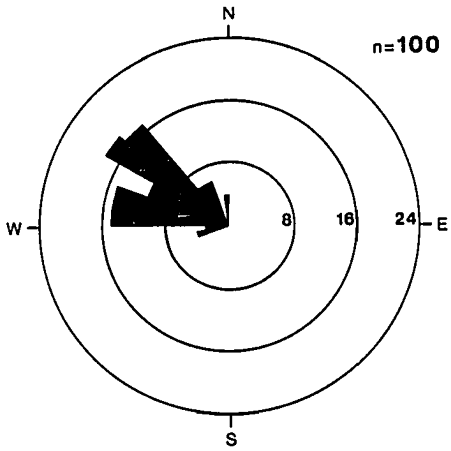
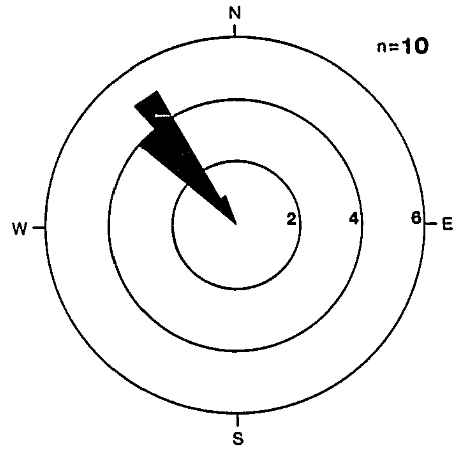
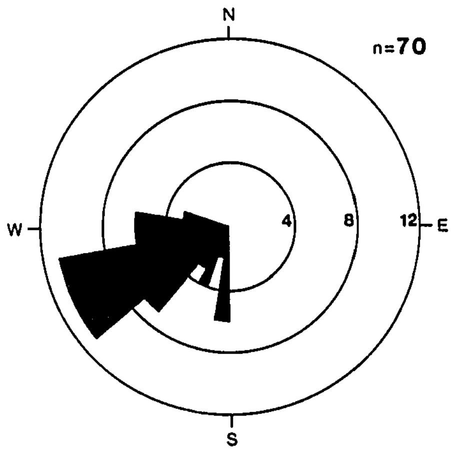
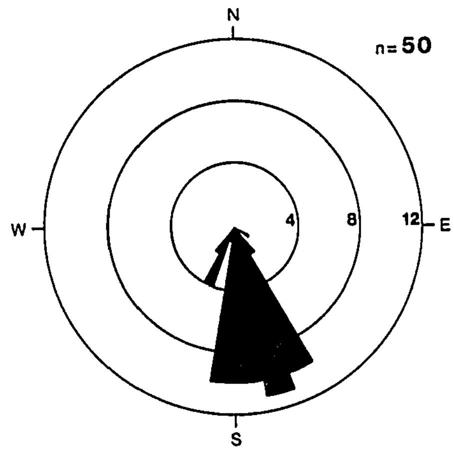
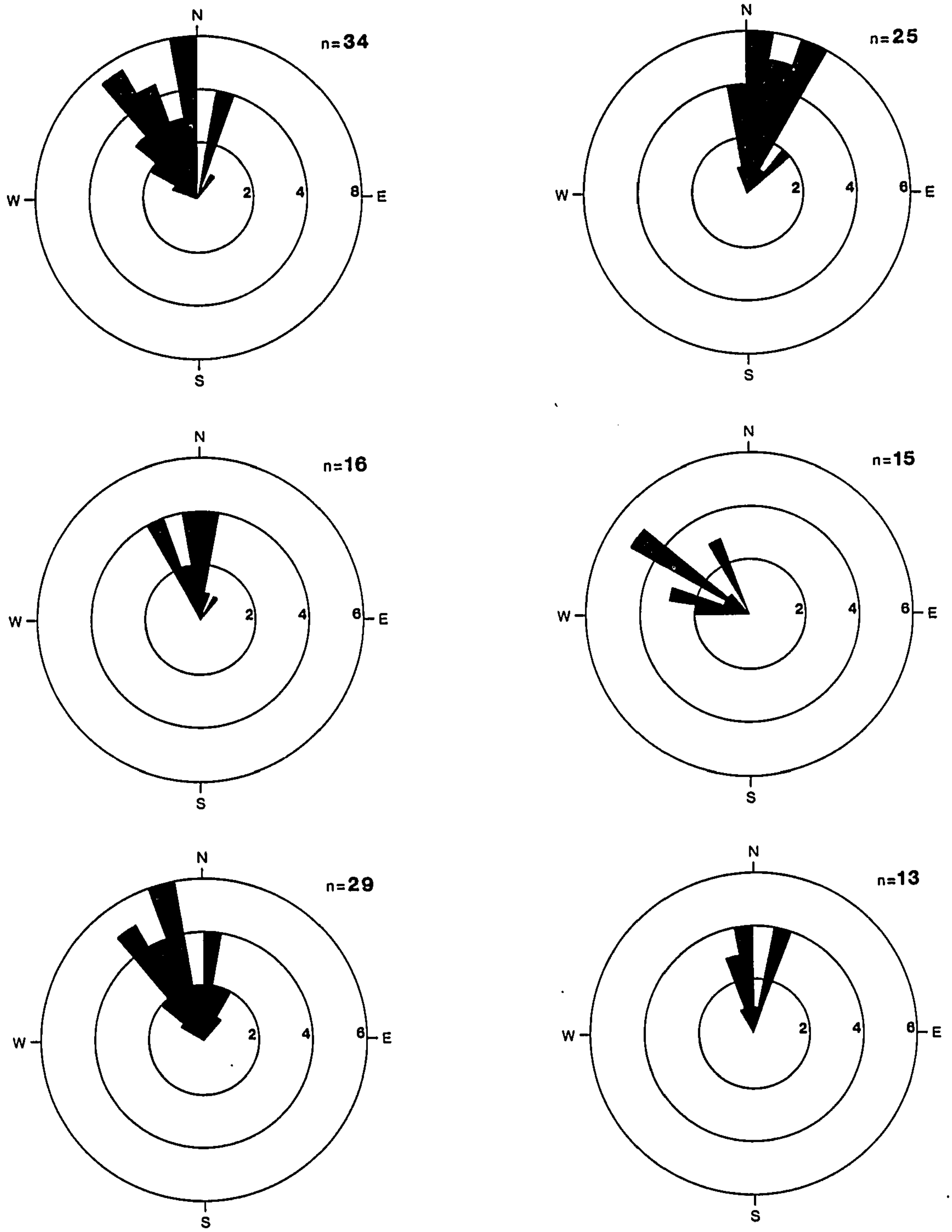
n=17
W
2
4
6
E
S

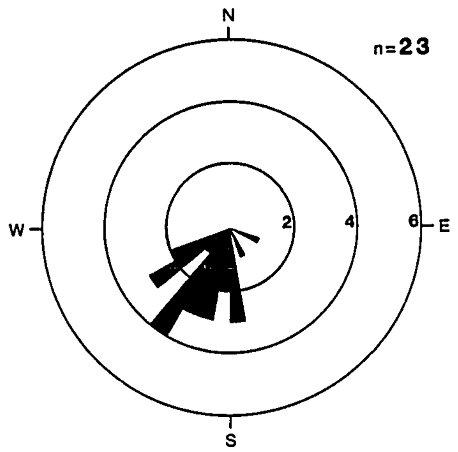

N
n=100
W
8
16
24
E
S

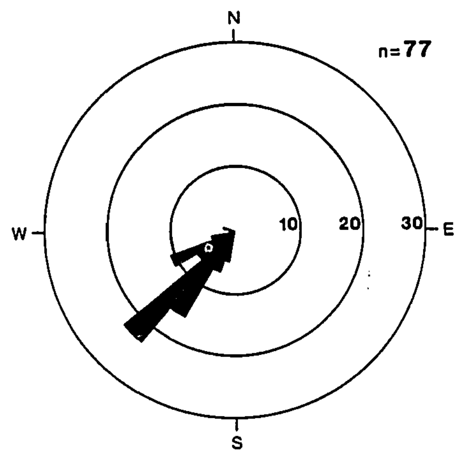

N
n=10
W
2
4
6
E
S

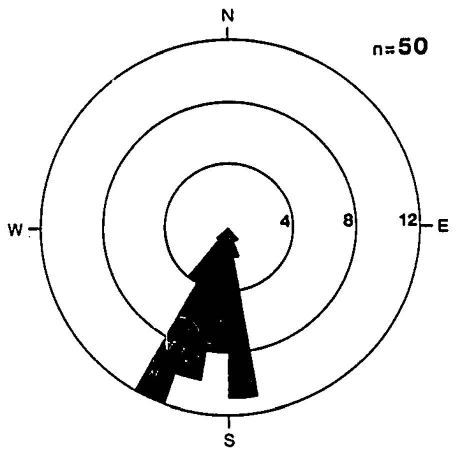

N
n=70
W
4
8
12
E
S

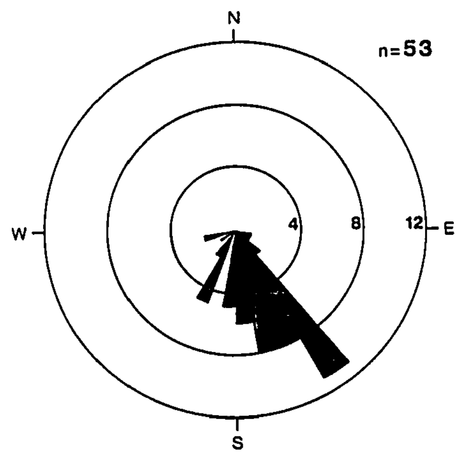

N
n=50
W
4
8
12
E
S

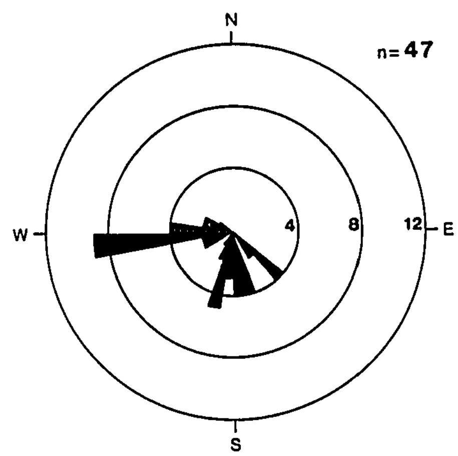

N
n=34
W
2
4
8
E
S
N
n=25
W
2
4
6
E
S
N
n=16
W
2
4
6
E
S
N
n=15
W
2
4
6
E
S
N
n=29
W
2
4
6
E
S
N
n=13
W
2
4
6
E
S

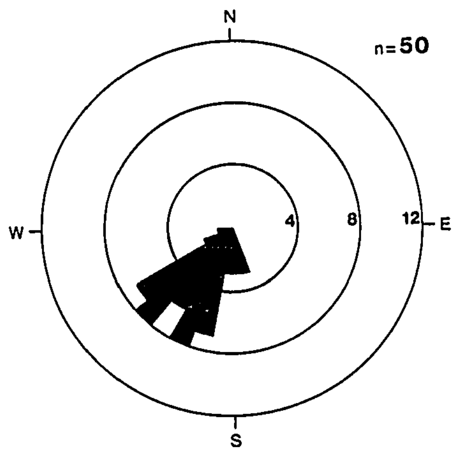

N
n=23
W
2
4
6
E
S

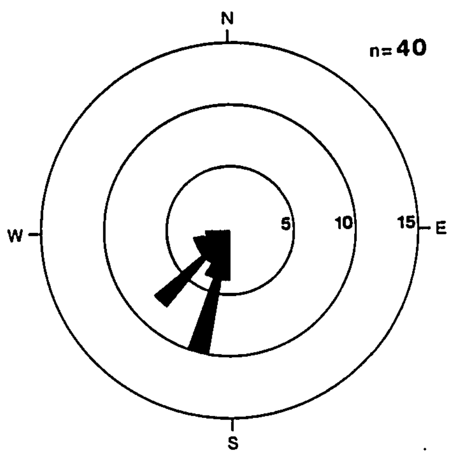

N
n=77
W
10
20
30
E
S

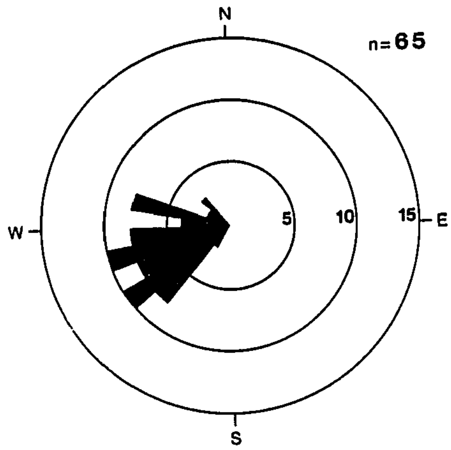

N
n=50
W
4
8
12
E
S

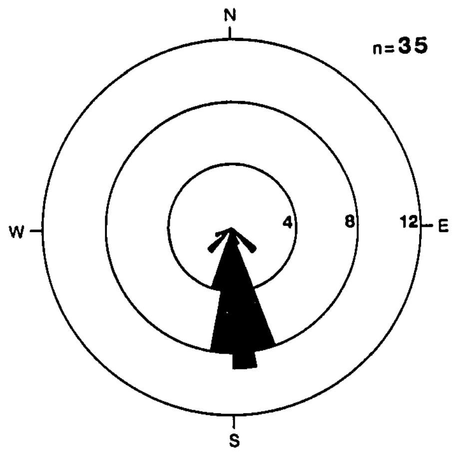

N
n=53
W
4
8
12
E
S

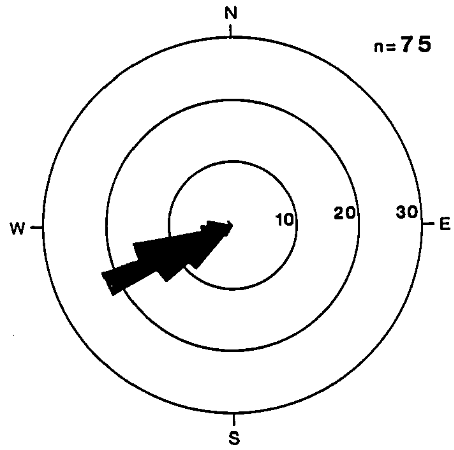

N
n=47
W
4
8
12
E
S

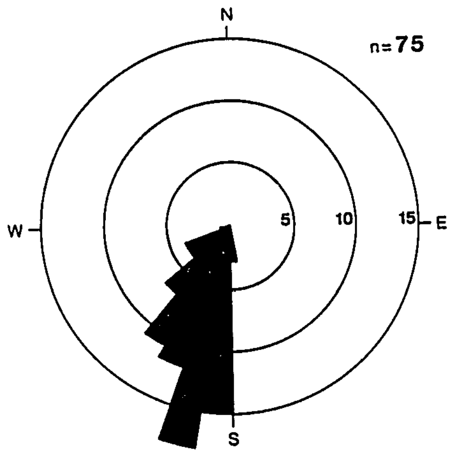

N
n=50
W
4
8
12
E
S

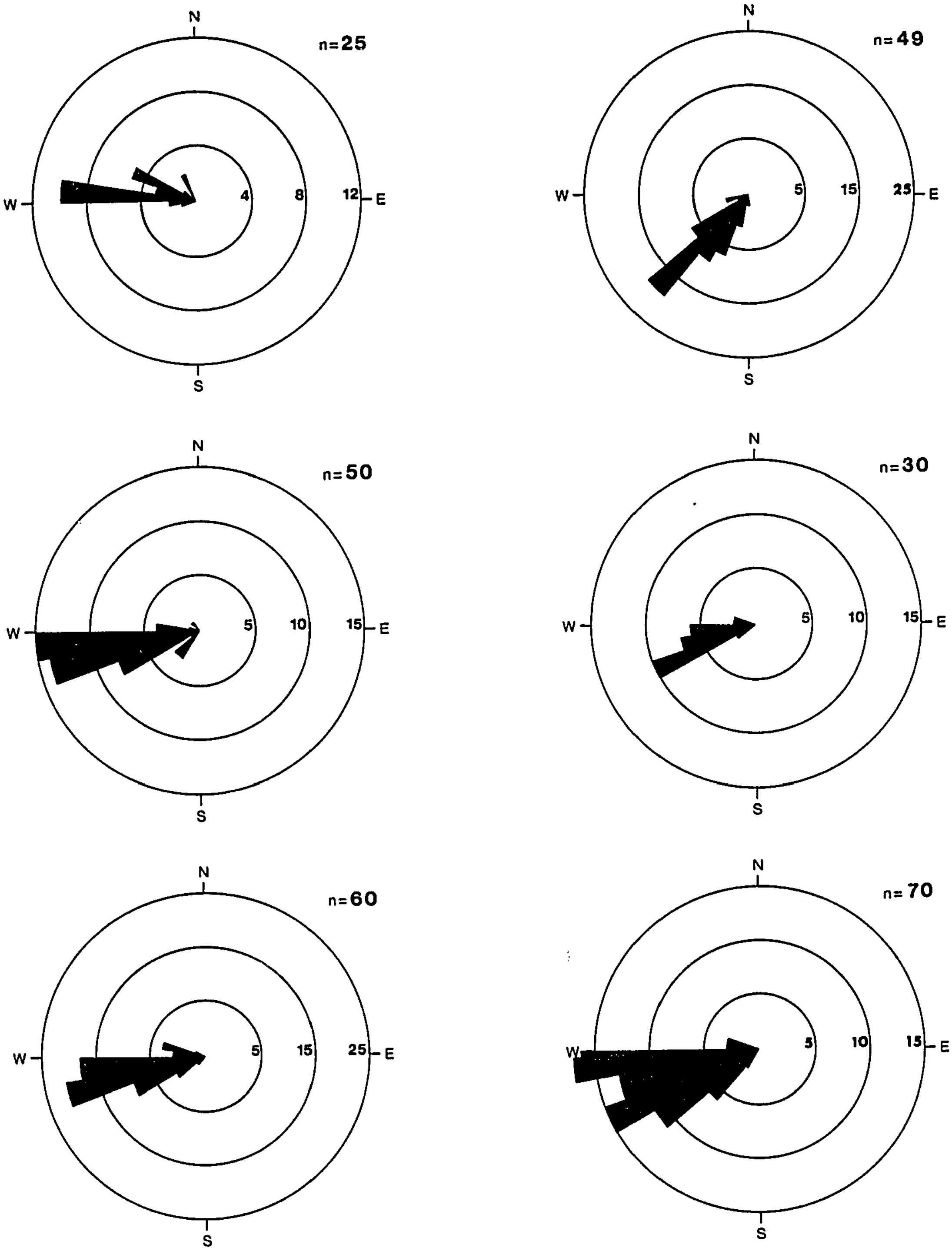

N
n=40
W
5
10
15
E
S


N
n=65
W
5
10
15
E
S

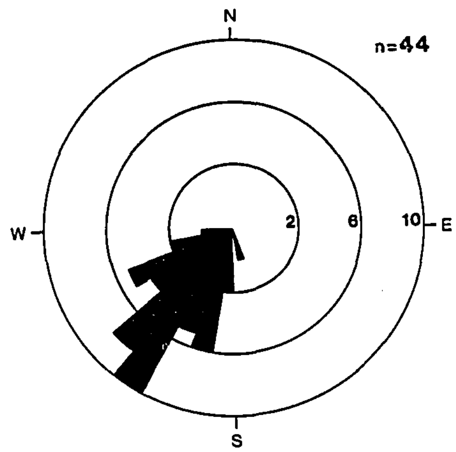

N
n=35
W
4
8
12
E
S

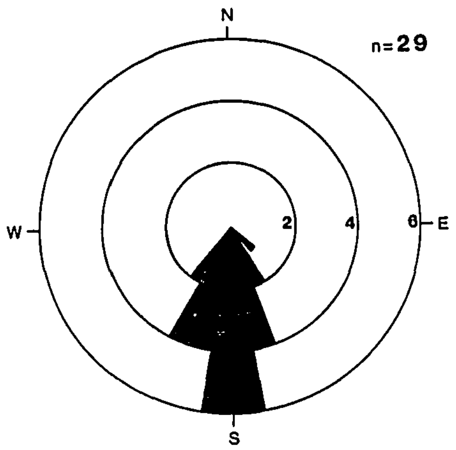

N
n=75
W
10
20
30
E
S

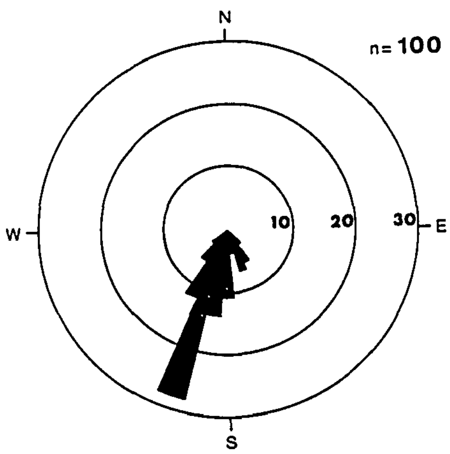

N
n=75
W
5
10
15
E
S

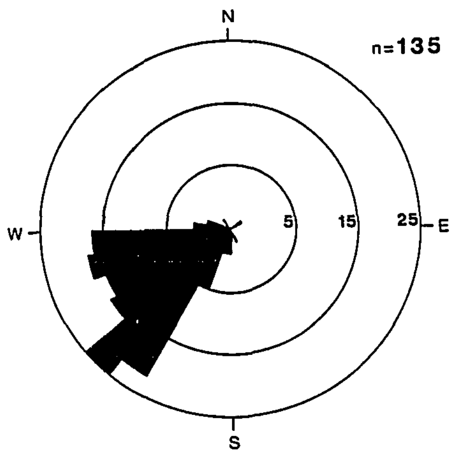


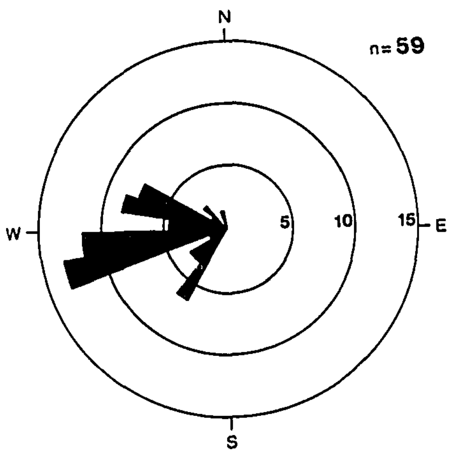
N
n=25
W
4
8
12
E
S
N
n=49
W
5
15
25
E
S
N
n=50
W
5
10
15
E
S
N
n=30
W
5
10
15
E
S
N
n=60
W
5
15
25
E
S
N
n=70
W
5
10
15
E
S


N
n=44
W
2
6
10
E
S


N
n=29
W
2
4
6
E
S


N
n= 100
W
10
20
30
E
S


N
n=135
W
5
15
25
E
S


N
n=59
W
5
10
15
E
S

APPENDIX B

Locations of some Type II and Type III Upper Himes (Greybull Interval) Channels

Type II	Type III
T58N R96W SW1/4 S34	T55N R95W center of line separating S15 and S22
T55N R95W NW1/4 SE1/4 S27	T54N R94W NW1/4 NW1/4 S6
T54N R94W S17	T54N R94W SE1/4 NW1/4 S6
T54N R94W E1/3 E1/2 S29	T53N R93W S1/2 S1/2 S35
T55N R94W S1/2 NE1/4 S25	
T54N R94W N1/4 S13 and S1/4 S12	

APPENDIX C

Qualitative mineralogy of clay minerals determined by X-ray diffraction

Clay mineral analyses were conducted on 39 Sundance, Morrison, Cloverly and Sykes Mountain samples. The samples were prepared and analysed under the guidance of Dr. Howard J. White at the Phillips Exploration and Production Lab in Bartlesville, Oklahoma. The samples were first disaggregated by gentle grinding and passed through a 100 mesh sieve. The fine fraction was then placed in a 250 ml bottle. H_2O and $NaCO_3$ were added and the bottle was then placed in a sonic mixer for an hour. Prior to mixing, selective samples were treated with 1/2 molar solution of HCl to remove carbonate cement. The bottles were then centrifuged at 350 rpm for 3 minutes and 48 seconds to force the settling of the greater than 5 micrometers (approx. 7.5 phi) size fraction. The suspended fine fraction was then decanted and used to make oriented clay mounts on 0.65 micrometer filter paper using a millipore filter suction system. The clay mounts were transfered to glass slides while still damp. The samples were allowed to dry and then X-rayed. The glass slides were then placed in an ethylene glycol chamber at 60° C for overnight then X-rayed again. The samples were finally heated to 550° C for 1/2 hour and X-rayed again. The results are summarized below. Clay minerals present include montmorillonite (m), illite (il), kaolinite (ka), and mixed layered montmorillonite / illite (mo/il). Peak intensities are described by the following abbreviations: very, very weak (vvw); very weak (vw); weak (w); moderate (m); strong (s); very strong (vs); and broad (br).

SAMPLE	AIR DRY d(A) I	GLYCOL. d(A) I	550° C d(A) I	INTP.	FM.
BHRI 4		16.0 w		mo	Jrs
	12.6 vs			mo/il	
			9.7 s	mo/il	
		8.7 w		mo (?)	
	6.2 w(+)			mo/il (?)	
		5.6 m		mo/il	
	5.1 vvw		4.8 s	il	
	4.5 mbr	4.5 m	4.4 m	mo/il	
		4.2 mbr		ka	
	3.6 vw			ka	
		3.4 s		il	
	3.3 vw			il	
	3.1 vs		3.2vs	mo	
BCAS 1c	13.6 sbr	14.5 mbr	13.2 w	mo	Jrs
	10.4 sbr	10.1 mbr	10.0 s	il	
	7.2 vw	7.2 vw		ka	
		5.6 wbr		mo	
	5.0 mbr	5.1 mbr	4.9 m	il	
	4.2 vw	4.3 wbr	4.2 vw	ka	
	3.6 mbr			ka	
	3.3 s	3.3 sbr	3.3 s	il	
BCAS 2c	13.4 sbr	14.2 mbr	12.6 wbr	mo/il	Jrm
	10.3 sbr	10.2 m	10.0 s	il	
	7.1 m	7.2 m		ka	
		5.6 vw		mo/il	
	4.9 mbr	5.0 m	4.9 m	il	
	4.2 w	4.3 wbr	4.2 vw	mo/il	
	3.6 mbr	3.6 m		ka	
	3.3 s	3.3 s	3.3 s	il	
BCAS2 4	10.4 vwbr			mo/il	Jrm
		10.0 vw	10.0 wbr	il	

SAMPLE	AIR DRY d(A) I	GLYCOL. d(A) I	$550^{o}C$ d(A) I	INTP.	FM.
	7.1 vvw	7.2 vw		ka	
	5.0 vw	5.0 vw	5.0 wbr	il	
	4.5 w		4.5 w	ka	
	4.2 w	4.2 wbr		ka	
			4.0 vw	ka	
	3.7 w	3.7 w	3.7 w	ka (?)	
	3.6 w			ka	
	3.3 m	3.3 m	3.3 m	il	
	3.2 w	3.2 vw	3.2 vwbr	il	
BCAS2 6		16.2 w		mo	Jrm
	12.8 mbr			mo/il	
	10.5 vw			mo/il	
			9.8 m	il	
		8.5 w		mo/il	
		7.1 vvw		ka	
		5.6 w(+)		mo/il	
	5.0 wbr		4.8 w(+)	il	
	4.5 w	4.4 wbr	4.5 wbr	mo	
	4.3 w	4.2 w	4.2 wbr	ka	
	3.6 vw		3.5 vw	ka	
	3.3 m	3.4 s	3.3 m	il	
	3.1 wbr		3.2 s	mo/il	
BCAS2 9	13.2 mbr	14.2 wbr		mo/il	Jrm
	10.4 wbr			mo/il	
		10.1 w	10.1 mbr	il	
	7.1 w	7.2 w		ka	
		5.5 w		mo/il	
	5.0 wbr	5.0 vw	4.9 wbr	il	
		4.3 vvw		ka	
	3.6 w	3.6 w		ka	
	3.3 w	3.3 m	3.3 mbr	il	
	3.0 w	3.0 vw	3.0 vw	mo/il	
BCAS2 11a	13.5 mbr	14.3 wbr	12.8 vwbr	mo/il	Jrm
	10.1 mbr	10.0 mbr	10.1 m	il	
	7.1 w(+)	7.1 w(+)		ka	
	5.0 w	5.0 w	5.0 w(+)	il	
	4.8 vw			il (?)	
	4.3 vw	4.3 vw	4.2 vw	ka	
	3.6 w(+)	3.6 w(+)		ka	
	3.3 s	3.3 s	3.3 s	il	
	3.2 vwbr		3.2 w	il	
BCAS2 11c		13.6 vvw		mo/il	Jrm
	11.9 wbr			mo/il	

SAMPLE	AIR DRY d(A) I	GLYCOL. d(A) I	550^{0} C d(A) I	INTP.	FM.
	10.3 mbr			mo/il	
		10.0 w	10.0 m	il	
	7.1 w	7.1 w		ka	
	5.0 wbr	5.0 w	5.0 wbr	il	
	4.5 w	4.5 w	4.5 w	mo	
	4.2 vw	4.2 w	4.2 w	ka	
	3.6 w	3.6 vw		ka	
	3.3 s	3.3 s	3.3 s	il	
	3.0 w	3.0 w	3.0 w	mo/il	
SCC 1	13.2 mbr	14.4 vwbr		mo/il	Jrm
	10.4 mbr	10.0 m	10.1 m	il	
	7.2 w	7.2 w		ka	
	5.0 w	5.0 wbr	5.0 wbr	ill	
	4.5 w	4.5 w	4.4 w	mo/il	
	4.2 w	4.3 w	4.2 w	mo/il	
	3.6 vw	3.6 vvw		ka	
			3.4 m	mo/il (?)	
	3.3 s	3.3 s	3.3 s	il	
	3.2 vvw	3.2 vw	3.2 wbr	mo/il	
SCC 2a			13.4 vvwbr	mo/il	Jrm
	12.4 wbr	13.4 vwbr	11.5 vvwbr	mo/il	
	10.5 mbr	10.2 mbr	10.3 mbr	il	
	7.1 vw	7.2 vw		ka	
	5.0 wbr	5.0 wbr	5.0 wbr	il	
	4.5 wbr	4.5 wbr	4.5 wbr	mo/il	
	4.3 wbr	4.2 wbr	4.2 vw	mo/il	
	3.6 vwbr			ka	
	3.3 s	3.3 s	3.3 s	il	
	3.2 vw			mo/il (?)	
SCC 3	13.6 vwbr			mo/il	Jrm
	10.0 wbr	10.1 w	10.1 m	il	
	7.1 vw	7.3 vw		ka	
	5.0 vw	5.0 vw	5.0 w	il	
	4.3 vvw	4.3 vw		ka	
	3.5 vvw	3.6 vvw		ka	
	3.3 m	3.4 m	3.3 m	il	
	3.2 vw	3.3 vvw		mo/il	
BCAM 1	13.6 wbr	14.5 vvw	12.6 vvw	mo/il	Jrm
	10.1 wbr	10.2 w	10.0 m	il	
	7.1 vw	7.1 vw		ka	
	5.0 vvw	5.0 vvw	4.9 vvw	il	
	3.3 m	3.3 m	3.3 m	il	

SAMPLE	AIR DRY d(A) I	GLYCOL. d(A) I	550° C d(A) I	INTP.	FM.
BCAM 1.1		16.0 vwbr		mo	Kcl
	12.1 mbr			mo/il	
		10.0 vwbr	10.1 m	il	
	7.2 w	7.3 w		ka	
	5.7 w	5.7 w		mo	
	5.0 vw	5.1 vw	5.0 m	il	
	4.5 vw	4.5 vw	4.5 w	mo (?)	
	4.3 w	4.3 w	4.3 w	ka	
	3.6 w(+)	3.6 w(+)		ka	
	3.3 m	3.3 s	3.3 s	il	
	3.1 mbr		3.2 sbr	mo	
	3.0 w	3.0 w		mo (?)	
BCAM 11	14.0 wbr	14.2 w		mo/il	Jrm
	10.4 wbr	10.0 w	9.9 mbr	il	
	7.0 w	7.1 w		ka	
	4.9 vw	4.9 vvw	4.9 vw	il	
	3.5 vvw			ka	
	3.3 m	3.3 m	3.3 m	il	
BCAM2 2a	13.2 mbr	13.3 mbr	12.6 wbr	mo/ill	Jrm
	10.4 mbr	10.1 mbr	9.8 s	il	
	7.2 m	7.1 m		ka	
		5.7 vw		mo/il	
	5.0 m	5.0 mbr	4.9 mbr	il	
	4.5 w	4.5 w	4.4 w	ka	
	4.3 wbr	4.3 w	4.2 w	ka	
WBI 1		12.6 wbr		mo/il	
	11.0 sbr			mo/il	
		10.1 mbr	10.2 s	il	
	5.0 mbr	5.0 wbr	5.0 mbr	il	
	4.5 w	4.5 wbr	4.5 wbr	ka	
	4.2 w	4.3 wbr	4.2 wbr	mo/il	
	3.3 sbr	3.3 s	3.3 sbr	il	
WBI 6		16.1 wbr		mo	Kcl
	12.6 wbr			mo/il	
			9.8 w(+)br	mo/il	
		8.5 vwbr		mo (?)	
		5.6 w		mo	
			4.8 w(+)	il	
	4.5 wbr	4.5 w(+)	4.5 m	mo	
	4.3 wbr	4.2 w(+)	4.3 m	il (?)	
	3.3 s	3.3 s	3.3 s	il	
	3.1 wbr		3.2 m	mo	

SAMPLE	AIR DRY d(A)	I	GLYCOL. d(A)	I	550° C d(A)	I	INTP.	FM.
WBI 7			16.0	vw			mo	Kcl
	12.6	s					mo/il	
					9.7	m	il	
			8.6	w			mo (?)	
	6.2	w(+)					mo/il	
			5.6	m			mo/il	
					4.8	m	il	
	4.5	m	4.5	m	4.4	m	mo	
	4.3	mbr	4.3	mbr	4.2	mbr	ka (?)	
	3.3	s	3.3	vs	3.3	s	il	
	3.1	s			3.2	s	mo	
WBI 8			15.5	vwbr			mo	Kcl
	12.3	s					mo/il	
					9.8	m	il	
			8.6	wbr			mo (?)	
	6.1	wbr					mo/il (?)	
			5.5	wbr			mo/il	
					4.8	m	il	
	4.4	wbr	4.4	wbr	4.4	wbr	mo/il	
	4.2	wbr	4.2	mbr	4.2	wbr	ka	
	3.3	m	3.3	s	3.3	s	il	
	3.1	mbr			3.2	s	mo	
WBI 10			13.2	vvw			mo/il	Kcl
	12.4	wbr					mo/il	
			9.9	vvw	10.0	w(+)	il	
	4.9	vw			5.0	vvw	il	
	3.3	w(+)	3.3	w(+)	3.3	w(+)	il	
					3.2	vwbr	mo/il (?)	
WBI 10.1	13.0	mbr	14.0	wbr	12.6	vvw	mo/il	Kcl
	10.3	mbr	10.0	w	10.0	m	il	
	7.4	m					ka	
	7.3	m	7.2	m			ka	
			5.6	w			mo/il (?)	
	5.5	vw	5.5	w			mo/il	
	5.0	w	5.0	w	5.0	w	il	
					4.4	vw	ka	
	4.2	vwbr	4.2	w	4.3	vw	ka	
	3.8	vvw					ka	
	3.6	m	3.6	m			ka	
	3.4	m	3.4	m	3.4	m	il	
	3.3	m	3.3	s	3.3	s	il	
					3.2	wbr	il (?)	

SAMPLE	AIR DRY d(A) I	GLYCOL. d(A) I	550^{o} C d(A) I	INTP.	FM.
WBI 11		15.2 w		mo	Kcl
	13.8 wbr			mo/il	
	10.5 wbr	10.0 w	10.0 w(+)br	il	
	4.9 wbr		5.0 wbr	il	
	4.5 w	4.5 vw	4.4 wbr	mo	
	3.3 w(+)	3.3 w(+)	3.3 w(+)	il	
			3.2 wbr	mo	
	3.0 w	3.0 w	3.0 w	mo/il	
BCAI 11	13.5 mbr	13.0 vwbr		mo/il	Kcl
		10.0 vwbr	9.8 mbr	il	
	7.2 m	7.2 m	7.2 m	ka	
	6.3 w	6.3 w		mo/il	
	5.0 wbr		4.9 wbr	il	
	4.5 wbr	4.5 w	4.4 w	mo/il	
	4.2 wbr	4.3 w	4.3 w	mo/il	
	3.6 m	3.6 m	3.6 m	ka	
	3.3 m	3.3 m	3.3 m	il	
BCA2 3		13.0 vw		mo/il	Kcl
	12.1 sbr			mo/il	
		10.0 vwbr	10.0 s	il	
	6.1 w			mo (?)	
		5.6 wbr		mo/il	
	5.0 w	5.0 w	5.0 m	il	
		4.5 vvw		mo	
	4.2 wbr	4.2 wbr	4.2 w	il (?)	
	3.3 s	3.3 s	3.3 s	il	
	3.1 sbr		3.2 sbr	mo	
BCA2 7		14.2 wbr		mo/il	Kcl
	12.8 vs			mo/il	
			9.9 s	il	
		8.7 w(+)		mo (?)	
	6.2 m			mo/il (?)	
		5.6 m		mo/il	
			4.8 m	il	
	4.2 vwbr	4.2 vwbr		ka	
		3.4 s		ka	
	3.3 vvw			il	
	3.1 s		3.2 s	mo	
BCA2 8		15.9 m		mo	Kcl
	12.4 vsbr			mo/il	
		10.0 vw	9.7 sbr	il	
		8.6 wbr		mo	
	7.2 s	7.2 s		ka	

SAMPLE	AIR DRY d(A) I	GLYCOL. d(A) I	550^{o} C d(A) I	INTP.	FM.
	6.2 wbr			mo/il (?)	
		5.6 w(+)br		mo/il	
			4.8 mbr	il	
	4.4 vvw	4.4 w	4.4 vvw	mo/il	
	4.2 vwbr	4.3 wbr	4.2 vvw	ka	
	3.6 vs	3.6 vs	3.5 vw	ka	
	3.3 w	3.4 s	3.3 sbr	il	
	3.1 mbr		3.2 s	mo	
BCA2 10		14.5 vwbr		mo/il	Kcl
	12.4 wbr			mo/il	
			9.9 wbr	il	
	7.2 s	7.2 s	7.1 w(+)	ka	
	3.6 s	3.6 s	3.6 w(+)	ka	
	3.3 w	3.4 wbr	3.3 wbr	il	
			3.2 vwbr	mo/il (?)	
NN3 1		13.6 vwbr		mo/il	Kcl
	12.8 mbr			mo/il	
			9.9 wbr	il	
	7.2 s	7.3 s	7.2 vvw	ka	
	4.4 wbr	4.3 vwbr		ka	
	3.6 s	3.6 s	3.6 vvw	ka	
	3.3 w	3.4 w	3.3 w	il	
			3.2 vw	mo/il (?)	
NN3 7a	7.3 s	7.4 s		ka	Kcl
	4.4 wbr	4.5 wbr		mo	
	4.2 wbr	4.2 w	4.3 w	ka	
	3.6 s	3.6 s		ka	
	3.4 m	3.4 m	3.3 m	il	
	3.0 s	3.0 s	3.0 m	mo/il	
NNI 7b		13.2 vvw		mo/il	Kcl
	12.0 mbr			mo/il	
		9.9 w	10.1 sbr	il	
		5.3 wbr		mo/il	
	5.0 mbr		5.0 mbr	il	
	4.5 w	4.5 w	4.5 wbr	mo/il	
	4.3 m	4.3 w	4.3 wbr	mo/il	
	3.8 w			mo/il	
	3.5 vw			ka	
	3.3 s	3.3 s	3.3 s	il	
	3.2 w	3.2 w	3.2 wbr	il (?)	
	3.0 m	3.0 m	3.0 m	mo/il	

SAMPLE	AIR DRY d(A) I	GLYCOL. d(A) I	550° C d(A) I	INTP.	FM.
NNI 11a	11.9 wbr	13.6 wbr		mo/il	Kcl
	10.3 m			mo/il	
		9.8 m	9.8 m	il	
	7.1 m	7.1 m		ka	
	5.0 w	5.0 w	5.0 w(+)	il	
	4.4 vw	4.4 vw		ka	
	4.2 w	4.2 w(+)	4.2 w(+)	ka	
	3.6 m	3.6 m		ka	
	3.3 vs	3.3 vs	3.3 vs	il	
			3.2 w	il (?)	
NNI 11b		13.2 wbr		mo/il	Kcl
	10.4 sbr	10.3 w(+)	9.8 s	mo/il	
	7.2 s	7.4 s	7.1 vw	ka	
	5.0 m	5.1 w(+)	5.0 s	il	
	4.2 w	4.3 w	4.2 w	ka	
	3.6 vs	3.6 s	3.5 w	ka	
	3.3 vs	3.4 vs	3.3 vs	il	
PFS B	13.2 wbr			mo/il	Kcl
	10.1 w	10.1 wbr	10.1w	il	
	7.1 m	7.2 w	7.2 vvw	ka	
	5.0 w	5.0 w	5.0 w	il	
		4.5 w	4.5 w	ka	
	4.4 w	4.4 w		ka	
	4.3 m	4.3 w	4.3 w	ka	
	3.6 m	3.6 m	3.6 w	ka	
	3.4 m	3.4 m	3.4 s	il	
	3.3 s	3.3 s	3.3 s	il	
BullI 1a	13.4 mbr			mo/il	Kcl
		12.0 wbr		mo/il	
	10.1 mbr	10.1 mbr	10.0 sbr	il	
	7.2 s	7.3 s		ka	
	5.0 mbr	5.0 wbr	5.0 mbr	il	
	4.5 w	4.5 wbr	4.4 vwbr	ka	
	4.3 w	4.3 wbr	4.2 w	ka	
	3.6 s	3.6 s		ka	
	3.3 s	3.3 s	3.3 s	il	
LSMtn		13.2 wbr		mo/il	Kcl
Recon	10.9 mbr			mo/il	
		9.9 m	10.0 s	il	
	7.1 m	7.2 m		ka	
	5.0 m	5.0 m	5.0 m	il	
	4.2 w	4.2 w		ka	
	3.6 s	3.6 s		ka	

SAMPLE	AIR DRY d(A) I	GLYCOL. d(A) I	550° C d(A) I	INTP.	FM.
	3.3 s	3.3 s	3.3 s	il	
			3.2 w	il (?)	
BCAS2 12		13.0 wbr		mo/il	Kcl
	10.3 mbr			mo/il	
		10.0 m	10.0 m	il	
	8.1 vw			mo (?)	
	7.3 s	7.2 s		ka	
	5.0 m	5.0 m	5.0 m	il	
			4.6 w	il	
	4.4 w		4.4 w	ka	
	4.3 m		4.3 w(+)	ka	
		4.2 m	4.2 m	ka	
	3.6 vs	3.6 vs		ka	
	3.3 vs	3.3 vs	3.3 vs	il	
Sykes 1	12.1 wbr	13.6 vwbr		mo/il	Ksm
	10.4 wbr	10.3 w		mo/il	
			10.1 mbr	il	
	7.3 w(+)	7.3 w		ka	
	5.0 w	5.0 wbr	5.0 wbr	il	
	4.5 w	4.5 w	4.4 w	ka	
	4.2 w	4.3 w(+)	4.2 w	ka	
	3.6 m	3.6 m		ka	
	3.3 s	3.3 s	3.3 s	il	
Sykes 2	12.6 mbr	13.6 wbr		mo/il	Ksm
	10.4 mbr	10.3 wbr		mo/il	
			10.0 m	il	
	7.2 m	7.2 m	7.0 vw	ka	
	5.0 wbr	5.0 w	5.0 w	il	
	4.2 w	4.2 w	4.2 w	ka	
	3.6 s	3.6 s	3.5 vvw	ka	
	3.3 s	3.3 s	3.3 s	il	
Sykes 3	12.6 mbr	13.6 wbr		mo/il	Ksm
	10.4 mbr	10.3 wbr		mo/il	
			10.0 s	il	
	7.2 m	7.2 m	7.3 w	ka	
	5.0 w(+)	5.0 w	5.0 m	il	
	4.2 w	4.2 w	4.2 w	ka	
	3.6 s	3.6 s	3.5 vw	ka	
	3.3 s	3.3 s	3.3 s	il	