Stratigraphy and depositional environment of the

Mesaverde Group, northeastern portion

of the Big Horn Basin, Wyoming

by

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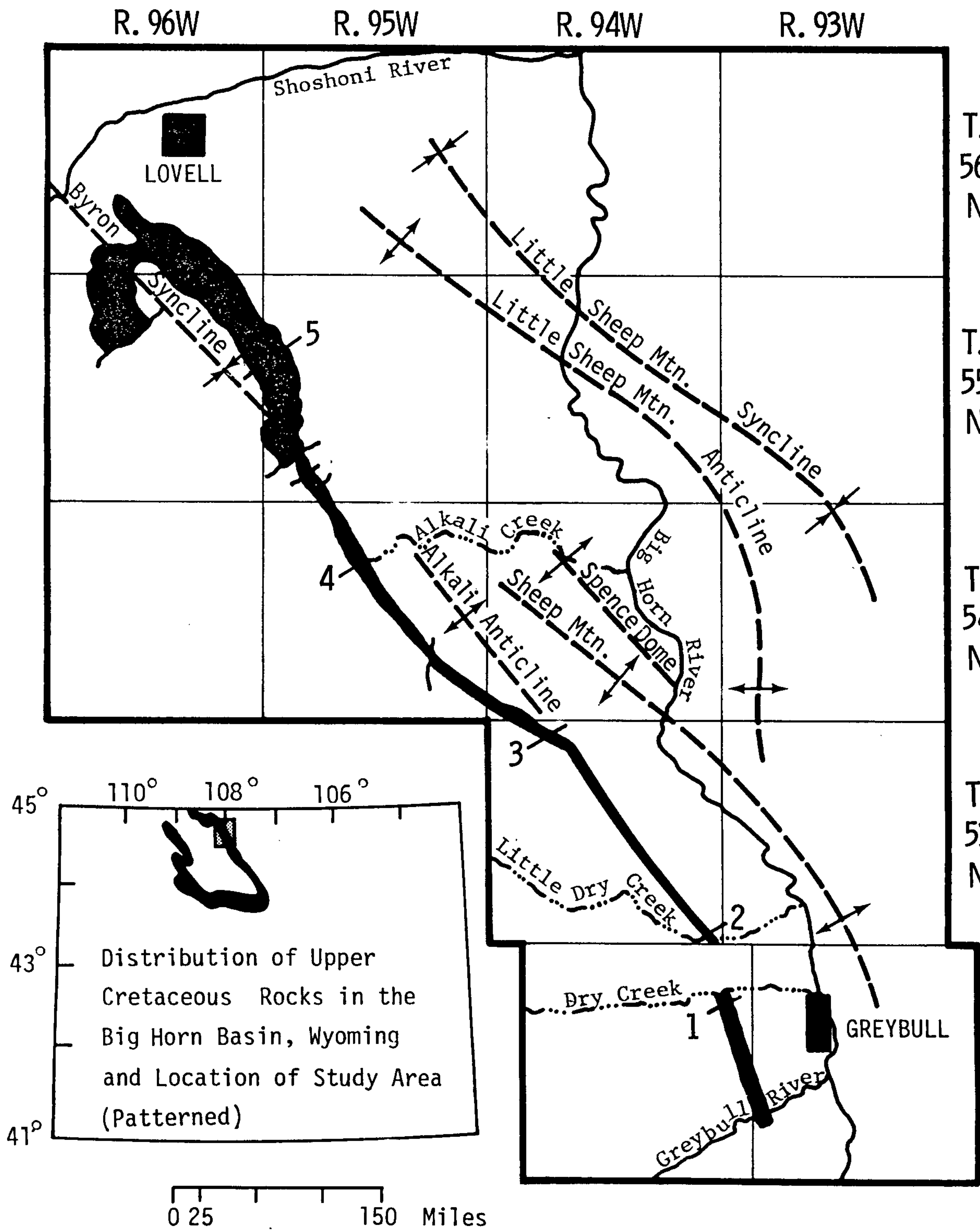
INTRODUCTION

This report presents a study of the stratigraphy and depositional environments of the Mesaverde Group in the northeastern portion of the Big Horn Basin, Wyoming. Exposures of the Mesaverde Group were examined during the summer of 1977 along a northwesterly trending outcrop belt extending from the Greybull River, two miles south of Greybull up to five miles south of Lovell, Wyoming (fig. 1).

Upper Cretaceous sediments in the Big Horn Basin as well as other Western Interior basins have received considerable attention in the last 20 years as a result of discoveries of coal and petroleum reserves. Although regional lithofacies patterns have proven useful in petroleum exploration, little work has been done in the area of paleoenvironmental reconstruction utilizing sequences of primary sedimentary structures. Also the Upper Cretaceous terminology in southern Montana and in the Southern Big Horn Basin is confused and in need of revision. In these regions the terms Montana Group and "Mesaverde" are used for identical rock units. There is then, a need for stratigraphic analysis of the Mesaverde Group in the northern parts of the Big Horn Basin to establish a firm basis for consistent terminology between Montana and areas to the south of the present study area.

The objectives of this study are (1) to describe in detail the rock units presently termed Mesaverde in the northeastern portion of the Big Horn Basin, (2) to clarify the terminology utilized between

Figure 1. Location of study area and measured sections of the Mesaverde Group in the Big Horn Basin.


R. 96W
R. 95W
R. 94W
R. 93W
Shoshoni River
LOVELL
Byron
Syncline
5
Little Sheep Mtn.
Little Sheep Mtn.
Syncline
Anticline
Big Horn River
Alkali Creek
4
Alkali Anticline
Sheep Mtn.
Spence Dome
3
Little Dry Creek
2
Dry Creek
1
GREYBULL
Greybull River
T. 56 N
T. 55 N
T. 54 N
T. 53 N
T. 52 N
45°
110°
108°
106°
43°
41°
Distribution of Upper Cretaceous Rocks in the Big Horn Basin, Wyoming and Location of Study Area (Patterned)
0 25
150 Miles

areas previously described as Mesaverde Formation or Mesaverde Group in southern Montana and the southern Big Horn Basin, (3) to interpret the depositional environments recorded by Mesaverde rocks and (4) to reconstruct the Campanian paleography in the study area and to relate it to previous paleoenvironmental reconstructions in the southern portion of the Big Horn Basin.

Previous Investigations

Geological investigations of the Upper Cretaceous rocks in north-central Wyoming and central Montana began in the mid 1800s with the initiation of the Meek and Hayden surveys. No real interpretive work was initiated, however, until Stanton and Hatcher (1903) and Stanton et al. (1905) working near the Judith River in central Montana, presented evidence regarding the age and stratigraphic position of the Judith River Beds.

Hatcher was the first to recognize the intertonguing nature of the marine and continental strata and presented the first cross-section illustrating the lateral relationships. Hatcher subdivided the rocks of the then Montana Group into the Eagle sandstone, Claggett Shale, Judith River, and Bearpaw Formations, terms which are still applicable throughout Montana and northern Wyoming. Stebinger (1914) pro- provided further evidence regarding the age of the Montana Group through correlations with the Upper Cretaceous Belly River Beds in southern Alberta. Later work by Bowen (1919) confirmed Hatcher's speculations regarding gradations from continental to marine conditions eastward in central Montana during Late Cretaceous Time.

In the Big Horn Basin area, the presence of soft coal in Upper Cretaceous strata led to repeated investigations by the United States Geological Survey. Fisher (1908) was able to trace the subdivisions of the Montana Group defined by Hatcher (1903) and Stanton et al. (1905) southward into the northeastern portion of the Big Horn Basin. Washburne and Woodruff (1907) working independently, accurately traced the Eagle Sandstone from central Montana to the southeastern extremity of the present study area. Hence, the terminology for Mesaverde equivalents throughout the eastern margin in the basin was accurately established during initial investigations.

Controversy regarding the age, terminology, and correlation of Upper Cretaceous rocks erupted as Hewett (1914) applied the name Gebo to Montana Group equivalents. Hewett did however recognize the upper part of the Gebo to be equivalent in age to the Eagle and Claggett and quickly discarded this usage in subsequent publications (figs. 2 and 3).

The term Mesaverde was not applied to rocks of Campanian age in the Big Horn Basin until Lupton (1916) recognized basic similarities in lithology and stratigraphic position with beds already described as Mesaverde in southern Wyoming and Utah. This name was then retained by the geological survey for subsequent mapping in the Big Horn Basin and is still used extensively today.

Recently the Mesaverde in the Big Horn Basin has been the subject of many regionally oriented stratigraphic and paleontologic studies (Asquith, 1974, Cobban, 1969, Gill and Cobban, 1966 and 1973,

Figure 2. Development of nomenclature for the Mesaverde Group in the Big Horn Basin (1905-1926).

			STANTON AND HATCHER, 1905		WASHBURNE, 1909 EAST		WOODRUFF, 1909 WEST		HEWETT, 1914	HEWETT AND LUPTON, 1917, 1926
SYSTEM	SERIES	STAGE	GROUP	FORMA-TION	GROUP	FORMA-TION	GROUP	FORMA-TION	FORMATION	FORMATION
CRETACEOUS	UPPER	CAMPANIAN	MONTANA	JUDITH RIVER & BEARPAW	MONTANA	JUDITH RIVER & BEARPAW	MONTANA	UNDIFFERENTIATED MONTANA	GEBO	MESAVERDE
				CLAGGETT		CLAGGETT				
				EAGLE		EAGLE		EAGLE		

Figure 3. Development of nomenclature for the Mesaverde Group in the Big Horn Basin (1958-1979).

			RICH, 1958	SEVERN, 1961			MILLER ET AL, 1965		GILL & COBBAN, 1966		MACKENZIE, 1975			THIS PAPER		
SYSTEM	SERIES	STAGE	FORMATION	GROUP	FORMATION	MEMBER	WEST	EAST	FORMATION	MEMBER	GROUP	FORMATION	MEMBER	GROUP	FORMATION	MEMBER
CRETACEOUS	UPPER	CAMPANIAN	MESAVERDE	MESAVERDE	JUDITH RIVER	TEAPOT ?	MESAVERDE	JUDITH RIVER	MESAVERDE	TEAPOT	MESAVERDE	JUDITH RIVER	TEAPOT	MESAVERDE	JUDITH RIVER	TEAPOT
					CLAGGETT			CLAGGETT				CLAGGETT			CLAGGETT	
					EAGLE			EAGLE				EAGLE			EAGLE	

Miller, Barlow, and Haun, 1965, MacKenzie, 1975, Rea and Barlow, 1975, and Weimer, 1961 and 1970). Mesaverde terminology in the eastern half of the basin still retains the original subdivisions as described by Stanton and Hatcher (1903) and Stanton et al. (1905). In the southwestern areas, marine units are not present in the Mesaverde, making subdivisions of the entirely nonmarine section very tenuous. Controversy is still continuing regarding the elevation of the Mesaverde from formation to group status. However, recent authors have tended to consider the unit a group (MacKenzie, 1975, Severn, 1961). In this report the Mesaverde is treated as a group with the Eagle, Claggett, and Judith River Formations and the Teapot Sandstone Member recognized in ascending order. The Eagle, Claggett, and Judith River Formations and the Teapot Sandstone Member are easily traceable in the field and on areal photographs and are mappable on a scale of 1:25,000 and should be treated with proper rank.

Method of Study

After accurately establishing lower and upper boundaries, five critically located exposures of the Mesaverde Group were measured and described in detail. Emphasis was placed on careful documentation of vertical changes in lithology and primary sedimentary structures which are important in the determination of sedimentary environments represented by Mesaverde rocks. Major sandstone and shale bodies were traced laterally in the field and on areal photographs to determine geometric relations and lateral facies changes.

Over 550 oriented rock samples were collected and keyed to specific stratigraphic horizons. Selected samples from key beds were then subjected to various laboratory analyses including mechanical analysis, thin section examination, x-ray diffraction and SEM analysis. Sandstone mineralogy was determined by petrographic examination, although staining techniques were applied to over 60 mounts to determine potash and plagioclase feldspar ratios. An average of 300 grains were counted in over 30 thin-sections to determine quantitative mineralogy and also compositional changes occurring between sedimentary environments.

More than 250 samples were studied to determine grain-size distributions useful in distinguishing between various sedimentary environments. Grain-size parameters (Folk and Ward, 1957, Passega, 1957) were plotted graphically as a means of characterizing depositional agents operating within the wide variety of sedimentary environments recorded in Mesaverde rocks.

Regional Stratigraphic Setting

The Mesaverde strata described in this study are exposed along the northern and eastern margin of the Big Horn Basin in the north-central portion of Bighorn County, Wyoming. Throughout the outcrop belt, the beds strike north-northwest, approximately parallel to the inferred depositional strike (Gill and Cobban, 1973, Houston and Murphy, 1977, Severn, 1961), and dip from 11 to 55 degrees west, in a basinward direction (fig. 1).

The Mesaverde Group represents only a small part of a series of clastic wedges deposited in the Cretaceous Western Interior seaway which extended from present-day Alaska to Mexico (Gill and Cobban, 1973, Masters, 1965, 1967).

In Colorado and Utah, the sediment source for Mesaverde sediments is considered to be the Sevier Orogenic Belt (Armstrong and Oriel, 1965, Masters, 1965) expressed today as a north-south trending overthrust belt in central Utah and north central Wyoming. Late Cretaceous intrusions including the Idaho Batholith and the Boulder Batholith along with the Elkhorn Volcanics in west central Montana, have been correlated with major regressional episodes during Campanian time and are thought to be important sediment sources for the Mesaverde in the Big Horn Basin (Houston and Murphy, 1977). Throughout the Western Interior, sediments deposited during regressive phases are thicker, coarser-grained, and more nonmarine towards source areas to the west (Spieker, 1949).

Throughout the Late Cretaceous, sedimentation patterns in the Western Interior epicontental sea were characterized by frequent strandline migrations due to variations in the amount of sediment influx and basin subsidence (Masters, 1965). During the Campanian Epoch the shoreline in Wyoming trended north-south, parallel to the adjacent source areas. Local variations in shoreline trends were due to tectonic activity during sedimentation or deltaic depocenters (Asquith, 1974, Weimer, 1970, and Zapp and Cobban, 1960). In southern Montana and northern Wyoming, shoreline trends averaged north, 30 to 50 degrees west with indentations related to local delta centers (Asquith, 1974, Gill and Cobban, 1973).

In the Big Horn Basin, Mesaverde strata consist of wedges of both marine and transitional marine sediments which interfinger eastward with the Claggett tongue of the Cody. In detail, the major regressive-transgressive clastic wedges can be subdivided into two regressive units separated by westward extending tongues of marine shale (Miller, Barlow, and Haun, 1965, MacKenzie, 1975). The base of each regressive sequence consists of beach sandstones and shelf shales, which in turn are overlain and interfinger with strata representing transitions from marine to continental environments. These in themselves display several cycles of deposition.

Basal units of both the Eagle and the Judith River cycles consist of massive beach sandstones, 5 to 35 feet thick, which interfinger complexly with overlying marine siltstones and shales. The Judith River cycle differs from the Eagle in the study area in that continental environments are recorded by its upper 400 feet, while the Eagle consists of predominately beach and shelf deposits. Both of these regressional phases record periods of frequent strandline oscillation resulting in thick, repetitive sequences at the base of each cycle. In contrast the Cody and Claggett transgressive phases occurred during comparatively short time intervals commonly resulting in the truncation of the marine sequences normally expected at the top of a complete regressive-transgressive cycle (Gill and Cobban, 1973).

PETROLOGY

Petrologic examination of Mesaverde sediments included detailed collection of over 300 kilograms of oriented samples associated with key stratigraphic horizons and major sand and shale units throughout the study area. Preparation of over 250 rock samples for mechanical analysis included disaggregation, pretreatment, and sieving through one-half unit phi screens, and hydrometer analysis of the less than four phi size grades. Sandstones were prepared for grain-mount and thin section work by vacuum impregnation and were examined for distinguishing textural and compositional characteristics. Petrologic characteristics provide valuable insight with regards to hydrodynamic conditions within the basin of deposition and are a valuable aid in the reconstruction of ancient sedimentary environments.

Texture

Textural parameters of Mesaverde sediments were resolved by the use of both mechanical and microscopic techniques. Disaggregation, pretreatment, and grain-size analysis were performed using conventional techniques (Folk, 1974, Royse. 1968).

Most sandstone samples were cemented by calcite or dolomite and were pretreated with 10 percent HCL solution, washed and dried before mechanical separations were performed. Sediments were separated into one half unit phi fractions by means of nested sieves and grain size components smaller than four phi were analyzed using the Bouycos 152-H hydrometer.

The results of ten randomly selected hydrometer runs were compared with pipetting from an aqueous solution as a test for the accuracy and precision of the hydrometer method. Deviations of the Bouycos 152-H hydrometer from the results of pipette analysis of identical samples were minimal. Average deviation for each unit phi size were less than 1.5 percent for each sample studied.

Eagle sediments

Texturally, the majority of Eagle sediments may be classified as silty sands (Folk, 1954) with subordinate amounts of sand and muddy sand (fig. 4). Mean grain-size ranges from $\phi 4.42$ to $\phi 3.94$ with the average graphic mean for all Eagle sands $\phi 3.70$. Most sand grains are subrounded to subangular and moderately spherical and are poorly to very poorly sorted (table 1.)

Commonly Eagle and Judith River sandstones are cemented with secondary sparry calcite constituting 20 to 30 percent of the total rock volume. In thin section, calcite cemented sands frequently display replacement of framework elements as evidenced by embayments and replacement of detrital grains by calcite (plate 1A). Often, framework and calcite grains display poikolitic relationships with masses of optically contigous calcite enclosing isolated framework grains (plate 1A). Detrital grains are in point or line contact when the calcite cement is not present.

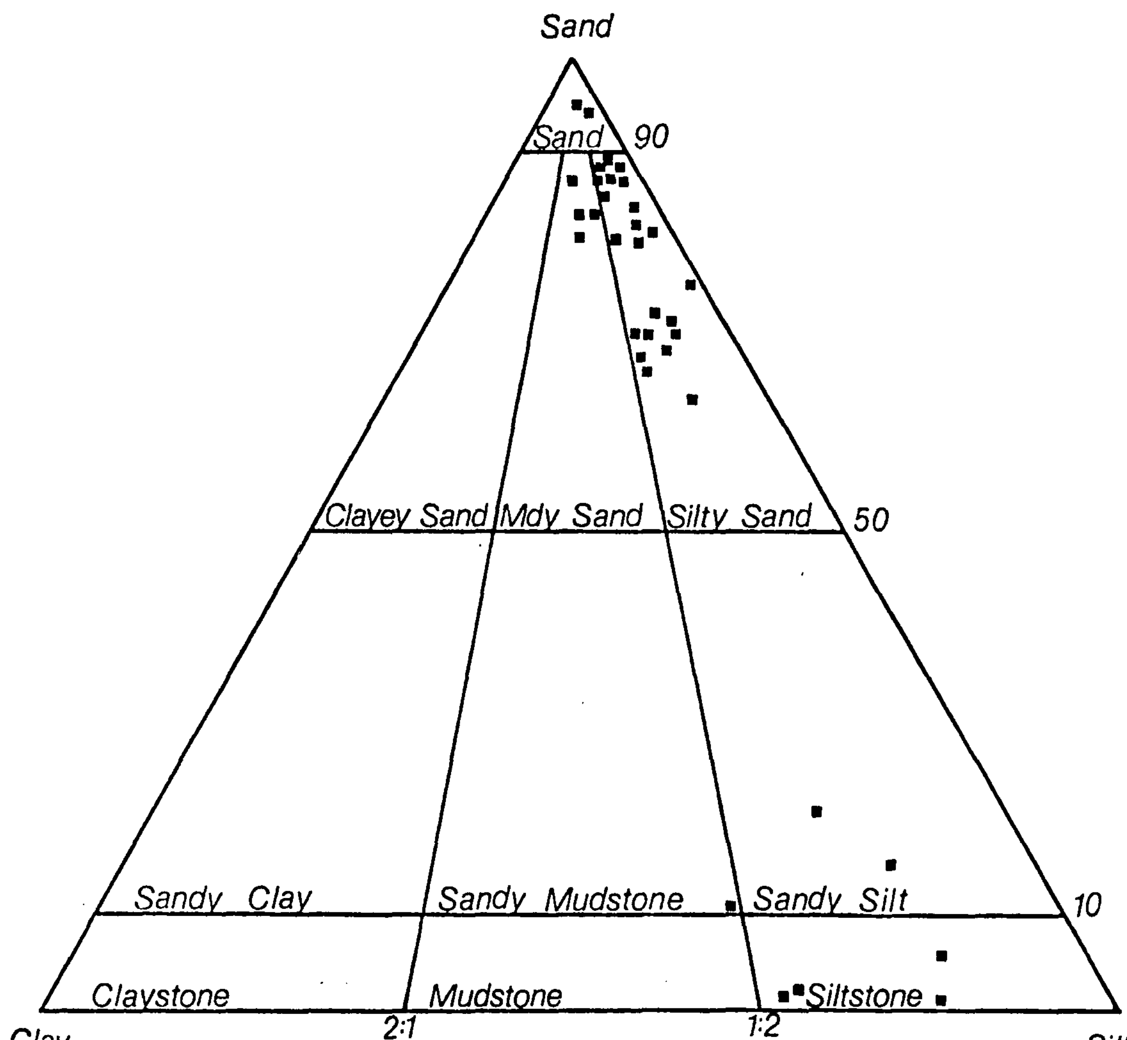


Figure 4. Size classification of Eagle Sediments, after Folk (1954)

Fine-grained sediments are sandy siltstones and siltstones based on terminology by Folk (1954). Analysis of these sediments indicate that they are poorly to very poorly sorted with inclusive graphic standard deviation ranging from 1.60 to 3.35 (Folk, 1974). These rocks occur as thin lenticular or wavy beds in sandstones or as relatively thick beds.

Judith River sediments

Judith River sediments are mainly sandstones and silty sands (Folk, 1954)(fig. 5). Grain-size data for the Judith River is presented in table 1 and displays well-developed coarsening-upwards sequences which are largely absent in the Eagle. Mean grain-size values are strikingly uniform throughout the Judith River with values ranging from $\phi 3.78$ to $\phi 2.79$. These values probably indicate relatively uniform conditions with respect to rate of sediment influx and basin subsidence in the study area throughout the time represented by these rocks.

Inclusive graphic standard deviation ranges from 0.08 to 2.01 while most sediments are moderately well to poorly sorted. Roundness and sphericity differ very little throughout the Judith River section and are quite similar to Eagle sandstones.

Teapot sediments

Teapot sediments display vertical and lateral variations in grain-size which contrast sharply with the general homogeneity displayed by

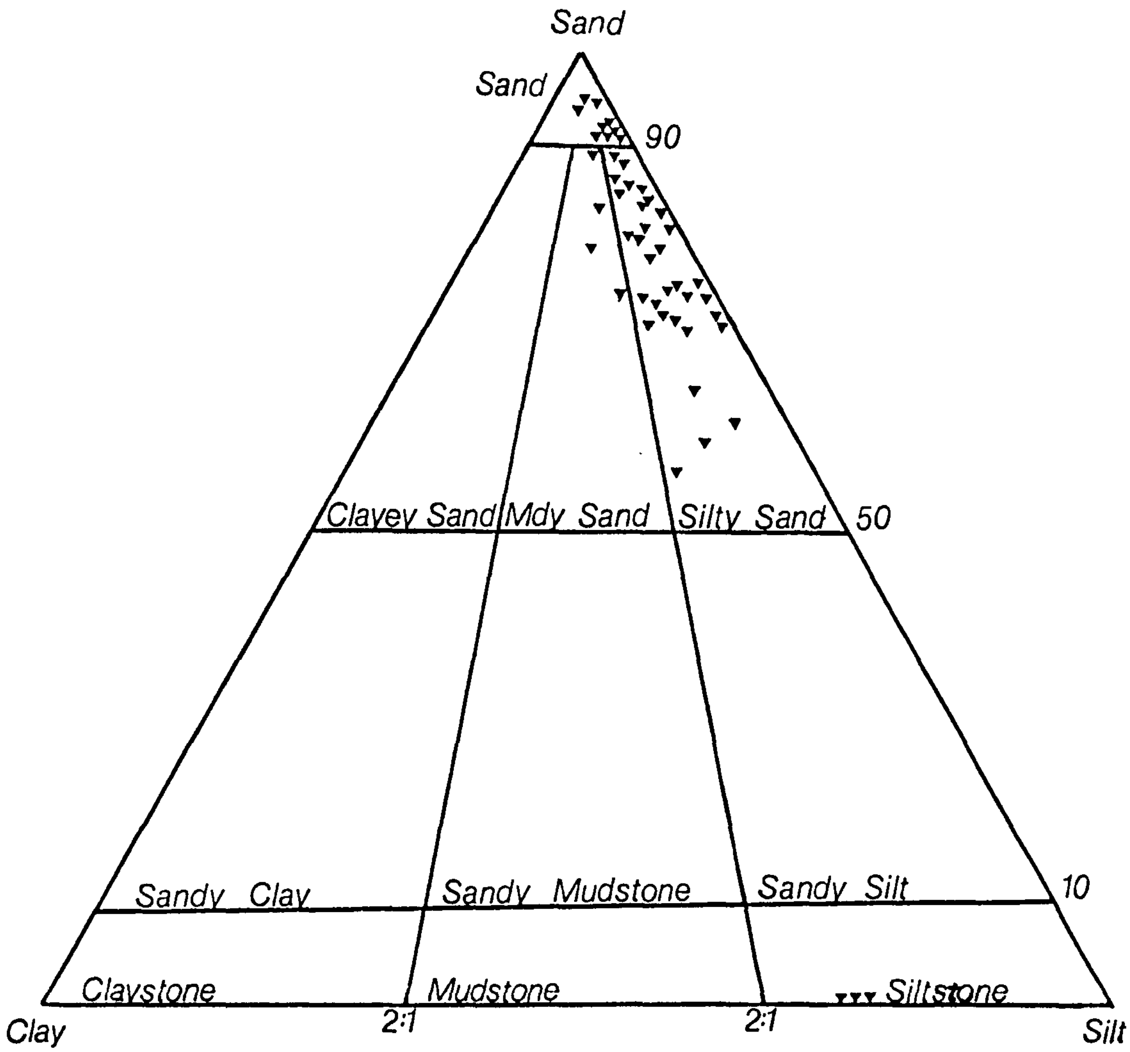


Figure 5. Size classification for Judith River Sediments, after Folk (1954).

the Eagle and Judith River. Diagnostic properties of the Teapot Sandstone Member are as follows:

1. A general, though fluctuating coarsening-upwards sequence throughout the unit.
2. Distinct fining-upwards textural sequences are developed over one to three foot intervals which are repeated many times.
3. Lateral variations in grain-size over comparatively short intervals with an increase in Mz values in a northwesterly direction.
4. Bimodal distributions in Teapot sediments are common but are not related to hydrodynamic conditions at the time of deposition. Major populations occur in the medium sand and clay size fractions; the abundance of clays caused by authigenic clay mineral formation.

<u>Statistical parameters, comparisons and depositional processes</u>

Grain-size distributions were determined and statistical parameters (Folk and Ward, 1957) were calculated to facilitate comparisons of major lithofacies of the Mesaverde Group. Table 1 lists values obtained for Mean Grain Size (Mz), Inclusive Graphic Standard Deviation (Sorting), Inclusive Graphic Skewness (SKi), and Graphic Kurtosis (Kg). The coarsest one percent and median grain-size (Passega, 1957) were plotted graphically on scatter diagrams as a means of comparison as were the above parameters for use as aids in environmental interpretations of Mesaverde sediments.

Table 1. List of statistical parameters for representative samples within the Mesaverde Group

Sample Number	Mean	Deviation	Skewness	Kurtosis	Coarse 1%	Median
<u>Eagle Formation</u>						
1 02	3.83	1.32	0.49	3.05	2.43	3.69
1 03	3.23	0.79	0.34	1.75	2.11	3.18
1 05	3.09	0.55	0.32	3.09	2.06	2.97
1 06	3.40	0.78	0.20	2.04	1.97	3.41
1 06	3.03	0.52	0.21	1.01	2.02	2.96
1 08	2.40	1.50	0.42	2.30	2.03	3.27
1 08	3.87	1.46	0.53	3.30	2.08	3.64
1 09	6.28	2.10	0.69	2.07	4.01	5.55
1 09	6.83	2.95	0.54	1.57	4.03	5.78
1 10	3.81	1.30	0.54	2.31	2.07	3.49
1 11	3.18	1.35	0.60	3.12	2.15	3.02
1 11	5.73	1.85	0.67	3.47	4.02	4.91
1 13	2.99	0.42	0.39	2.62	2.02	2.94
1 14	3.25	0.89	0.64	1.65	2.28	3.01
1 14	3.56	1.03	0.23	1.74	2.05	3.49
1 16	4.03	1.59	0.54	2.53	2.09	3.70
1 19	4.14	1.49	0.68	3.78	2.58	3.77
1 20	3.56	0.55	0.15	2.03	2.16	3.57
1 21	3.39	0.78	0.31	1.44	2.25	3.33
1 22	5.86	2.00	0.68	3.42	4.02	4.95
1 23	4.11	2.12	0.73	4.35	2.13	3.61
1 24	3.48	1.01	0.24	2.13	2.05	3.46
1 26	3.34	0.94	0.18	1.47	2.04	3.34
1 27	5.11	2.12	0.70	2.60	4.02	4.95

Table 1. (continued)

Sample Number	Mean	Deviation	Skewness	Kurtosis	Coarse 1%	Median
Basal Lower Judith River						
1 27	6.14	2.26	0.74	2.00	4.02	4.95
1 29	3.93	1.13	0.14	4.21	2.09	3.53
1 30	3.85	1.34	0.32	3.19	2.05	3.42
1 31	3.42	0.80	0.10	1.45	2.03	3.23
1 31	3.18	0.58	0.54	2.51	2.27	3.00
1 32	3.35	0.51	0.52	2.00	2.35	3.27
1 32	3.03	0.59	0.52	3.09	2.19	2.80
1 32	4.10	1.35	0.47	2.24	2.08	3.67
1 34	4.41	0.08	0.44	2.47	2.05	3.57
1 34	3.96	1.05	0.38	2.49	2.08	3.56
1 34	4.02	1.20	0.38	2.77	2.07	3.66
1 34	3.90	1.14	0.33	2.45	2.06	3.55
1 36	4.31	1.12	0.71	1.49	2.35	3.84
1 36	3.42	0.89	0.54	1.49	2.06	3.00
1 36	3.14	0.75	0.52	1.69	2.08	2.89
1 39	3.61	0.63	0.30	2.51	2.23	3.61
1 39	3.06	0.01		1.27	2.12	3.69
1 39	3.65	0.60	0.32	2.77	2.46	3.65
1 39	3.42	0.70	0.24	2.00	2.06	3.61
1 39	3.71	1.15	0.34	2.22	2.05	3.51
1 39	3.33	0.71	0.27	1.59	2.25	3.31
1 39	3.29	1.21	0.45	2.92	2.16	3.23
1 39	3.17	0.79	0.47	1.80	2.19	3.05
1 39	2.97	0.94	0.34	1.30	2.01	2.89
1 39	3.22	0.89	0.50	2.04	.224	3.53
1 40	3.87	1.04	0.57	1.83	1.13	3.26
1 41	3.79	0.52	0.27	2.01	2.21	3.54

Table 1. (continued)

Sample Number	Mean	Deviation	Skewness	Kurtosis	Coarse 1%	Median
Basal Lower Judith River (continued)						
1 42	3.78	0.70	0.26	2.71	2.15	3.56
1 42	4.29	1.98	0.40	3.01	2.37	3.74
1 43	4.01	2.01	0.56	6.08	2.14	3.73
4 11	3.77	0.47	0.48	2.03	2.51	3.53
4 11	3.43	0.40	0.36	3.46	2.84	3.34
4 11	3.43	0.40	0.95	3.46	2.84	3.34
4 11	3.36	0.47	0.70	2.94	2.59	3.25
4 11	3.31	0.47	0.70	2.94	2.59	3.25
4 11	2.99	0.33	0.09	3.70	1.84	2.97
4 11	2.49	0.45	0.14	1.89	1.67	2.50
4 11	2.58	0.54	0.31	2.16	1.86	2.58
6 3 2	3.42	0.60	0.09	1.95	2.12	3.16
2 8 1	3.72	0.50	0.79	3.27	2.10	3.61
Barrier Island Units						
1 44	3.76	0.21	0.83	2.48	2.29	3.60
1 44	2.73	1.77	0.88	2.34	2.31	4.63
1 44	3.16	1.13	0.30	1.82	2.02	3.15
1 45	3.09	0.70	0.53	1.45	2.12	2.89
1 45	3.38	1.12	0.42	2.58	2.27	3.30
1 45	2.25	0.53	0.41	1.53	1.59	2.18
1 45	2.62	0.92	0.29	2.48	1.09	2.87
1 45	2.74	1.01	0.42	3.09	1.13	2.36
4 13	3.66	0.43	0.06	1.99	2.83	3.61
4 13	3.52	0.46	0.42	2.31	2.65	3.30
4 13	3.35	0.45	2.51	3.73	2.29	3.28

Table 1. (continued)

Sample Number	Mean	Deviation	Skewness	Kurtosis	Coarse 1%	Median
Barrier Island Units (continued)						
4 13	3.42	0.43	0.67	2.66	2.88	3.23
4 13	3.20	0.63	0.48	2.25	2.11	3.00
4 13	2.19	0.68	0.40	2.44	1.64	2.91
4 13	2.58	0.65	1.21	3.93	1.58	2.50
4 13	2.96	0.69	0.69	2.67	1.61	2.66
3 10	3.20	0.60	0.63	2.37	2.12	3.03
3 10	2.90	0.68	0.70	2.93	1.66	2.69
3 10	3.08	0.65	0.77	2.80	1.74	2.76
3 10	2.90	0.68	0.70	2.93	1.66	2.69
3 10	3.02	0.73	0.55	2.18	1.66	2.73
6 10	3.51	0.45	0.34	2.42	2.61	3.37
6 10	3.46	0.54	0.34	2.42	1.95	3.37
6 10	3.43	0.54	0.15	2.50	2.24	3.18
6 10	3.34	0.56	0.62	2.61	2.19	3.11
6 10	3.17	0.61	0.41	2.94	1.70	2.98
6 10	3.43	0.45	0.59	3.38	2.45	3.31
6 10	3.26	0.60	0.45	2.24	2.13	3.03
6 10	2.67	0.65	0.92	3.38	1.73	2.60
6 10	2.68	0.65	0.91	3.59	1.59	2.55
Upper Judith River						
1 46	4.72	2.08	0.47	2.88	2.36	4.58
1 46	2.89	0.93	0.45	1.72	2.02	2.77
1 46	4.71	2.15	0.82	2.17	2.56	3.77
1 47	2.93	1.26	0.50	1.74	2.00	2.76
1 48	3.88	0.77	0.43	2.01	2.31	3.77

Table 1. (continued)

Sample Number	Mean	Deviation	Skewness	Kurtosis	Coarse 1%	Median
<u>Upper Judith River (continued)</u>						
1 48	3.16	0.82	0.12	1.57	2.03	3.21
1 48	3.54	0.92	0.29	3.64	2.08	3.52
1 49	3.18	0.80	0.44	1.80	2.16	3.08
1 49	3.28	0.98	0.12	2.29	2.04	3.38
1 49	4.13	0.03	1.04	1.04	2.20	3.88
1 49	3.82	0.86	0.49	4.45	2.28	3.70
1 50	4.05	1.27	0.57	4.07	2.39	3.87
1 50	3.82	1.24	0.63	3.04	2.16	3.58
1 51	3.72	0.85	0.44	3.32	2.27	3.67
1 51	3.72	1.09	0.40	3.14	2.11	3.61
1 51	3.71	1.06	0.32	1.20	2.07	3.58
1 51	4.00	1.50	0.56	3.55	2.19	3.80
1 51	3.54	0.93	0.21	1.80	2.06	3.49
1 52	3.53	0.51	0.18	2.16	2.37	3.27
3 15	3.27	0.56	0.73	2.50	2.14	3.07
3 15	3.22	0.56	0.77	2.64	2.12	3.03
3 15	3.28	0.56	0.20	2.69	2.08	3.09

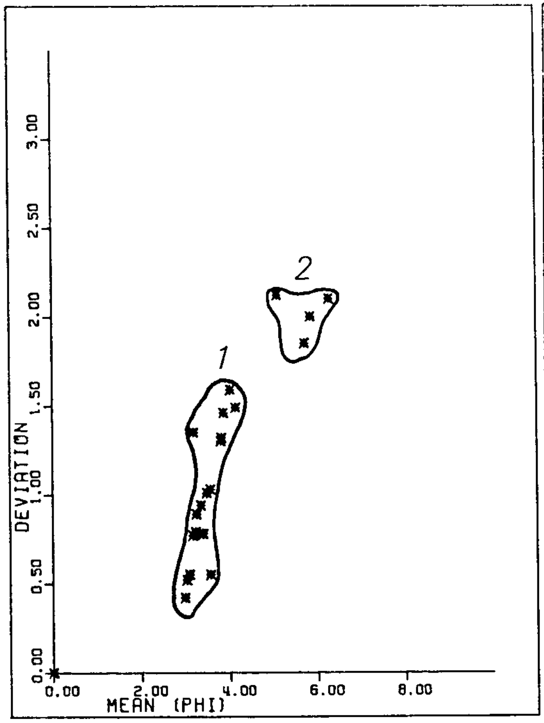
Depositional Agents and Statistical Comparisons

Grain-size parameters have proven useful in the characterization of sediment population distribution patterns as described by Folk and Ward (Folk and Ward, 1957, Mason and Folk, 1958). Statistical parameters such as skewness and kurtosis show in a general sense, the deviations of a sediment population for the normal distribution, but are of little use in paleoenvironmental reconstruction when treated singularly. However, when these parameters are selectively compared on scatter diagrams, certain major environments and associated depositional processes can be recognized (Cameron and Hardarshar, 1977, Folk and Ward, 1957, Mason and Folk, 1958, Nordstrom, 1977, Passega, 1957, 1964, Passega and Byramjee, 1969, and Royse, 1968).

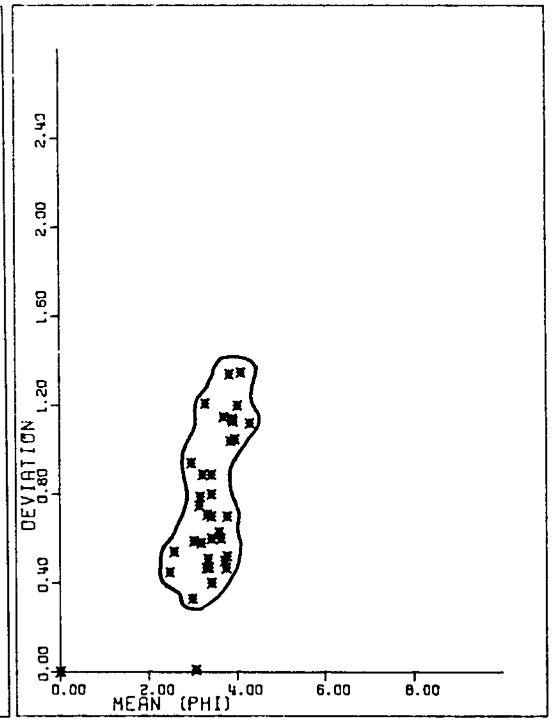
<u>Inclusive graphic standard deviation versus mean grain-size</u>

Scatter diagrams of mean grain-size versus graphic standard deviation reveal two basic patterns considered significant by the writer in interpreting depositional environments represented by the Mesaverde Group (figs. 6A-B). Figure 6A shows two distinct groupings of Eagle sediments consisting of sandstone (1) and siltstones (2) in which sorting systematically decreases from moderately well-sorted to poorly sorted along with a progressive decrease in grain-size (fig. 6B). Lower and Upper Judith River sediments show an almost identical pattern to that of the Eagle suggesting similarities between depositional processes acting during the deposition of these sediments.

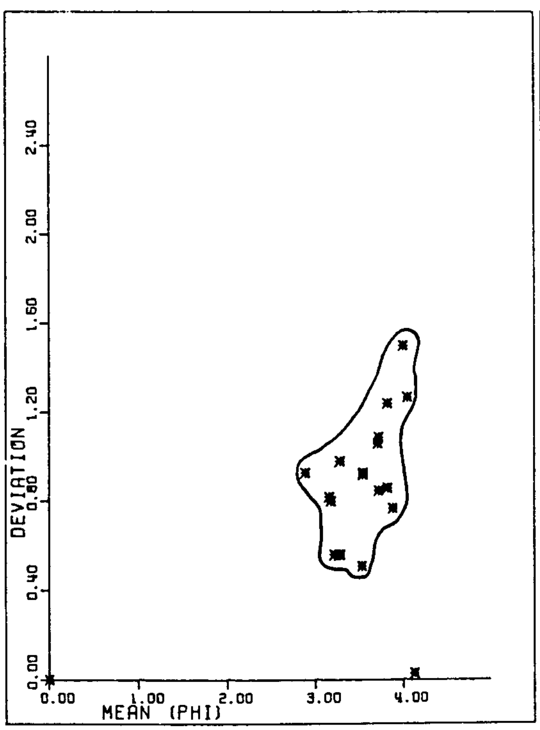
Figures 6A through 6D. Scatter diagrams of mean versus inclusive graphic standard deviation for Eagle and Judith River sediments.


DEVIATION
0.00
0.50
1.00
1.50
2.00
2.50
3.00
0.00
2.00
4.00
6.00
8.00
MEAN (PHI)
1
2

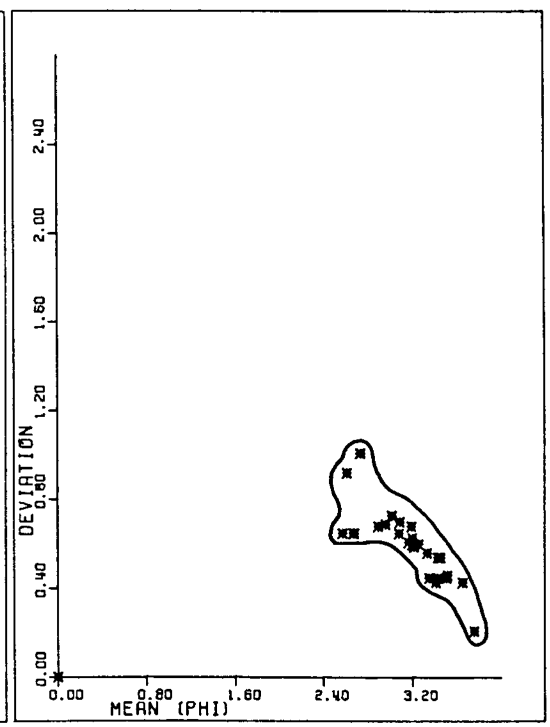
A


DEVIATION
0.00
0.40
0.80
1.20
1.60
2.00
0.00
2.00
4.00
6.00
8.00
MEAN (PHI)

B


DEVIATION
0.00
0.40
0.80
1.20
1.60
2.00
2.40
0.00
1.00
2.00
3.00
4.00
MEAN (PHI)

C


DEVIATION
0.00
0.40
0.80
1.20
1.60
2.00
2.40
0.00
0.80
1.60
2.40
3.20
MEAN (PHI)

D

Recognition of a significant relationship between grain-size and sorting was introduced by Folk and Ward (1957). Graphic comparisons by these authors resulted in sinsoidal plots with cobble and silt size material generally displaying poorer sorting than sand size material. It was also realized at this time that curve positions on comparative plots can be quite variable and are controlled by the size of materials available for transport.

Figures 6A, B and C display linear decreases in sorting values along with decreases in grain-size similar to that expected in marine beach and shelf deposits. The coarsest sediments are deposited along the foreshore of the beach while finer materials remain in suspension and are carried offshore (Clifton, 1976). Increased sediment sorting in the foreshore and upper shoreface beach deposits is the result of the winnowing activity of swash and backswash, tidal, and longshore currents which concentrate a narrow range of particle size. The poorer sorting frequently encountered in the shoreface-offshore transition zone and the offshore zone may be the result of sediment mixing during storms in which sand ripples or dunes periodically migrate seaward.

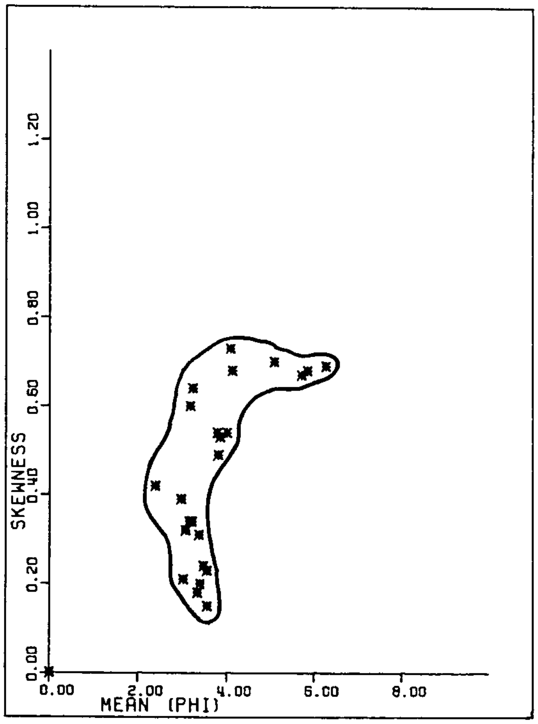
Comparative graphs of coarsening-upward barrier bar sequences in the Judith River Formation reveal patterns which contrast with the patterns described above. As shown by figure 6D, sorting systematically decreases upward with increasing grain-size. Similar regressive sand sequences subjacent to coal-bearing beds are common throughout the Cretaceous deposits in the Western Interior which are considered barrier or distributary mouth bars, depending on their proximity to major deltaic systems (Weimer, 1970). Patterns displayed by these sand bodies could be explained as a result of rapid seaward progradation of deltaic distributary channels resulting in coarsening-upwards cycles which are poorly sorted proximal to the river channels. Evidently the amount of time required for current sorting processes to operate did not keep pace with the rapid progradational rates associated with barrier bars.

Skewness versus mean size

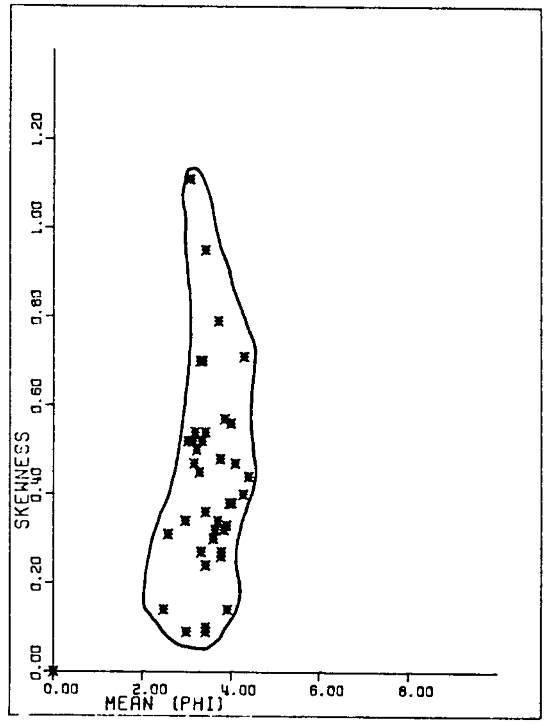
Scatter diagrams of skewness versus mean grain-size are useful in distinguishing beach deposits from the upper portions of barrier bar units. Beach sandstones in the Eagle and Judith River Formations share a common characteristic in that skewness decreases dramatically with grain-size in all cases (figs. 7A and C). Sediments interpreted as lower shoreface on the basis of bedform sequences by the writer are finer-grained with higher skewness values while upper shoreface and foreshore beds are coarser-grained with lower skewness values. These results would be expected in a high energy beach environment where strong shoreward swash current activity winnows out finer-grained components concentrating coarser materials near the upper foreshore and berm crest zones of the beach (Houston and Murphy, 1977). This writer interpretes high skewness values to be the result of the removal of finer-grained material from foreshore zones and deposition on back beach zones. A back beach dune interpretation is substantiated by the association of overlying lagoonal sediments and associated bedform features. Friedman (1961, 1967) and Mason and Folk (1958) were able to distinguish between beach and associated barrier island dune deposits by plotting skewness against mean grain-size. They attributed the differences in skewness to variances

Figures 7A through 7D. Scatter diagrams of mean versus skewness

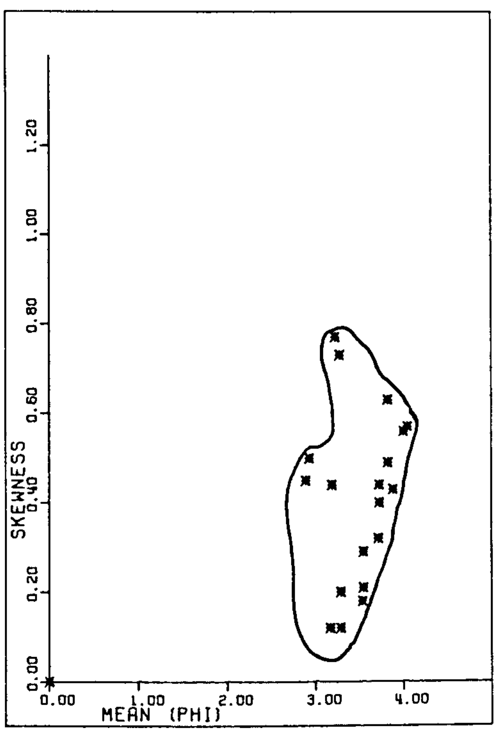
Eagle and Judith River sediments.


SKEWNESS
0.00
0.20
0.40
0.60
0.80
1.00
1.20
MEAN (PHI)
0.00
2.00
4.00
6.00
8.00

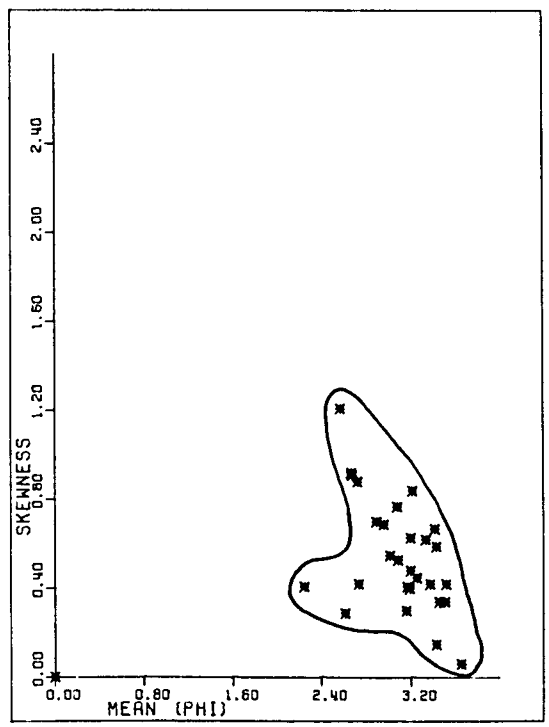
A


SKEWNESS
0.00
0.20
0.40
0.60
0.80
1.00
1.20
MEAN (PHI)
0.00
2.00
4.00
6.00
8.00

B


SKEWNESS
0.00
0.20
0.40
0.60
0.80
1.00
1.20
MEAN (PHI)
0.00
1.00
2.00
3.00
4.00

C


SKEWNESS
0.00
0.40
0.80
1.20
1.60
2.00
2.40
MEAN (PHI)
0.00
0.80
1.60
2.40
3.20

D

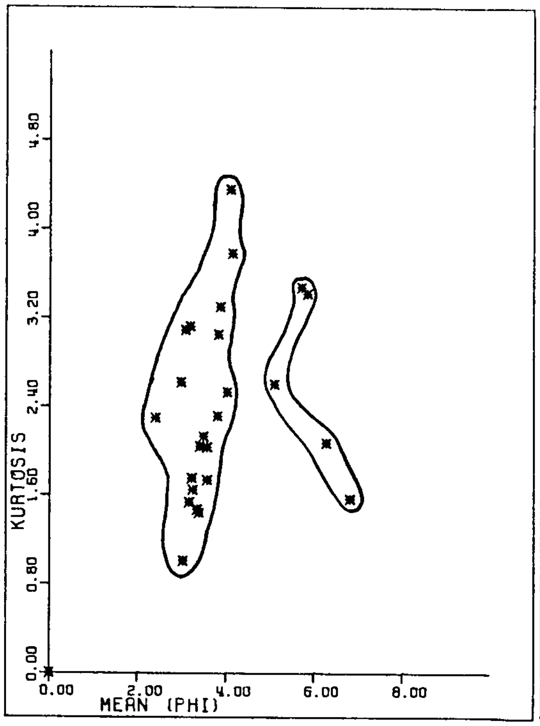
in current and wind sorting processes in beach and beach crest barrier island environments.

Thick barrier island sandstone bodies in the Judith River Formation yield scatter diagrams of skewness versus mean grain-size much unlike those of beach deposits (fig. 7D). In all cases, these units show skewness increasing upwards as grain-size increased, a trend not expected as a result of beach sorting processes. The results of research by Friedman (1961, 1967), and Folk and Ward (1957) suggest trends of this nature may be the result of eolian processes operating in the beach crest or back beach zones of a barrier island environment. Finer components are frequently blown landward from the adjacent beach and may be preserved at the top of ancient barrier bar sandstones.

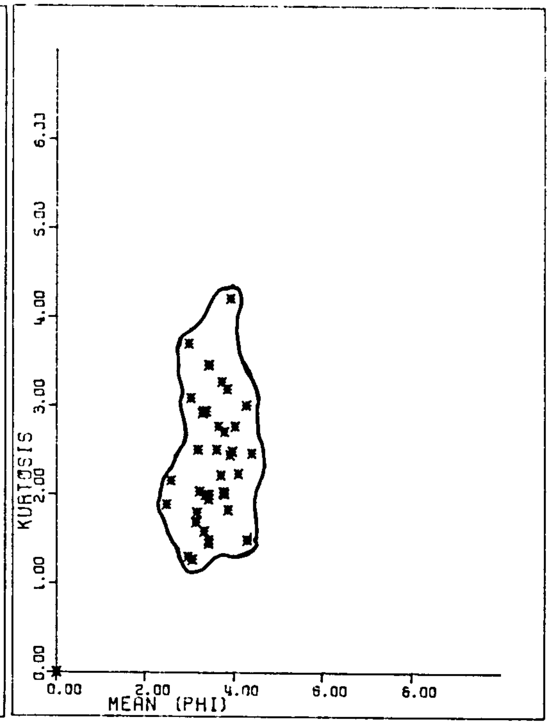
Kurtosis versus mean size

Kurtosis is a measure of the peakedness of the particle size distribution. It is a comparison of the sorting values in the central portion of the distribution to those of the tails of the distribution. Kurtosis values for all Mesaverde sediments are quite high ranging from 1.00 to 4.50 with most values leptokurtic to very leptokurtic (Folk, 1974). Beach sandstones in the Eagle and Judith River Formations display decreasing kurtosis values with decreasing grain-size (figs. 8A-C). In coarsening-upwards beach deposits kurtosis values systematically decrease upwards from 3.00 to 1.40 (fig. 8D). These values are not indicative of beach sorting processes and suggest that rates of sediment influx increase throughout regressive phases and are

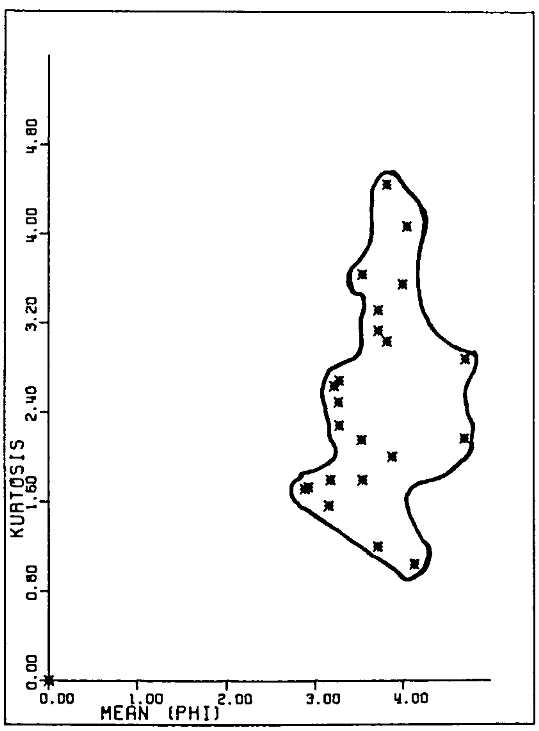
Figures 8A through 8D. Scatter diagrams of mean versus kurtosis for Eagle and Judith River sediments.


KURTOSIS
0.00
0.80
1.60
2.40
3.20
4.00
4.80
MEAN (PHI)
0.00
2.00
4.00
6.00
8.00

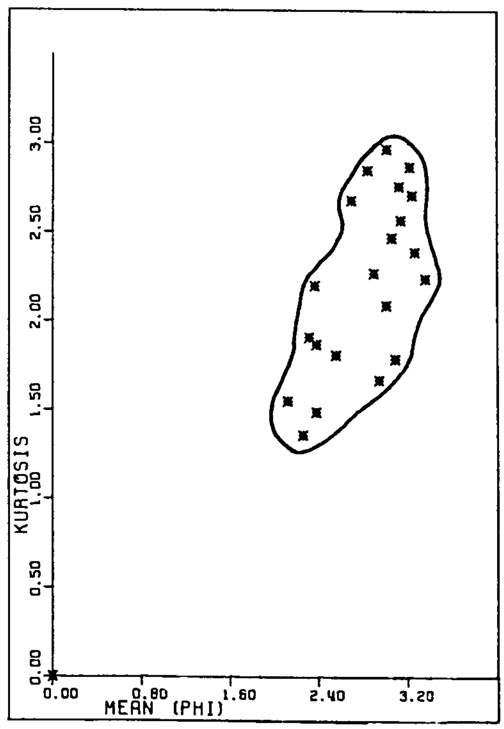
A


KURTOSIS
0.00
1.00
2.00
3.00
4.00
MEAN (PHI)
0.00
2.00
4.00
6.00
6.00

B


KURTOSIS
0.00
0.80
1.60
2.40
3.20
4.00
4.80
MEAN (PHI)
0.00
1.00
2.00
3.00
4.00

C


KURTOSIS
0.00
0.50
1.00
1.50
2.00
2.50
3.00
MEAN (PHI)
0.00
0.80
1.60
2.40
3.20

D

more rapid than current sorting processes which rework the sediments. Thus, the amount of time available for sediment reworking may be a controlling factor when abundant sand is available as in the case of fossil beach placers described by Houston and Murphy (1977).

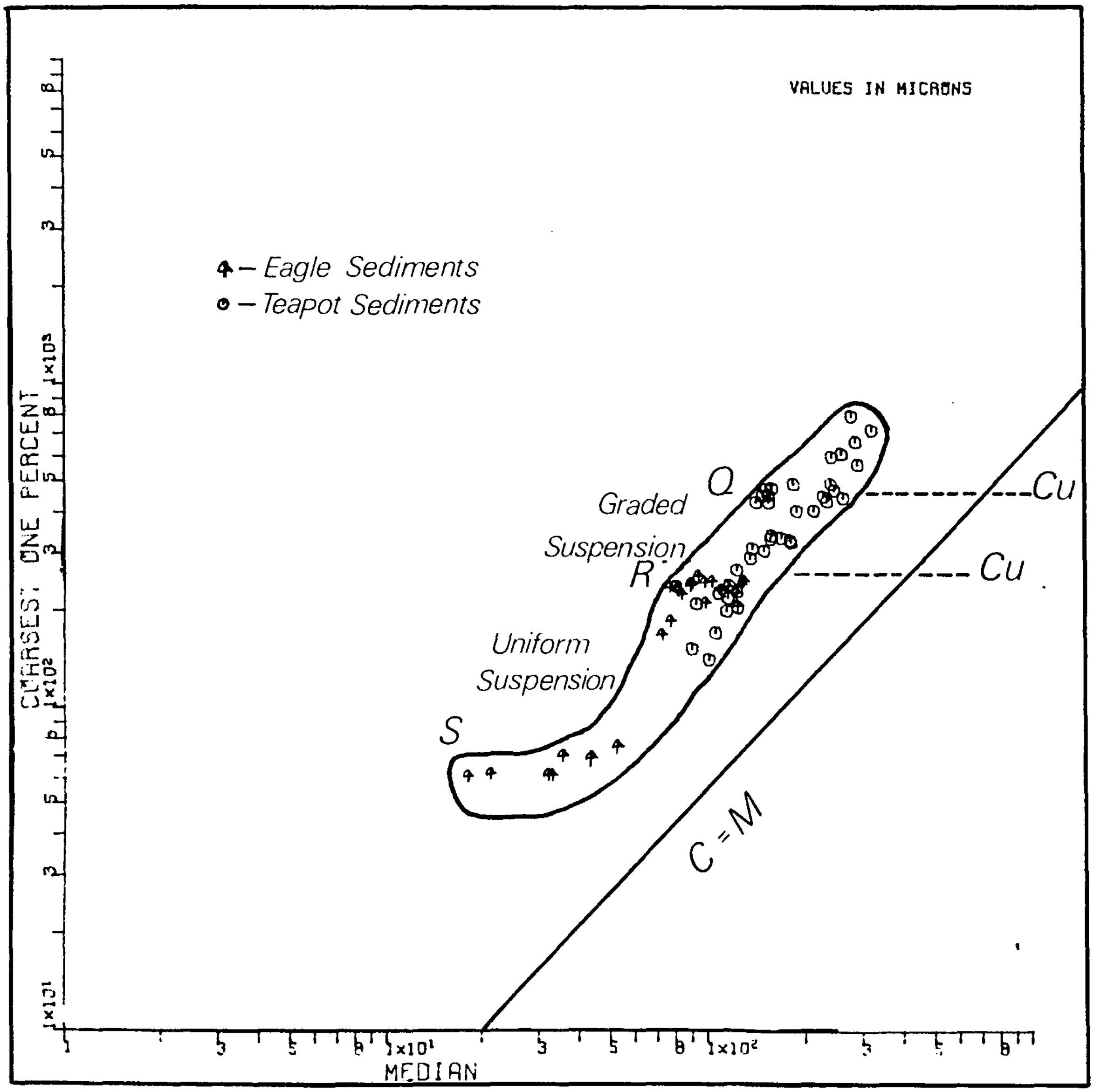
CM diagram

During the last 20 years, a wide variety of scatter diagrams have been used to characterize the various depositional agents operating within major sedimentary environmental complexes. Among the more useful is the CM diagram in which the mean grain diameter is compared graphically to the coarsest one percent of a given sediment population (Passega, 1957). The parameter CM is then, a measure of the competency of a given depositional agent reflecting the sensitivity of interaction between the coarsest size material available for transport and the energy of the depositional medium.

Passega (1957, 1964) and Passega and Byramjee (1969) have found that environments including alluvial fan, turbidity current, stream channel, fluvial and beach are characterized by certain graphic patterns which are indicative of the sedimentary processes operating within each environment.

A CM diagram has been constructed (fig. 9) displaying patterns indicative of Eagle beach sands and Teapot fluvial sediments comparable to those described by Passega (1957) and Passega and Byramjee (1969) for beach deposits and Williams and Rust (1969) for alluvial sediments. This diagram illustrates the application of CM diagrams in differentiation of depositional mediums as discussed by the above authors.

Figure 9. CM diagram, representing a comparison of coarsest one percent and median grain size for Eagle and Teapot sediments. Note that most Eagle sediments fall into segment R-S, the uniform suspension, while Teapot sands shown by line R-O were deposited under mostly graded suspension flow conditions.


VALUES IN MICRONS
– Eagle Sediments
– Teapot Sediments
COARSEST ONE PERCENT
MEDIAN
Q
R
S
Graded Suspension
Uniform Suspension
Cu
Cu
C = M
1x10¹
1x10²
1x10³

The Eagle formation contains fine-to very fine-grained sands and siltstones which were deposited in a beach-shelf transition and which represent deposition from a uniform suspension (fig. 9). As shown by segment R-S, Eagle formation sediments represent a continuum from higher energy bed load deposition (Passega, 1957) to lower energy suspension deposition with decreasing grain-size, reflecting a transition from higher energy foreshore to lower energy offshore beach environments. Teapot sandstones are fine to medium-grained and represent alluvial channel complexes deposited by both graded suspension and by rolling and bouncing along the channel bed (fig. 9 segment R-O).

Although CM diagrams are useful characterizations of depositional processes operating within sedimentary environments, they must be used only as one of several lines of evidence in paleoenvironmental reconstruction. Rock body geometry, primary structures, and other lithologic and paleontologic evidence should also be considered in making these interpretations.

Composition

Average sandstone compositions were determined with a petrographic microscope by systematic counting of 300 grains of representative thin sections and grain mounts and with x-ray diffractometry. Representative samples and associated clay mineral suites are shown in table 2 and are expressed as whole rock percentages.

Mesaverde sandstones may be classified (Folk and Ward, 1957) as litharenites and feldspathic litharenites, with quartz, chert, dolomite, and feldspars among the dominant mineral types (fig. 10).

Table 2. Quantitative x-ray data from lower Judith River and Teapot sandstones

Sample No.	Mixed Layer	Chlorite	Kaolinite	Illite	Monmorillonite
Lower Judith River Sands:					
1-40-3	44.02%	2.10%	16.90%	28.80%	8.18%
1-42-5	17.57%	6.90%	66.43%	4.87%	4.23%
1-44-1	00.00%	1.35%	79.65%	11.35%	10.92%
1-46-3	5.38%	4.03%	67.59%	11.35%	00.98%
1-46-4	16.05%	0.00%	78.12%	7.43%	00.98%
1-47-7	00.00%	18.00%	48.09%	34.42%	00.00%
Teapot Sands:					
1-53-1	00.00%	00.00%	80.56%	19.44%	00.60%
1-53-3	00.00%	1.00%	78.98%	18.12%	1.10%
1-53-4	16.41%	4.44%	76.42%	2.73%	00.00%
1-53-8	00.00%	3.97%	78.45%	18.34%	00.00%
1-53-9	00.00%	00.00%	75.21%	19.47%	00.00%
1-53-10	00.00%	1.06%	97.62%	00.00%	1.32%

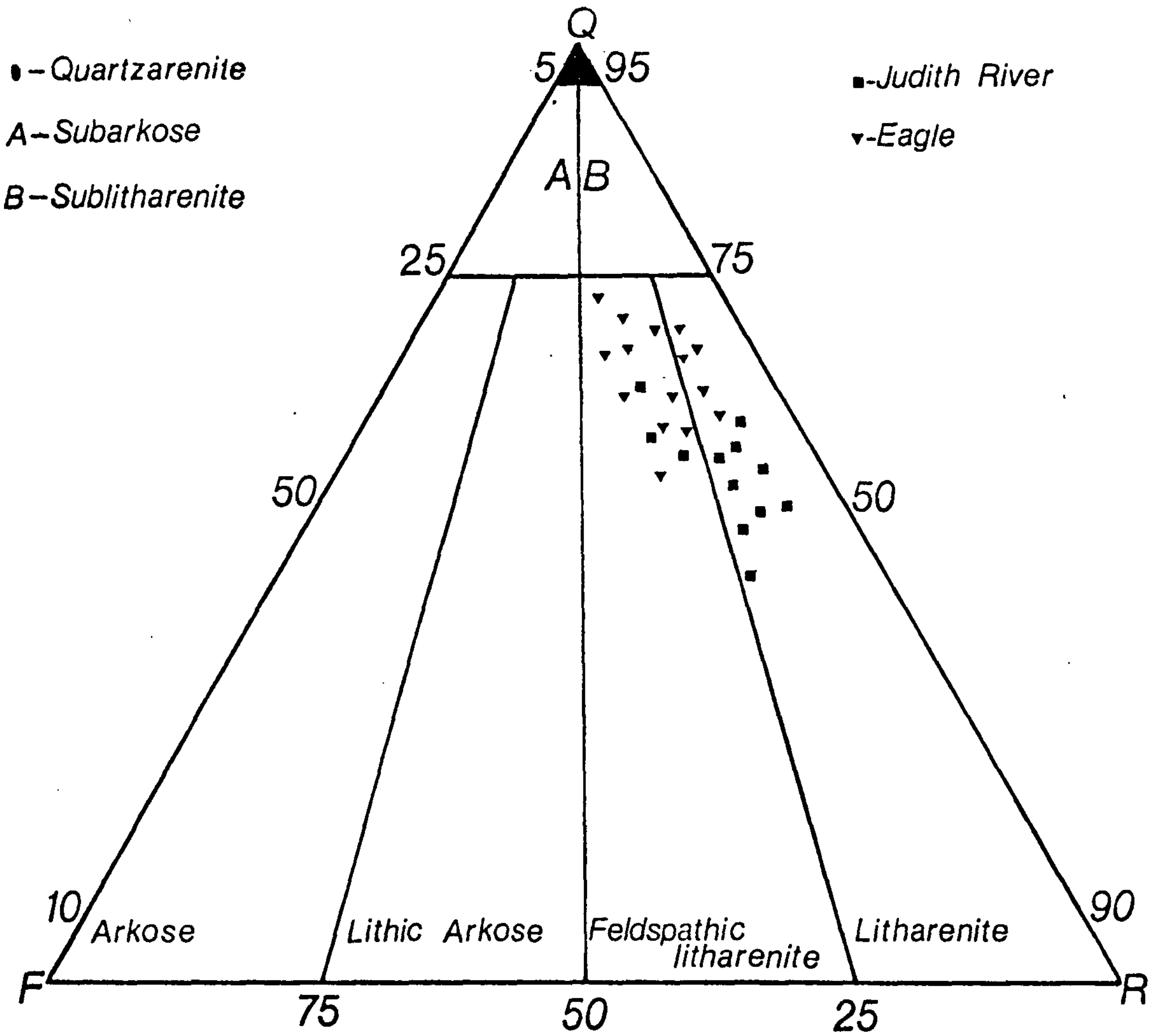


Figure 10. Compositional triangle displaying dominant lithology of Eagle and Judith River sediments. After Folk and Ward (1957).

Sandstones within the study area are homogeneous showing few mineralogical changes vertically or laterally. Over 95 percent of the total samples collected contain secondary sparry calcite cement (1-39%), quartz (28-57%), polycrystalline quartz (1-3%), chert (10-42%), plagioclase (0.3-3.7%), potassium feldspar (3-11%), glauconite (0.3-2.0%), and biotite (0.3-3.0%). Unstable mineral components such as pyroxenes, amphiboles, and rock fragments are absent in all rock samples studied. Feldspars and plagioclase grains are typically fresh with no evidence of alteration to clay mineral species.

Over 75 grain mounts and 30 thin sections of sandstones were examined to determine vertical and lateral changes in quartz-feldspar and feldspar-plagioclase ratios throughout the thesis area. These ratios indicate no trends in any major lithofacies, which is not surprising, as stratigraphic sections were taken roughly parallel to the inferred depositional strike of the Campanian shoreline. Distances away from the source areas as reflected by the constancy in mineral species percentages in the area of study apparently remained constant throughout deposition of the Mesaverde.

Eagle sandstones

Eagle sandstones can be differentiated from the majority of Judith River sands by the presence of various genetic types of dolomite and detritial glauconite. The presence of these mineral components throughout the Upper Cretaceous deposits in the Western Interior (Sabins, 1962, 1965, MacKenzie, 1975) has been documented during regional textural analysis. The importance of studying genetic types of dolomite lies in the fact that these minerals are restricted to offshore beach and shelf deposits and are absent in transitional marine and continental facies, thus providing another means of differentiating depositional environments preserved in the stratigraphic sequence.

In the northeastern portion of the Big Horn Basin, three genetic types of dolomite have been identified during thin section studies and include the following types previously described by Sabins (1962).

> Detrital dolomite: This type consists of allogenic fragments mechanically eroded and transported from areas outside of the basin of deposition, consisting of aggregates of well-rounded dolomite grains (plate 1B).
>
> Primary dolomite: These grains are formed from precipitation inside the basin of deposition, usually solitary rhombic crystals which have been abraded to various degrees displaying depositional fabrics (plate 1C).
>
> Secondary dolomite: Dolomite formed after the deposition of framework components composed of small euhedral rhombs, usually enclosed in a sparry clacite cement (plate 1D).

Detrital dolomite   Detrital dolomite grains (Sabins, 1962) as other types of dolomite are restricted exclusively to Eagle and the basal unit of the lower Judith River beach sediments. Detrital grains are rare, usually less than 5 percent of the total dolomite component in more sandstones. In all samples, individual detrital dolomite grains are similar in size to other framework elements suggesting deposition under similar hydrodynamic conditions.

Further evidence for allogenic origin is the association of detrital dolomite with other detrital grains as point or line contacts with no evidence of replacement by dolomite. The majority of the grains are well-rounded due to the relative softness of the mineral and are polycrystalline with no evidence of relict structures. Frequently, individual grains are coated with a thin limonite stain which is a striking feature in thin section.

Primary dolomite   Primary dolomite grains (Sabins, 1962) occur as single rhombic crystals with varying degrees of roundness developed during intrabasinal transport, as established by associated allogenic grain-boundary relationships. This type is the most abundant in all sandstone samples constituting over 50 percent of the total dolomite percentage. Thin section analysis reveals limonite staining along grain boundaries and cleavage fractures due to post depositional weathering of iron.

In all thin sections in which primary dolomite is present, the size of the grains is directly related to that of the associated detrital quartz grains. Primary dolomite grain size is plotted against frequency per thin section in figure 11. Average values are given due to samples plotting on the same point. The data shows that samples with coarser detrital quartz contain coarser dolomite grains suggesting that these two components were deposited by the same depositional medium under comparable energy conditions.

Primary dolomite grains display both point and line contacts with associated framework grains. These are clearly of the primary depositional type with no evidence of replacement of clastic grains by dolomite grains. In many samples, calcite cements detrital components,

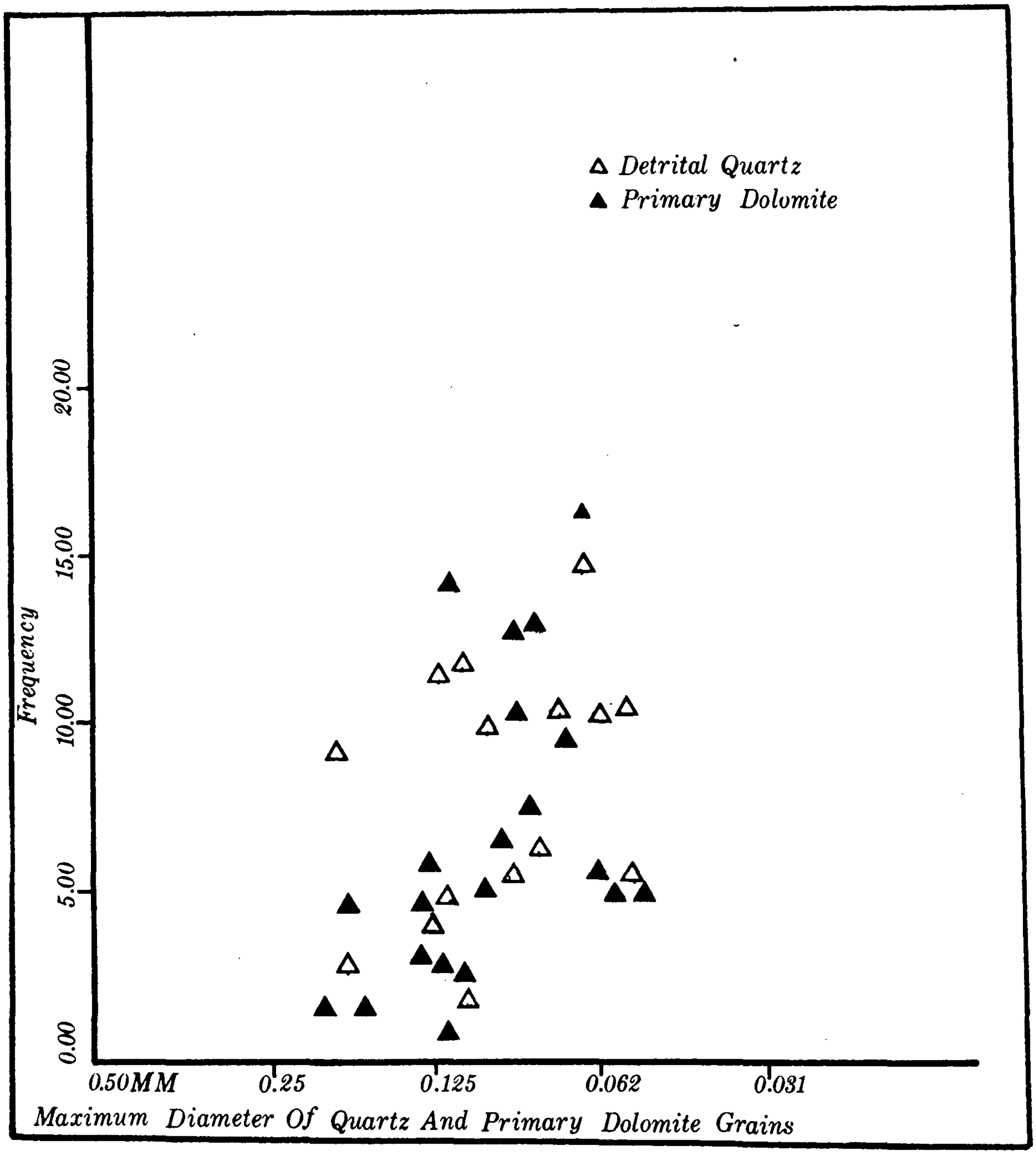


Figure 11. Scatter diagram displaying the direct relationship between mean grain size of quartz and primary dolomite grains versus their frequency.

poikolitically enclosing both primary dolomite and framework grains. Optically continuous patches of sparry calcite are not in the same optical orientation with any adjacent dolomite grains suggesting formation of calcite after the primary dolomite.

Secondary dolomite   Secondary dolomite (Sabins, 1962) always occurs as small euhedral crystals averaging 0.04-0.06mm in greatest diameter. These rhombs occur mainly in calcite cemented sandstones where secondary calcite fills void spaces between sand grains. Secondary dolomite rhombs are associated with large optically contigous patches of calcite over 1.0mm in diameter replacing calcite and sometimes embaying detrital quartz grains.

Grains of detrital, primary, and secondary dolomite are important mineralogical constituents of Eagle and lower Judith River sediments consisting of up to 29 percent of the average volume. Regional lithologic and biostratigraphic correlations suggest that the dolomite is restricted to late Cretaceous marine beach and shelf environments and is not associated with evaporite deposits (MacKenzie, 1975, Sabins, 1962). Greatest abundances of dolomite types are encountered in basinward portions of marine sandstones, while landward equivalents contain little or no dolomite. Dolomite formation is then environmentally controlled and identification of genetic types can provide valuable insight concerning environments of deposition.

Clay mineral composition and significance

Over one hundred and twenty-five samples of less than two micron sediment has been analyzed by the writer from the Mesaverde Group exposed near Dry Creek along with 25 samples of mudrock from the Claggett Formation to determine vertical and lateral variations in clay mineral content and whether or not these variations might be environmentally controlled. Standardization of all treatments was used in preparation of unorientated clay mounts to insure maximum x-ray diffraction peaks. Forty-eight samples were chosen for quantitative mineral identification. Methods for measuring peak size and area developed by Schultz (1962) were utilized. The results of representative sandstone samples are presented in table 2.

Eagle sandstones Kaolinite was a major constituent of all Eagle sands averaging over 70 percent of the total clay mineral content (fig. 12). Illitic clays ranged from 20 to 30 percent and consisted primarily of 9.8Å clays of the mica type. Chlorite and montmorillonite were found only as traces of the total clay mineral percentage in most samples. Upper Eagle sands contained 40 to 50 percent mixed-layer illite-montmorillonite not found elsewhere in the Eagle deposits. Percentages of expandable material in the mixed-layer lattice averaged approximately 30 percent (Weaver, 1956). Throughout the Eagle Formation no significant trends in kaolinite content were noted, although montmorillonite content increased slightly upward in the interval.

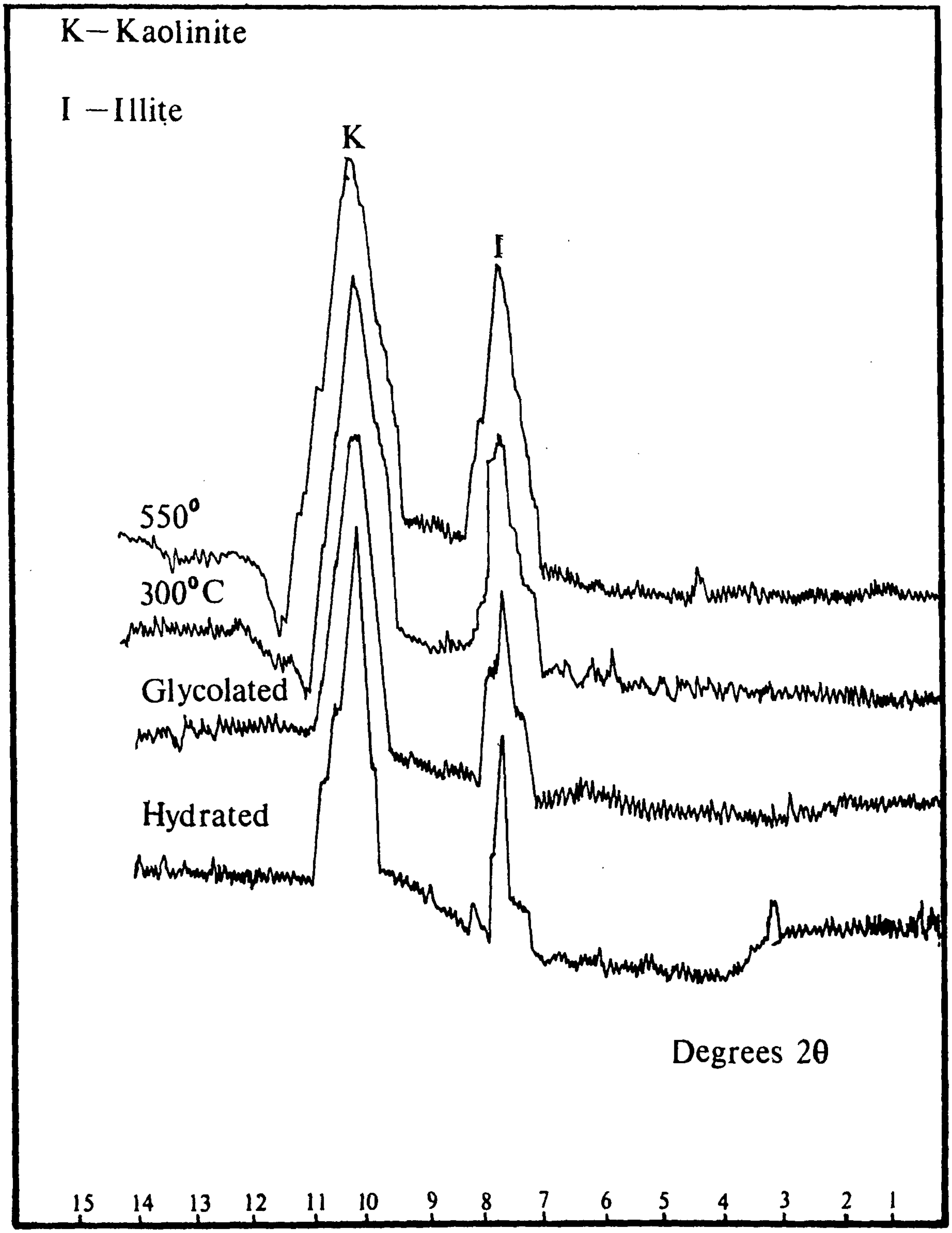


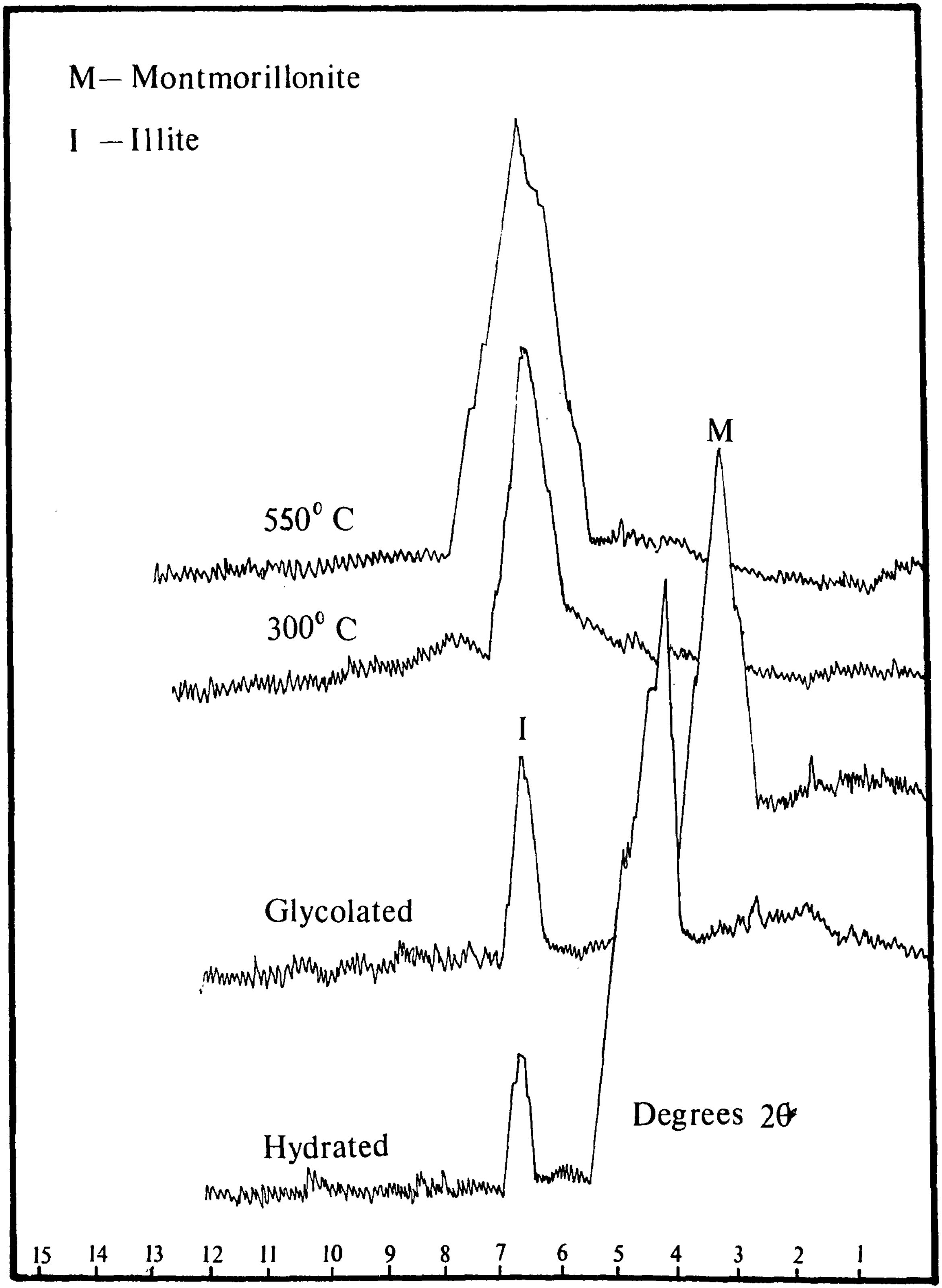
Figure 12. X-ray diffraction chart displaying prominent Illite and Kaolinite peaks in Eagle sands.

Claggett mudrocks Analysis of shale and mudstone samples from the Claggett Formation throughout the area of study indicate montmorillonite to be the major clay mineral constituent ranging from 32 to 55 percent (fig. 13). 10Å illite percentage ranges from 3 to 47 percent while kaolinite values averaged less than 10 percent. These values are comparable to results obtained by Weaver (1961) who found montmorillonite to be the most abundant clay mineral in marine shales of Mesaverde equivalents in the Washakie Basin, Wyoming.

A sandstone sample 60 feet from the base of the Claggett Formation at Dry Creek contained small, rounded clasts of bentonitic shale along with a thin, discontinuous layer of bentonite less than one inch thick. This layer was not found in other measured sections to the north but may be correlatable with the Ardmore Bentonite which occurs near the base of the Claggett Formation (Gill and Cobban, 1973, McGookey, 1972) in southern Montana and eastern Wyoming. Montmorillonite content in Claggett mudrocks increases from 32 percent near Dry Creek to almost 60 percent to the north near Lovell, Wyoming. This trend reflects increasing amounts of volcanoclastic debris deposited in the Claggett sea in the direction of the source area, the Elkhorn Volcanics in central Montana (Gill and Cobban, 1973, McGookey, 1972).

Judith River sandstones Clay mineral content in Judith River sands is very similar to that of the Eagle Formation. Kaolinite percentage is quite high ranging from 66 to 97 percent, while illite contents range from 3 to 28 percent. Montmorillonite and chlorite

Figure 13. X-ray diffraction chart displaying prominent Montmorillonite and Illite peaks in Claggett Mudrocks.



M– Montmorillonite
I –Illite
M
550° C
300° C
I
Glycolated
Hydrated
Degrees 2θ
15 14 13 12 11 10 9 8 7 6 5 4 3 2 1

are found only as traces and are never major clay mineral constituents in Judith River sands. Kaolinite percentage displays very little variation from the marine shelf sands near the base of the Formation to the fluvial Teapot sands. There is clearly very little correlation between kaolinite content and obvious environmental changes which are recorded throughout the sequence. This suggests that the sands showing very high kaolinite content may have been subject to diagenetic and post diagenetic groundwater alteration.

The results of the clay mineral analysis by the writer agree with those of Weaver (1961) who studied the clay mineralogy of Mesaverde equivalents in the Washakie Basin, Wyoming. Weaver's data from the Lewis and Almond Formations indicate a decrease in kaolinite content in sediments landward from the strandline.

Within the Lewis Shale, montmorillonite is most abundant in open marine sediments, while illite, chlorite, and mixed-layer clays tend to be slightly more abundant in shallow marine deposits and kaolinite more abundant in beach sandstones (Weaver, 1961).

This data is comparable to results obtained by the writer in that kaolinite which is dominant in upper Eagle shoreface sands is replaced by montmorillonite in Claggett offshore sediments. In general, a change from low to high kaolinite content may indicate a change from open marine to nearshore conditions. Thus, in a broad regressive-transgressive cycle, such as the Eagle-Claggett, the time of maximum transgression should be represented by minimum kaolinite content.

Authigenic clays in Teapot sandstones Petrographic and scanning electron microscope analysis indicate authigenic kaolinite is present in Teapot sandstones but absent in Eagle and Judith River sands. Authigenic kaolinite occurs as crystalline clots and vermicular pseudo-hexagonal plates filling pores and as pore linings coating framework grains. Wilson and Pittman (1977) have summarized clay mineral data from over 3,000 sandstones and established a number of useful criteria for distinguishing authigenic from allogenic clays. Many features attributed to authigenic clay mineral formation by Wilson and Pittman (1977) are comparable to the features observed in the study of the clays in Teapot sands. It must be pointed out however, that no single textural or compositional feature indicative of authigenic clays is in itself a positive indicator of the presence of authigenic clay minerals. A number of criteria established in a single sandstone sample usually is sufficient for positive identification.

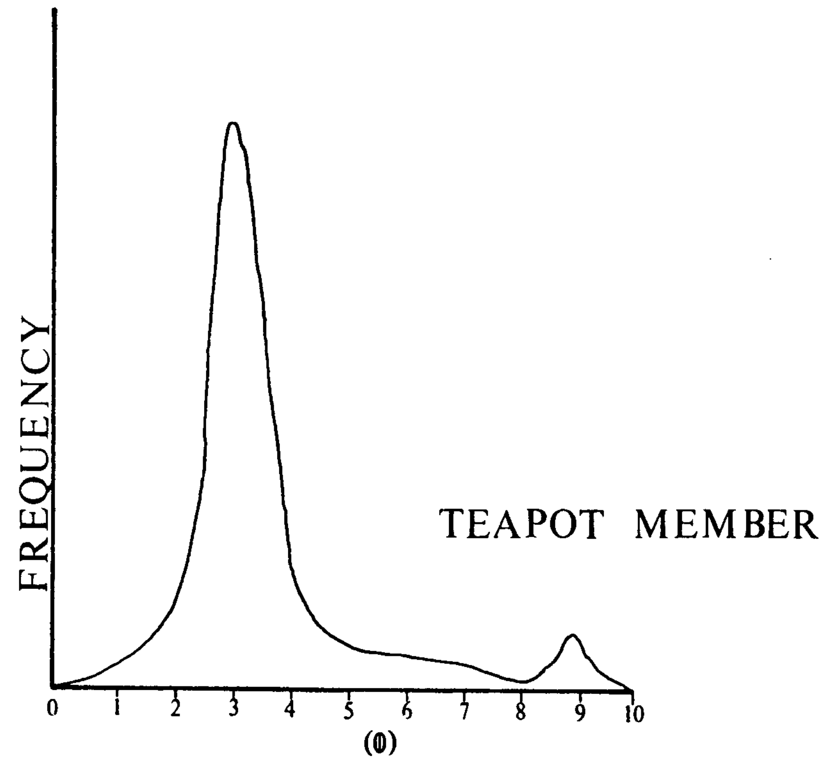
Many authigenic clay mineral suites are monomineralic reflecting their formation under a limited range of subsurface physical and chemical conditions (Wilson and Pittman, 1977). Teapot sands average over 80 percent kaolinite with many samples showing well over 90 percent of the total clay mineral assemblage to be kaolinite.

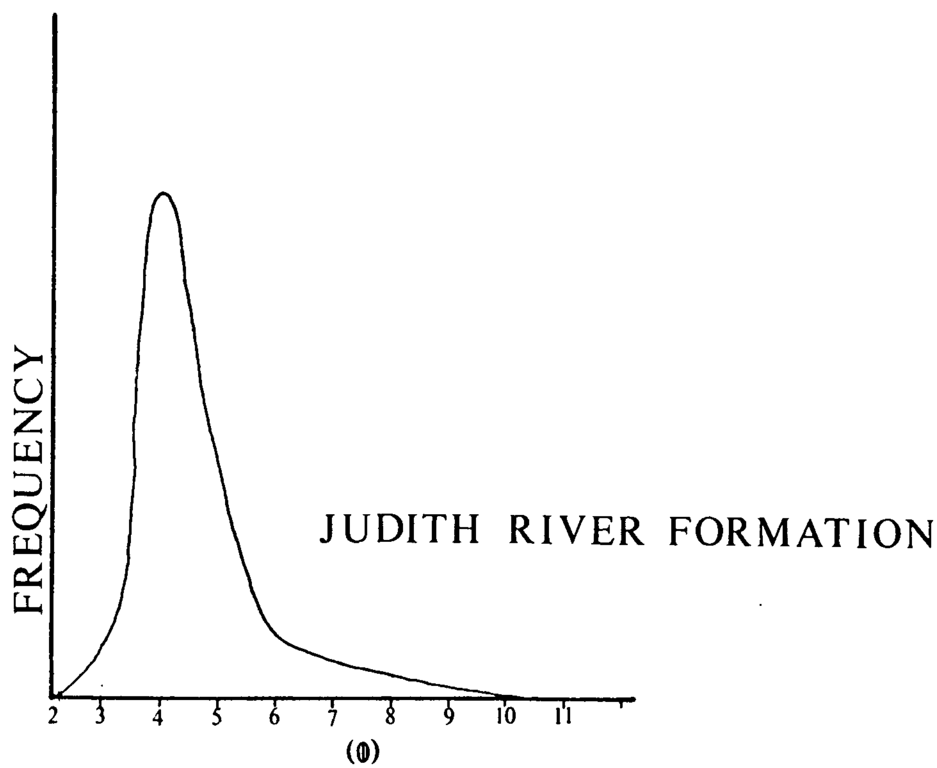
Authigenic clays in Teapot sandstones exhibit crystalline habits, the most common of which consists of a delicate vermicular growth comprised of a series of stacked pseudohexagonal plates sometimes filling entire pore spaces in sands (plates 2A through 2D). It is thought that the delicacy of this morphologic type precludes possibilities of extended periods of transport (Ross and Kerr, 1931, Wilson and Pittman, 1977). Authigenic kaolinite also fills pores as large crystalline clots which often obscure detrital grain boundaries.

Authigenic kaolinite also occurs as pore linings in sandstones forming thin coatings on grain surfaces and are absent at grain-to-grain contacts, a feature common to authigenic clays. As demonstrated by Shelton (1964), many authigenic clays obtain high degrees of crystallinity not common in allogenic clays. All teapot sandstones display sharp x-ray peaks as defined by peak height and area measures introduced by Schultz (1969).

Sandstones which contain authigenic clay minerals often exhibit distinct breaks in grain-size distributions (Shelton, 1964) resulting in bimodal size distributions. Comparisons of Teapot with Judith River sands (fig. 14) show silt size material is present in small amounts compared to clay in Teapot sands while Judith River sands display unimodal distributions. Bimodal distributions are the result of diagenetic growth of kaolinite and are not a consequence of hydrodynamic conditions present during the time of deposition. Textural parameters including sorting, skewness, and kurtosis used frequently as paleoenvironmental indicators, can be grossly misinterpreted if authigenic clays are mistaken for allogenic types. Interpretations based on sieve and settling tube data can be seriously in error if diagenetic processes have altered the textural properties of sandstones.

Figure 14. Comparison of grain size distribution in Teapot sands and Judith River sands.


FREQUENCY
TEAPOT MEMBER
0 1 2 3 4 5 6 7 8 9 10
(0)


FREQUENCY
JUDITH RIVER FORMATION
2 3 4 5 6 7 8 9 10 11
(0)

The presence of authigenic clays in sandstones including the Teapot are important in the petroleum industry as they may control porosity and permeability of reservoir sands. Permeability can be drastically reduced by pore lining clay which grows outward towards the center of void spaces from grain boundaries. Pore clots by authigenic clays greatly reduce reservoir quality and frequently cause production problems during drilling due to their ability to block pores.

STRATIGRAPHY

Development of Terminology

The term Mesaverde was first proposed by Holmes (1877, p. 245) in descriptions of sequences of massive sandstones and coal beds exposed as flat lying strata on the Mesaverde, Montezuma County, Colorado. Since initial work, "Mesaverde" has been used in both a rock and time stratigraphic sense and has been used extensively and loosely in descriptions of coal-bearing Upper Cretaceous strata exposed throughout the Western Interior. Because of the complex intertonguing relations involving marine and continental strata, there have been almost as many nomenclatural schemes as there have been students of the Mesaverde. This has been complicated by the difference in purpose between structural mapping, subsurface, and surface investigations. Many local and regional studies have adopted terminology much different from adjacent areas without regard to the fact that these rocks were initially deposited within the same basin consequently displaying many comparable lithologic, geometric, and paleontologic characteristics.

The term Montana Group was first introduced by Eldridge (1889) in descriptions of Upper Cretaceous strata in central Montana. The nomenclature used by Eldridge received wide acceptance and was adopted by Weed (1889), Stanton and Hatcher (1903) and Stanton et al. (1905) during their investigations of Upper Cretaceous rocks in central and southern Montana. Stanton and Hatcher (1903, p. 63) subdivided the Montana

Group into the following formations which are in ascending order: the Eagle Formation, Claggett Formation, Judith River, Bearpaw, and Fox Hills Formations.

In the Big Horn Basin (Fisher, 1908, Cobban and Reeside, 1952, Rea and Barlow, 1975, Gill and Cobban, 1973, Miller, Barlow, and Haun, 1965, Severn, 1961, Washburne and Woodruff, 1907, and MacKenzie, 1975) have recognized the subdivisions of the Montana Group, while most authors have retained the term Mesaverde in descriptions of strata superjacent to the Cody Formation and subjacent to the Meeteetse-Bearpaw equivalents. During stratigraphic and structural investigations throughout the last 15 years, the Eagle, Claggett, Judith River, and Teapot have been recognized and mapped in different parts of the Big Horn Basin, although the use of these terms has not been consistent even between adjacent areas. Differences in nomenclature are due largely to the fact that marine rocks contained within the Cody and Claggett Formations are largely absent in the southwestern portion of the basin making subdivisions of lithologically similar continental units rather tentative.

Throughout the last 20 years controversy has been the theme concerning the status of the Mesaverde as a group or formation in the Big Horn Basin. Severn (1961) was the first to describe general facies relationships of the Mesaverde and suggested that it be raised to group status with the Eagle, Claggett, and Judith River considered formations. Severn (1961) saw the units to be southward extensions of the type Montana Group and showed them easily traceable southward from southern Montana along the eastern Big Horn Basin outcrop belt. Also recognized by Severn (1961) was the Teapot Sandstone, the upper member of the Judith River Formation, correlatable northward from its type section in the Powder River Basin.

Miller, Barlow, and Haun (1965) restricted the term Mesaverde to the southwestern portion of the basin, westward of the point of maximum marine transgression where the entire section is considered continental. In eastern areas, these authors recognized the presence of the Eagle and Judith River Formations as representing clastic wedges extending eastward into dominately marine Cody and Claggett Formations.

Further confusion regarding terminology developed as Miller, Barlow, and Haun (1965) did not include the Eagle, Claggett, and Judith River within what they called the Mesaverde. MacKenzie (1975) in studies of the southern portion of the Big Horn Basin considered the Mesaverde group status with the Eagle, Claggett, Judith River Formations and the Teapot Sandstone Member laterally continuous, mappable units throughout the area.

In this report, the Mesaverde is considered a group with the Eagle, Claggett, and Judith River Formations and the Teapot Sandstone Member present as lithologically distinct and mappable units throughout the study area. Because the writer has studied only a local area, he will retain the term Mesaverde as previously applied in the Big Horn Basin. The term Montana Group would be most applicable while following guidelines established by the code of stratigraphic nomenclature. The names Eagle Sandstone and Claggett Shale should not be retained as these units were first named as formations at the type sections. Neither the Eagle or the Claggett Formations in the type areas consist entirely of sandstone and shale respectively, thus the use of these terms should be abandoned and replaced by the term Formation.

Stratigraphic Relations Within the Mesaverde Group

Stratigraphic relationships within the Mesaverde Group in the Big Horn Basin and in the study area are complex with intricate intertonguing between marine and continental strata occurring over very short distances. Regional stratigraphic cross-sections (Asquith, 1974, Gill and Cobban, 1973) show that Late Cretaceous shorelines prograded eastward as a series of pulses from constantly shifting deltaic depocenters separated by interdeltaic areas. Biostratigraphic and subsurface evidence indicates Campanian shorelines were complex with transgression, regression, and stillstand areas occurring in close proximity at the same time periods (Asquith, 1974, Gill and Cobban, 1973, Weimer, 1961, 1970). The relatively simple picture of sequences of time-equivalent, nonmarine pulses separated by similar marine tongues, may be in need of considerable revision (Asquith, 1974). As evidence for a series of coalescing deltas in the present Big Horn Basin area accumulates, there is an increasing need for a new look at the layer cake interpretations of facies changes and associated lithologic distributions within the Mesaverde Group.

Eagle Formation

Facies relationships within the Cody Shale and Eagle Formation are very complex within the study area. Rapid lateral facies changes and pinchouts involving hundreds of feet of offshore marine shales and littoral sandstones often occur in lateral distances of less than five miles, even through the outcrop trend along the eastern outcrop belt is inferred to be approximately parallel to the depositional strike (Houston and Murphy, 1977, Gill and Cobban, 1973, and Severn, 1961) (fig. 15). Lateral pinchouts involving over 300 feet of Upper Cody and Lower Eagle sediments occurs over distances of less than six miles.

Throughout the Eagle, two laterally persistent northwestward thickening tongues of the upper part of the Cody Shale occur exhibiting very complex, interfingering relationships and rapid lateral pinchouts with littoral Eagle sediments (fig. 15). The lowermost upper Cody tongue which is not present at Dry Creek is 80 feet thick at Little Dry Creek only 1.5 miles away. This tongue interfingers with the lowermost Eagle sand unit west of Alkali Anticline and becomes part of the main body of the Cody near Lovell. The uppermost shale tongue of the upper part of the Cody thickens northwestward from 70 feet at Dry Creek to over 200 feet near the south nose of Alkali Anticline.

As with the lower tongue, interfingering relationships with the Eagle sands are inferred by pronounced thickening and thinning throughout the outcrop area. Intertonguing relationships between the Cody and Eagle Formations have been noted in the area west of Alkali Anticline

MEASURED SURFACE SECTIONS–MESAVERDE GROUP

SE–NW OUTCROP BELT

GREYBULL TO LOVELL

BIG HORN BASIN

BIG HORN COUNTY, WYOMING

VERTICAL SCALE

0 75 150 FEET

HORIZONTAL SCALE

0 1.5 3 4.5 6 MILES

EXPLANATION

SHALE

MUDSTONE

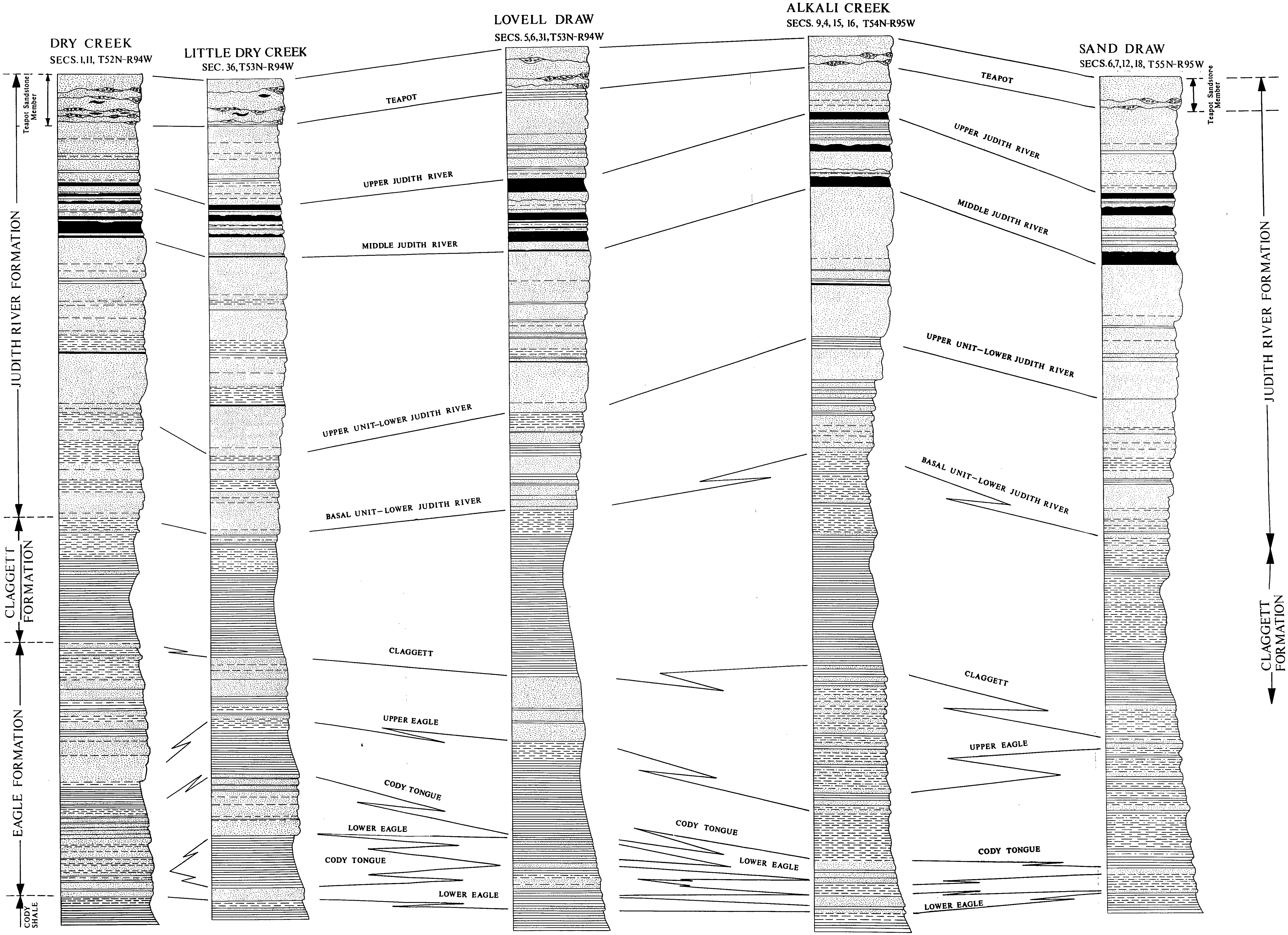
CARBONACEOUS MUDSTONE

SILTSTONE

SANDSTONE

IRONSTONE CONCRETION LENSES

Figure 15. Graphic sections.



DRY CREEK
SECS. 1,11, T52N-R94W
LITTLE DRY CREEK
SEC. 36, T53N-R94W
LOVELL DRAW
SECS. 5,6,31, T53N-R94W
ALKALI CREEK
SECS. 9,4,15,16, T54N-R95W
SAND DRAW
SECS. 6,7,12,18, T55N-R95W
Teapot Sandstone Member
JUDITH RIVER FORMATION
CLAGGETT FORMATION
EAGLE FORMATION
CODY SHALE
TEAPOT
UPPER JUDITH RIVER
MIDDLE JUDITH RIVER
UPPER UNIT-LOWER JUDITH RIVER
BASAL UNIT-LOWER JUDITH RIVER
CLAGGETT
UPPER EAGLE
CODY TONGUE
LOWER EAGLE

by Cobban (1969) and by Rea and Barlow (1975) and have been interpreted is being associated with an easterly dipping shelf-slope transition zone with a paleoslope of about one half degree which may be associated with prodeltaic slopes (Asquith, 1974) and MacKenzie (1975).

The Eagle Formation in the Big Horn Basin consists of two eastward thinning clastic wedges separated by shale tongues of the Cody Formation. The lowermost regressive units, the Virgille and Telegraph Creek equivalents (Gill and Cobban, 1973) consist of massive beach and nearshore sands and are restricted to the northwestern portion of the basin. The uppermost tongue of the Eagle is present in much of the Big Horn Basin. It is represented by environments grading eastward from coastal and deltaic plain to transitional marine sediments and finally to prodelta and offshore marine deposits near the southeastern outcrop belt (MacKenzie, 1975).

The Eagle Formation is a mappable unit everywhere in the basin where it occurs subjacent to marine shales of the Claggett Formation (Miller, Barlow, and Haun, 1965). In southwestern areas, where the Claggett is absent or present only as littoral sands and silts, the Eagle and Judith River are entirely continental and are not easily differentiated. Thicknesses of over 1,000 feet of transitional marine and continental strata near Cottonwood Creek in T.45N., R.96W, grade laterally into littoral sands and offshore marine or prodeltaic shales to the east, thinning to a feather-edge at Mud Creek in T.44N., R.91W. (Severn, 1961, MacKenzie, 1975).

Comparisons of stratigraphic sections along the northwest-southeast outcrop belt measured by MacKenzie (1975) and the writer indicate pinchouts in the Eagle near the Basin-Manderson area. Stratigraphic sections measured near Manderson, T.50N., R.92W, show no Eagle present (MacKenzie, 1975), while measurements only 12 miles north at Dry Creek indicate nearly 400 feet of Eagle sediments. Similar relationships involving pinchouts between littoral sands and offshore marine shales along the depositional strike are common in the Almond Formation and the Lewis Shale near the Rock Springs uplift in southwestern Wyoming. These areas are interpreted by Weimer (1966, 1970, 1975) as reversals in the normal eastward sandstone pinchouts and are indicative of strong inflections or embayments in the shoreline associated with deltaic depocenters.

Claggett Formation

The Claggett Formation is present throughout the eastern two-thirds of the Big Horn Basin as westward extending tongues of the upper Cody marine shales. Time equivalent strata in the western portion of the basin are represented by transitional marine and continental sediments. In the basin, the distribution of marine shales within the Claggett indicates the extent of marine transgression. Claggett strata are variable in thickness especially near Nieber anticline, possibly reflecting subtle influences of deformation during deposition (Miller, Barlow, and Haun, 1965). Although local areas of thinning do occur, the Claggett interval thins westward pinching out near the west central portions of the basin. Recent field work by

MacKenzie (1975) indicates the Claggett Formation is present as far west as Cottonwood Creek in the southwestern portion of the basin. Within the study area the Claggett is present as southeastward-projecting tongues of the Cody Shale, thinning from 325 feet near Lovell to 190 feet at Dry Creek near Greybull.

Judith River Formation

Judith River sediments are present throughout the Big Horn Basin consisting of a complete regressive-transgressive cycle in eastward outcrops grading westward into coastal or deltaic plain sediments. Individual sandstone beds near the base of the Judith River grade eastward into marine shales, defining the base of the unit as a series of step-ups with sediments gradually becoming younger in the direction of regression (Gill and Cobban, 1973, Miller, Barlow, and Haun, 1965).

The Judith River Formation in the Big Horn Basin averages 700 to 1,000 feet in southwestern areas, thinning eastward to near 400 feet along the extreme southeastern outcrops (MacKenzie, 1975). Detailed stratigraphic relationships have not been examined elsewhere in the basin. This is especially true in the extreme west and northwest areas where the Mesaverde is still largely unknown. Only general facies relationships between marine and continental strata are understood in the southern parts of the basin (Gill and Cobban, 1966, 1973, MacKenzie, 1975) while very little work with regards to paleoenvironments has been done.

Along the northeastern outcrop belt from Greybull to Lovell, the Judith River averages over 700 feet thick thinning slightly northwestward from 750 feet thick at Dry Creek to near 700 feet near Lovell. Throughout the area no significant facies changes or pinchouts occur within the Judith River with many sandstone bodies easily traceable laterally indicating the outcrop trend is approximately parallel to the depositional strike (Houston and Murphy, 1977). Individual units exhibit almost no variations in thickness or lithology, an ideal setting for the study of stratigraphic changes occurring along-strike, which contrast with the rapid lateral pinchouts in the interval perpendicular to the trend of the shore.

Age and Correlation

Eagle sediments in southern Montana and northern Wyoming are Early Campanian in age having been deposited during the range of *Desmoscaphites* through *Baculites* sp. smooth variety (Gill and Cobban, 1973). The Telegraph Creek-Eagle regression began about 85 my BP in Montana and Wyoming during the range of *Scaphites depressus* (Gill and Cobban, 1966, and 1973). In the eastern portion of the Big Horn Basin, deposition of the Eagle was not initiated until about 82 my BP during the range of *Scaphites hippocrepus II* (Gill and Cobban, 1973). Within the study area the base of the Eagle Formation is dated at about 82 my BP by the presences of *Scaphites hippocrepus II* (Cobban personal communication, 1978). The Eagle interval in the study area is laterally equivalent to the Virgille Sandstone and the Telegraph Creek Formation in the northwestern portions of the Big Horn Basin (fig. 16).

The Claggett transgression in the Big Horn Basin began during the ranges of Baculites obtusis and Baculites nclearni, about 80 my BP, at which time the strandline moved westward as much as 150 miles (Gill and Cobban, 1973). Baculites obtusis, early form (Cobban, personal communication, 1978) collected from concretions near the Claggett, Judith River Contact, provide a date of about 80 my BP for this zone. The Claggett Formation is characterized by numerous bentonite beds, the most important of which is the Ardmore Bentonite, occurring in the basal part of the unit in Montana. In the southern Powder River Basins, the Ardmore Bentonite directly overlies the Sussex Sandstone Member of the Cody Formation. The Claggett Formation in the study area is equivalent in age to the transitional marine and fluviatile beds in the Judith River Formation immediately subjacent to the Teapot Sandstone Member near the western margin of the Big Horn Basin (Gill and Cobban, 1973).

Rocks deposited during the Judith River regressive phase are considered Late Campanian. They were deposited during the range zones of Baculites perplexus and Baculites scotti, about 78.5 to 75.5 my BP (Gill and Cobban, 1973). During this time period the strandline regressed eastward about 190 miles and is recorded as a series of step-ups in the base of the Judith River (Gill and Cobban, 1973, Miller, Barlow, and Haun, 1965). The tripartite subdivision of the Judith River recognized in the eastern Big Horn Basin (Miller, Barlow, and

Figure 16. Correlation chart of Mesaverde equivalents in the Big Horn and Western Powder River Basins along with ammonite zonation.

<table>
<tr><th colspan="3">Western interior ammonite sequence</th><th>COTTONWOOD CREEK</th><th>ZIMMERMAN BUTTE</th><th>STUDY AREA</th><th>NOWATER CREEK</th><th>SALT CREEK, POWDER R. BASIN</th></tr>
<tr><td rowspan="6">SYSTEM</td><td rowspan="6">SERIES</td><td>Discoscaphites nebrascensis</td><td rowspan="8">MEETEETSE</td><td rowspan="8">MEETEETSE</td><td rowspan="8">MEETEETSE</td><td rowspan="8">BEARPAW ??</td><td rowspan="8">LEWIS SHALE</td></tr>
<tr><td>Discoscaphites roanensis</td></tr>
<tr><td>Sphenodiscus (Coahuilites)</td></tr>
<tr><td>Baculites clinolobatus 24</td></tr>
<tr><td>Baculites grandis 23</td></tr>
<tr><td>Baculites baculus 22</td></tr>
<tr><td rowspan="23">CRETACEOUS</td><td rowspan="17">UPPER CAMPANIAN</td><td>Baculites eliasi 21</td></tr>
<tr><td>Baculites jenseni 20</td></tr>
<tr><td>Baculites reesidei 19</td><td>TEAPOT</td><td rowspan="3">TEAPOT MBR.</td><td rowspan="2">TEAPOT</td><td rowspan="2">TEAPOT</td><td rowspan="5">TEAPOT MBR.</td></tr>
<tr><td>Baculites cuneatus 18</td><td rowspan="14">JUDITH RIVER</td></tr>
<tr><td>Baculites compressus 17</td><td rowspan="14">JUDITH RIVER</td><td rowspan="12">JUDITH RIVER</td></tr>
<tr><td>Didymoceras cheyennense 16</td><td rowspan="12">JUDITH RIVER</td></tr>
<tr><td>Exiteloceras jenneyi 15</td></tr>
<tr><td>Didymoceras stevensoni 14</td><td rowspan="2">UNNAMED MEMBER</td></tr>
<tr><td>Didymoceras nebrascense 13</td></tr>
<tr><td>Baculites scotti 12</td><td rowspan="2">PARKMAN</td></tr>
<tr><td>Baculites gregoryensis 11</td></tr>
<tr><td>Baculites perplexus (late form)</td><td rowspan="2">CODY TONGUE</td></tr>
<tr><td>Baculites gilberti 10</td></tr>
<tr><td>Baculites perplexus (early form)</td><td rowspan="10">STEELE SHALE</td></tr>
<tr><td>Baculites sp. (smooth)</td></tr>
<tr><td>Baculites asperiformis 9</td></tr>
<tr><td>Baculites mclearni 8</td><td rowspan="7">CODY</td></tr>
<tr><td rowspan="6">LOWER</td><td>Baculites obtusus 7</td><td>CLAGGETT</td><td rowspan="2">CLAGGETT</td></tr>
<tr><td>Baculites sp. (weak flank ribs) 6</td><td rowspan="4">EAGLE</td><td>CLAGGETT</td></tr>
<tr><td>Baculites sp. (smooth) 5</td><td>EAGLE</td><td rowspan="2">EAGLE</td></tr>
<tr><td>Scaphites hippocrepis III</td><td rowspan="3">CODY</td></tr>
<tr><td>Scaphites hippocrepis II 4</td><td rowspan="2">CODY</td></tr>
<tr><td>Scaphites hippocrepis I 3</td><td>CODY</td></tr>
</table>

Haun, 1965) has been traced eastward and southward into the southern Powder River Basin (Rich, 1958). In this area, the Mesaverde Formation consists of the Parkman Sandstone, a middle unnamed marine unit and the Teapot sandstone (fig. 16).

Eagle Formation

The Eagle Formation was first named by W. H. Weed (1889) in descriptions of interbedded sandstones, shales, and lignites exposed as bluffs near the mouth of Eagle Creek, 30 miles east of Fort Benton, Montana. The formation is the basal unit of the Mesaverde Group consisting of cyclic sequences of sandstones, siltstones, and shales. The Eagle Formation is conformably underlain and interfingers with marine units of the Cody Shale. In the research area the unit ranges from over 410 feet near Greybull thinning northwestward to a feather edge about six miles south of Lovell, Wyoming.

The contact between the Eagle and underlying Cody Shale is always gradational and is marked by the first massive, ridge forming sandstone or the first thick sequence of cyclic sandstones, siltstones, and shales. This ridge trends roughly parallel to the margins of the Big Horn Basin and forms the first conspicuous hogback or cuesta adjacent to strike valleys formed in poorly indurated Cody shales. The upper contact of the formation is placed at the top of thin sandstones immediately below thick grayish brown siltstones and shales of the Claggett Formation.

Sandstone bodies contained within the Eagle interval are laterally contiguous, cyclic units ranging from five to 35 feet thick and are separated by marine shales and siltstones less than 10 feet to over 200 feet thick. Rea and Barlow (1975) report comparable sand unit thicknesses in areas adjacent to the study area. In outcrop, the number and thickness of discrete sand bodies decreases systematically northwestward reflecting sand-shale pinchouts.

Detailed stratigraphic sections of the Eagle Formation indicate the existence of laterally persistent lower and upper sandstone bodies, separated by tongues of the Cody Shale. These units are herein informally referred to as the lower and upper Eagle sandstones, terms used strictly in a descriptive sense only in the area covered by this report. It should be stressed that these terms are probably not applicable to all areas of the Big Horn Basin and have no formal stratigraphic rank.

Lower Eagle sandstone

The lower Eagle sandstone is very well-exposed at Little Dry Creek near Greybull and consists of over 100 feet of massive light grey to buff sands (plate 3A) with thin shale and mudstone interbeds. These sands pinch out to a feather edge near Lovell and are replaced by marine siltstones and shales. Individual sand units are quite continuous laterally and easily traceable in the field over distances of several miles. These units within the lower Eagle display sharp, erosional basal contacts in lower portions of the section grading upward to transitional contacts with sands decreasing in abundance upward.

Near the south nose of Alkali Anticline, massive sandstones are conspicuously absent and are replaced by thick sequences of coarsening upward shales and sandy silts. Basal contacts of individual sand beds, many of which are less than one foot thick, range from erosional to transitional, a reflection of discontinuous and uneven migration of the sand-mud transition zone, a feature common to beach shoreface-offshore transition zones (Masters, 1965).

Upper Eagle sandstone

The upper Eagle sand unit at Dry and Little Dry Creek consists of over 190 feet of massive sandstones, interbedded with thin siltstones and shales. This unit pinches out and interfingers with upper Cody shales in outcrops about six miles south of Lovell, Wyoming. Individual sand bodies are discontinuous and lenticular in contrast to the majority of the Mesaverde. Basal contacts of sand units are characterized by seaward dipping erosional surfaces (plate 3B) and concentrations of unorientated rip-up clasts, mudstone lenses, and clay galls, eroded from subjacent fine-grained beds.

Northward, at Alkali Creek, the upper Eagle is thick, repetitive sequences of shale, siltstones, and sandstones. Sands average only one to three feet thick, while silt and shale units range from 10 to 20 feet thick throughout this interval. Basal contacts of sand bodies range from erosional to gradational, again reflecting multiple, discontinuous strandline migrations throughout the northeastern Big Horn Basin area.

Claggett Formation

The Claggett Formation was named by Stanton and Hatcher (1903) for exposures of dark marine shale overlying the Eagle Formation near the confluence of the Judith and Missouri Rivers near Fort Claggett, Montana. Along the northeastern part of the Big Horn Basin outcrop belt, the Claggett is dominated by siltstones, silty mudstones, mudstones, and shales conformably overlain and interfingering with beach and shelf sediments of the Judith River Formation.

In the research area, the base of the Claggett Formation is placed at the first thick sequence of siltstones, mudstones, or shales immediately overlying thin bedded sandstones in the Eagle Formation (plate 3C). In most areas this contact is sharp, reflecting rapid rate transgression accompanying initial Claggett deposition (Gill and Cobban, 1973). The contact of the Claggett with the overlying Judith River Formation is similar to the Cody-Eagle zone in that both contacts are gradational. Regressive beach and shelf sandstones of the Judith River grade downward into upper Claggett laminated shales, mudstones, and siltstones. Near Alkali Creek, this gradational coarsening upward sequence is well-developed over an interval of 100 feet as Judith River beach sediments are underlain by thick sequences of shales, mudstones and sandy siltstones.

The Claggett Formation is very poorly indurated throughout the entire outcrop area underlying a conspicuous strike valley. These sediments are easily weathered and are buried by several inches to four feet of regolith, with exposures restricted to lower portions of hogbacks upheld by the basal Judith River sands. Bedding within the Claggett interval is typically homogeneous and very thinly bedded to laminated. Cross bedding is conspicuously absent which may be due in part to the general textural homogeneity of mudrocks contained within the Claggett.

Claggett sediments in the Big Horn Basin are considered exclusively marine. At Simmerman Butte, T.44N., R.93W, Gill and Cobban (1966, 1969) have collected B. Nclearne and Baculites obtusis, while MacKenzie (1975) has collected marine pelyocepods and foraminifera (Haplophragnoidea) from the Claggett.

Judith River Formation

The name Judith River was first used by Meek and Hayden (1856, p. 267) in discussions of exposures of nonmarine strata then considered Early Tertiary lake deposits. No exact type locality was designated during the Meek and Hayden surveys but was presumed to be "along the Missouri River between the Judith River on the west and the Musselshell River on the east". Until 1903, the term Judith River was used for strata of differing ages and stratigraphic position and various authors referred to the unit as the Judith River beds or Judith River Group. Stanton and Hatcher (1903) divided the Montana Group into four formations one of which they designated the Judith River Formation which they described as "consisting of 500-600 feet of light colored, nonmarine beds, underlying the Bearpaw Shale and overlying the Claggett Formation."

Throughout the study area, the lower contact of the Judith River Formation is placed at the base of the first thick sandstone or the first cyclic sequence of shales, siltstones, and sands overlying the Claggett Formation. This contact is quite similar to the Cody-Eagle contact zone where lighter colored hogbacks and cuestas are upheld by relatively resistant Judith River sandstones.

The upper contact is placed at the top of the Teapot Sandstone Member subjacent to bentonitic, carbonaceous shales and sandstones of the basal Meeteetse Formation (plate 3D). The Teapot was named by Barnett (1913) for exposures of buff sandstone and carbonaceous shale and coal beds near the Salt Creek oil field in the Powder River Basin. The Teapot is a conspicuous ridge former, and upholds the last resistant hogback or cuesta adjacent to large strike valleys eroded in poorly indurated overlying Meeteetse strata.

Throughout the area covered by this report, the Judith River Formation can be subdivided into four distinct units by the writer, each possessing unique qualities with respect to lithology and primary sedimentary structures. These are in ascending order as follows: The lower, middle and upper Judith River, overlain by the Teapot Sandstone Member. Although these units are easily mappable in the study area, they are not considered formal stratigraphic units. The exception being the Teapot Sandstone Member, which is a recognizable unit in the southern and eastern portions of the Big Horn Basin (MacKenzie, 1975). Miller, Barlow, and Haun (1965) have also recognized a tripartite subdivision of the Judith River into lower and upper sandstones, and a middle continental unit. These units have been correlated with the Parkman-Teapot sequence in the Wind River and the Powder River Basins Rich, 1958. MacKenzie (1975) was not able to subdivide the Judith River in southwestern areas of the basin, although his descriptions of basal sandstones along the southeastern outcrops show lithologic and bedding sequences comparable to the lower Judith River as described by the writer.

Lower Judith River

The lower Judith River conformably overlies the Claggett Formation and can be divided into two units. These bodies are both part of an overall progradational, coarsening-upwards shoreline sequence.

The basal unit of the lower Judith River averages 150 feet with no significant pinchouts or facies changes throughout the area of study. Well-developed, cyclic, coarsening-upwards sequences of sandy shale, siltstone, and yellowish gray sandstone are repeated over intervals as much as 50 feet. Sandstones are lithologically similar to Eagle sands, although a distinct increase in carbonaceous content is apparent. This material although not abundant, is present in greater amounts in southeastern sections at Dry and Little Dry Creeks reflecting current dispersal from areas to the south and west, where coals and carbonaceous strata are abundant in the Judith River (MacKenzie, 1975).

The upper unit of the lower Judith River averages nearly 240 feet, and consists of two massive, buff sandstones, ranging in thickness from

65 to 95 feet, separated by thick sequences of sandstones averaging 30 feet in thickness, interbedded with thin mudstones and shales. The basal contact is transitional as thin shales in the lower unit gradually give way to thick, massive sandstones. All sandstone bodies within this interval are laterally extensive, sheet-like sands, easily traceable in outcrop and on areal photographs.

Individual sand units are extremely massive and lithologically very homogeneous, forming conspicuous resistant hogbacks and cuestas (plate 4A). Each unit gradually coarsens upwards, and while devoid of carbonaceous material internally, are always overlain by one to three feet of carbonaceous mudstones or shales. The uppermost massive sandstones are overlain by thick sequences of carbonaceous strata contained in the middle Judith River section.

The general homogeneity of sand bodies within the upper unit is interrupted by large, spherical to oblate ferriginous sandstone concretions, which become very abundant in upper portions of sandstone units (plate 4B). Colored various shades of reddish brown and yellow, the concretions average three to five feet in longest diameter, with their long axis parallel to the strike of the bedding.

Grain-size within the concretions is much coarser than the host rock, due to the presence of sparry calcite cement. These concretions are also very abundant near the upper contact of the Teapot, and form impressive dip slopes in many areas. MacKenzie (1975) has reported identical concretions in the Judith River, and ascribes their origin in part, to diagenetic formation of iron oxide compounds around organic nuclei. Fairbridge (1967) indicates that these types of sandstone concretions may be formed during the epidiagenetic phase in which tectonic uplift and subsequent erosion occur, causing pyrite and marcasite to alter to limonite and hematite, frequently forming the cementing agent in sandstones.

The upper unit of the lower Judith River Formation contains two deposits of titaniferous black sandstone, characterized by unusually high concentrations of magnetite and ilmenite. These features have been described by Houston and Murphy (1962, 1977) as titaniferous black sandstones or fossil beach placers. In the study area, the northernmost deposit has been documented by Houston and Murphy (1962) and is located about six miles south, southeast of Lovell, and crosses the township line between Section 7, T.55N., R.95W., and Section 12, T.55N., R.96W. A second, undocumented deposit located by the writer is located three miles southeast of the Lovell deposit in the NE 1/4 of Section 30, T.55N., R.95W., at the same stratigraphic level, and may have been part of a single extensive beach deposit.

The Lovell deposit is easily recognized in outcrop as an imposing, dark colored cliff, standing in sharp contrast to underlying buff host sands. This deposit is composed of two erosional remnants, exposed as broad dip slopes, with the southern remnant over 3,000 feet long, in a northwesterly direction, and the total length of the deposit exceeding 5,000 feet (plates 4E and D). The black, titaniferous sand averages three feet thick, thickening to four feet near the northernmost extremity. The long axis of the Lovell deposit as now exposed trends N.45-50 degrees west. The original shape of the deposit cannot be determined, but is approximately parallel to regional strandline trends in Wyoming during Judith River Time (Zapp and Cobban, 1960, Gill and Cobban, 1973, Houston and Murphy, 1977).

The southernmost black sand deposit is exposed as a circular shaped erosional-bound area nearly 500 feet in diameter. As with the Lovell deposit, it is recognizable by its black color which is quite different from the lighter color host sands. The deposit forms an impressive dip slope on its upper surface, with an associated scarp face composed of over 50 feet of massive, buff sandstone. The titaniferous black sand is confined to the upper one foot, so if indeed the two deposits were originally one continuous sheet, a considerable southward thinning is indicated.

Titaniferous fossil beach placers occur frequently in Upper Cretaceous littoral sandstones (Houston and Murphy, 1962 and 1977, MacKenzie, 1975). Throughout the Big Horn Basin these features are valuable paleoenvironmental and palcocurrent indicators, and are significant sedimentary features.

Middle Judith River

Strata included with the middle Judith River Formation are characterized by repeated sequences of carbonaceous mudstones, and shales, sandstones, and siltstones, ranging from 90 to 130 feet thick.

The characteristic which differentiates this unit from other parts of the Judith River is the abundance of carbonaceous beds over sandstones (plate 5A). The basal contact of the Middle Judith River is abrupt, with a sharp break between carbonaceous sediments and underlying massive sands. Individual carbonaceous shale and mudstone beds average 15 to 20 feet thick, and are very rich in fossil plant material, although no lignite or coal beds were found. Descrete sandstone beds range in thickness from one to 30 feet, and are very lenticular, frequently grading laterally in carbonaceous sediments. Basal contacts of sand units are mostly erosional, although no distinct cut and fill relationships were noted. Frequently, the lower six inches of sand beds contain carbonaceous mudstone clasts and fossil plant debris eroded from subjacent mudrocks. Fossil plant material is also present in sands in the form of detrital grains and dark films with clay skins along bedding surfaces. Thicker sandstones display fining-upward textural cycles, occasionally repeated several times in a single sequence.

Upper Judith River

Sediments contained in the upper Judith River uphold prominent cuestas and hogbacks throughout the study area. Rocks consist of repetitive sequences of sandstone and thin interbedded shales ranging from 90 to 100 feet in thickness. The basal contact of this unit is sharp, placed at the top of the uppermost carbonaceous sequence in the underlying middle Judith River section.

Sandstone bodies within this sequence are laterally contiguous, sheet sands, averaging 20 to 30 feet thick, and are strikingly similar to those of the basal lower Judith River. These representative sequences range from 20 to 50 feet in thickness, with shale beds gradually upward, until the sequence is composed entirely of sands. Basal contacts of sandstones upon underlying shales are gradational near the base of each coarsening upwards cycle and become erosional towards the top of each unit (plate 5B).

Carbonaceous debris is not common in upper Judith River strata, but are present in both mudstones and sandstones, as thin films and partings of unoxidized fossil plant material concentrated along bedding planes. Thick beds of carbonaceous sediments common in the middle Judith River are not present in upper Judith River beds.

<u>Regional unconformity at the base of the Teapot Sandstone</u>

Regional biostratigraphic evidence based on the evolutionary sequences of several ammonite genera, and regional sampling presented by Gill and Cobban (1966, 1973) indicate a regional angular unconformity exists at the base of the Teapot Sandstone Member throughout the southern and eastern Big Horn Basin area. Although an unconformity is not obvious in outcrop, faunal zones are shown to be dipping below the Teapot, eastward across the southern portion of the basin. The amount of erosion is shown to increase westward as progressively more strata are inferred missing. Over 1.00 feet of section and as many as 13 faunal zones may have been truncated by the proposed unconformity in the far west portion of the basin. Although paleontologic evidence exists, MacKenzie (1975) has concluded on the basis of subsurface correlations and surface investigations, that no single extensive unconformity exists within the Mesaverde Group. The writer has found no evidence indicating the presence of an unconformity at the base of the Teapot, although Cobban (personal communication) indicates the unconformity is subtle in outcrop.

An alternative to the interpretation of Gill and Cobban (1966b) proposed by Asquith (1974) may better explain the apparent disparity between faunal evidence and the lack of lithologic evidence for a significant unconformity. Asquith (1974) in subsurface correlations of bentonitic shales within the Cody Formation at Nieber Anticline, in the southeastern Big Horn Basin shows the presence of eastward dipping time-stratigraphic units reflecting sedimentation on a depositional topography consisting of a marine shelf, slope, and basin facies. These eastward dipping time lines may be associated with foreset and bottomset deltaic facies (Asquith, 1974). Rea and Barlow (1975) have found comparable depositional slopes in the same area in the upper Cody and lower Eagle interval, suggesting a progressive eastward shift in the locus of deposition, possibly associated with slopes near prodeltaic environments. Time stratigraphic units thin shoreward to the west and basinward to the east, away from the inferred deltaic depocenters.

MacKenzie (1975) has proposed a hypothesis similar to Asquith (1974) accounting for the overall westward decrease in stratigraphic intervals between the top of the Mesaverde Group and index faunal zones within the Group. The 1,000 plus feet of thinning at the base of the Teapot Sandstone Member ascribed to uplift and erosion (Gill and Cobban, 1966), can be accounted for by thinning in a delta or possibly coastal plain setting. The westward loss of index fossil zones can easily be explained by the fact that marine strata were never deposited, and are replaced by thin sequences of delta plain sediments. A westward facies change from marine shales to continental facies, with subsequent thinning due simply to nondeposition is the most plausible explanation for loss of ammonite zones. At any given time interval, time surfaces were inclined to the Teapot, while Judith River sediments were deposited along with transitional marine (delta front) and prodeltaic marine sediments progressively eastward.

Teapot sandstone member

The Teapot Sandstone Member, the upper unit of the Judith River Formation, upholds the highest cuestas and hogbacks throughout the outcrop trend. These light grey to white ridges (plate 5C) stand in sharp contrast to dark colors of the overlying Meeteetse Formation. Impressive dip slopes are formed on top of the Teapot, which are enhanced by the poorly indurated nature of the overlying unit. The Teapot is almost totally sandstone, with the exception of thin, lenticular shales occurring near the upper contact.

Near the southeastern margin of the field area, the Teapot averages over 80 feet, thinning northwestward to near 50 feet near Lovell. The basal contact of the Teapot is distinct, placed immediately below the first zone of spherical ironstone concretions which occur as discrete beds throughout the sequence.

Lithologically, the Teapot is distinctive as it is by far the coarsest-grained unit within the Mesaverde Group. Major lateral and vertical changes in grain size occur with very short intervals. Local discontinuous lenses of medium to coarse-grained sandstones are common near the base of the unit. Grain size increases upward, but is interrupted by repeated fining-upwards cycles ranging from one to three feet thick. Individual cycles display highly erosional basal contacts (plate 5D). Intimately associated with these contacts are clay choked sands, unoriented rip-up clasts, and carbonaceous mud galls, accounting for as much as 50 percent of the sediment volume. Also scattered throughout the Teapot are carbonaceous mudstone lenses, averaging several inches in longest diameter, oriented parallel to bedding surfaces (plate 5A).

The most distinctive aspect of the Teapot is the presence of interbedded lenses of rounded, unoriented ironstone concretions of diverse sizes, with little or no sand matrix (plates 6B-D). Individual lenses range in length from three feet to several tens of feet, characterized by highly erosional, scoured basal contacts. The concretions show no internal structure, and no secondary reaction rims are present in the host rock. Associated at times are unoriented carbonaceous and Kaolinitic rip-up clasts displaying no evidence of oxidation. The origin of these concretionary bodies is not known, but are primary depositional features as evidenced by their highly scoured basal contacts.

DEPOSITIONAL ENVIRONMENTS

The environments of deposition of modern and ancient sedimentary rock sequences can be delineated on the basis of rock body geometry, composition, texture, and stratification types and sequences. Geometric properties of paleoenvironmental complexes preserved in ancient rocks have resulted from the migration through time of the component subenvironments. Although very important, body geometries in many cases cannot be studied or may be quite different from the modern analogues, which in many cases represent only small portions of the total time-migration picture. Compositional properties are products of the source areas, and of the depositional and diagenetic environment, and are not always valid indicators of past environments.

Stratification type and sequence in bedding is indeed the most valuable aid in the determination of sedimentary environments. This was recognized very early by Sorby (1859) who predicted that the stratification in sedimentary rocks might some day be used in the reconstruction of ancient environments.

Harms and Fahnstock (1965) and Simons and Richardson (1961, 1962) and Simons et al. (1965) were the first to recognize the relationship of certain stratification types as being the result of migrating bedforms associated with low, transitional, or high flow regime conditions. It was further recognized by these and many other authors that stratification types which are associated with certain flow regime conditions are dependent upon water depth and bed configuration, particle size and shape, and the density of the fluid-sediment mixture. From information regarding processes and flow in alluvial channels, stratigraphers recognized the importance of sedimentary structure sequences in that they are the product of bedform migration through time, and that sequences of bedding types record the lateral movement through time of associated sedimentary environments.

In the last 20 years, much interpretive work has been done, and research regarding bedform migration has expanded from alluvial channel to all sedimentary environments. Of major importance to this study is the application of the flow regime concept and associated bedform phenomena to various beach environments by Clifton (1976), Clifton, Hunter, and Phillips (1971), Davidson and Greenwood (1974 and 1976). These authors have recognized basic similarities between wave dominated beach environments, and stream channels, and have presented useful models for the formation of primary structure sequences in barred and non-barred beach environments.

In detail, Clifton et al. (1971) has recognized a typical sequence of primary structures produced by the progradation of a high energy non-barred shoreline. Sequences of small to large scale bedforms shoreward in a beach profile are comparable to the flow regime model and associated bedforms in alluvial channels under waning flow conditions (Clifton, 1971).

The vertical sequence of primary structures in a beach sandstone is a record of the superposition of the lateral components of a beach profile. Upward in a progradational beach sequence, the stratification reflects physical processes operating in the following subenvironments: Offshore beach-lower shoreface transition, shoreface or seaward slope (Davidson and Greenwood, 1976), submarine bar zone, foreshore, berm crest, and backshore (Masters, 1965, 1967). Sedimentary structure sequences are varied and complex, reflecting the wide variety of beach processes, including tidal, longshore, and storm induced current and wave activity. Stratification indicative of offshore bars with associated rip currents and tidal inlets are commonly present in modern beach profiles (Davidson and Greenwood, 1976), and are recognizable in ancient beach deposits (Masters, 1965 and 1967). Although beach sequences are present in the Eagle, lower and upper Judith River sections, Eagle sediments are unique in that they contain only lower shoreface-offshore transition and shoreface environments. All lateral components of a typical beach profile are preserved in the Judith River Formation. Large scale sedimentary structures deposited in nearshore zones, or under highly asymmetric translational upper flow regimes (Clifton, 1976, Clifton, Hunter, and Phillips, 1971, and Davidson and Greenwood, 1976) are absent in Eagle rocks while common in Judith River beach sequences.

Primary Structures and Sequences of the Shoreface-Offbeach Transition Zone

Sediments deposited in the shoreface-offshore transition zone are most readily recognized in the Eagle formation where they interfinger or are gradational to underlying offshore marine sediments of the upper part of the Cody Shale. The contact between the two environments is a reflection of the sand-mud transition zone, where sand is periodically introduced during high energy storm surges (Masters, 1965). The resulting lithology consists predominately of laminated shales, mudstones, and siltstones, with occasional thin sandstone interbeds containing a variety of ripple lamination types. This lithologic type is similar in position to other Late Cretaceous deltaic and interdeltaic progradational outer beach zones described by Curry (1976a, 1976b), Curry and Crews (1976), Douglass and Blazzard (1961), Davis and Todd (1976), Howard (1972), Harms et al. (1975), Merewether, Cobban and Spencer (1976), and Masters (1965 and 1967) commonly underly massive sandstones deposited in the shoreface beach zone.

Stratification sequences include thick, 10 to 30 foot sequences of a laminated (Campbell, 1968) sandy siltstones, overlain by flaggy sandstones characterized by type B ripple-drift lamination of Jopling and Walker (1968). These sequences are found superjacent to shale tongues of the upper Cody, and reflect increasing wave current energy coupled with abundant suspension deposition.

Near the south nose of Alkali Anticline, shoreface-offshore transition sediments are present throughout the Eagle and are represented by thick, coarsening upward sequences of sandy shale, siltstone, and thin silty sandstones. Sands invariably display ripple surfaces with ripple crest orientation mainly northeast-southwest, subparallel to the inferred depositional strike (Gill and Cobban, 1973, Houston and Murphy, 1977, and Zapp and Cobban, 1960).

Wave tank experiments (Scott, 1954) have shown that in the shoreface and shoreface-offshore transition areas ripples can become strongly asymmetric landward, as shoreward current velocities become greater than seaward return velocities. The shoreward migration of ripple forms accompanied by sediment fallout may produce ripple-drift lamination, modern examples of which are described by Davidson and Greenwood (1976).

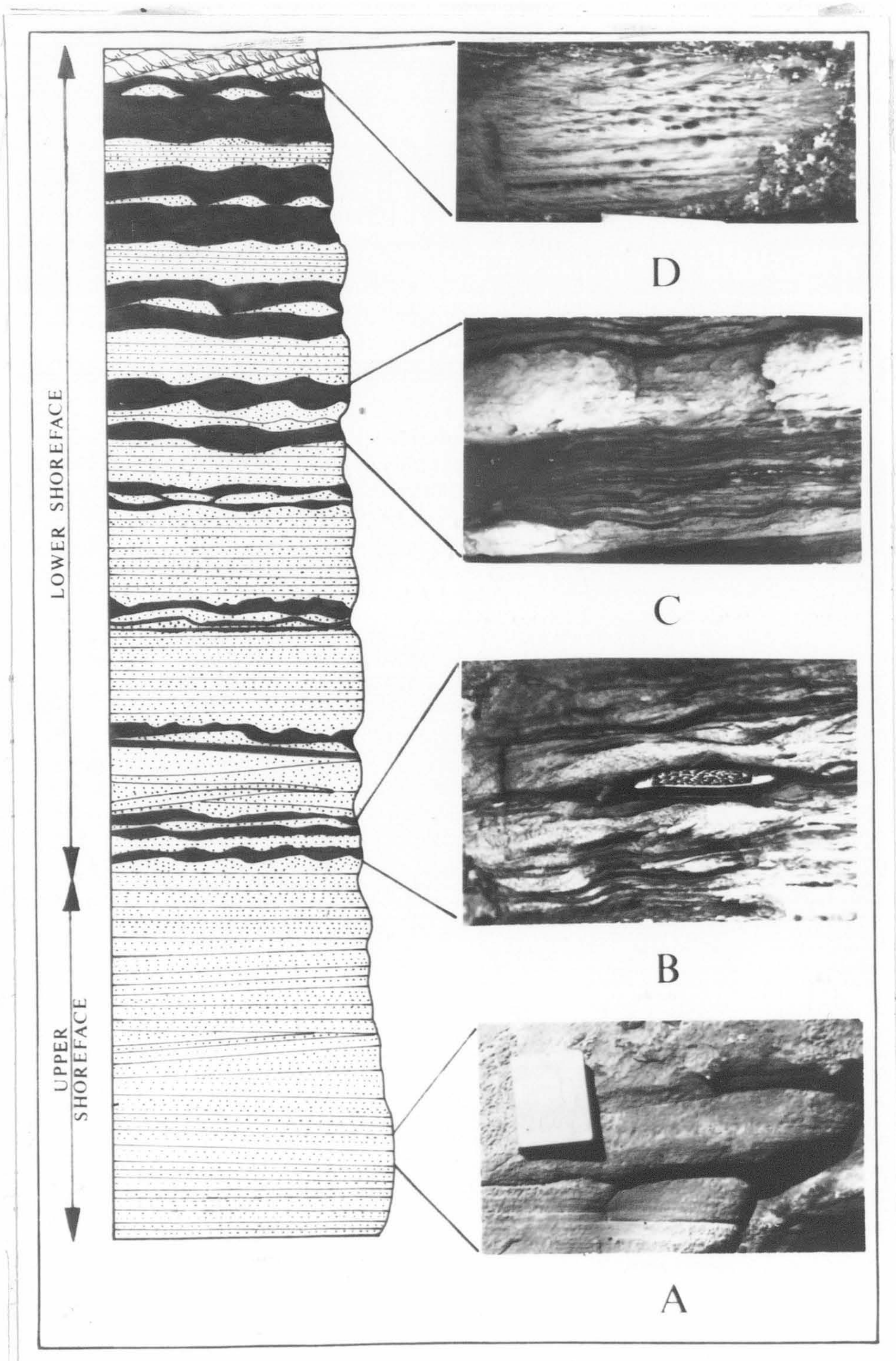
In the Eagle interval near Alkali Anticline, examples of ripple-drift lamination sequences are common. Bedsets of superimposed sinusoidal ripple lamination are overlain by type B ripple-drift lamination, which in turn, is overlain by type A ripple-drift sets (Jopling and Walker, 1968). This sequence is indicative of a gradation from high suspension, low bed load movement, to low suspension, high bed load movement, with no preservation of stoss side laminae in the ripple sets (type A Jopling and Walker, 1968) (plate 7A). This sequence indicates a transition from predominately low energy suspension deposition in offshore environments, to periodic higher energy deposition from bedload currents in the shorface-offshore transition area.

Shoreface Environments

Above the transition zone, shoreface sandstones are massive and frequently interbedded with shale and mudstone laminae, suggesting intermittent sand transport in this zone. Bedform features described in the "landward slope facies" by Davidson and Greenwood (1976) are comparable to those of shoreface sediments, and consist of alternating sequences of plane beds or low angle, large-scale planar or hummocky cross bedding and small-scale bedforms. These small-scale units were deposited in the asymmetrical low flow regime (Clifton, Hunter, and Phillips, 1971), under low current velocities. With increasing wave period these bedforms are washed out to form a plane bed. The shoreface zone as described by Davidson and Greenwood (1976) is characterized by composite bedsets of ripple to plane bedding.

Stratification sequences preserved in Eagle, lower and upper Judith River formation, shoreface sediments are similar to those described by Davidson and Greenwood (1976), with ripple forms preserved as wavy and lenticular bedding (Reineck and Wunderlich, 1968), ripple-drift lamination, or single wavy mudstone beds. These bedforms are interbedded with horizontal bedding, gently inclined planar crossbedding, or hummocky crossbedding. Typical shoreface sequences include thin units of asymmetric wavy mudstone and sandstone beds, overlain by thicker sets of plane beds or gently seaward dipping large scale planar crossbedding (fig. 17A-D). Plane beds are more common than small scale bedforms, reflecting shallow water depths and higher current

Figure 17. Section of upper Eagle sandstone as exposed at Little Dry Creek, with photographs documenting environmental interpretations based on bed form sequences.


LOWER SHOREFACE
UPPER SHOREFACE
D
C
B
A

velocities than required for ripple migration and preservation (Harms and Fahnstock, 1965).

Where large scale bedforms overlie wavy bedding in the upper shoreface zone, the basal contact of sands are seaward dipping erosional surfaces with frequent concentrations of rip-up clasts and mud galls, eroded from underlying mudstones. In a seaward direction, these contacts become gradational, reflecting lower current velocities associated with deeper water. Thicknesses of both small and large scale bedforms decrease seaward, again due to decreasing rates of sediment transport in deeper water.

Lower shoreface environments are preserved as sequences of large scale, low angle planar crossbed sets overlain by thin beds of laminated shale, indicating gradation from lower upper, to lower lower flow regime in a seaward direction. Occasionally, plane-to-ripple bedsets are capped by type B ripple-drift lamination, or bipolor ripple-drift sets, reflecting landward currents associated with shoaling waves (fig. 17-D).

A unique stratification type associated with plane-to-ripple bedsets is that of hummocky crossbedding originally described by Harms et al. (1975). Commonly the lower contacts of these sets are erosional, with laminae within a cross bed set thickening laterally to form fanlike vertical traces (plate 7B). The hydrodynamic implications of hummocky crossbedding.are not well-known, although they are found with rippled bedforms in association with offshore and beach sand bodies. Harms

(1975) suggests that hummocky crossbedding is formed mainly in shoreface and foreshore environments, during strong surges of varied direction, generated by periodic storm waves.

Overlying shoreface sediments are well-developed progradational beach sequences preserved in the lower and upper Judith River Formation. Similar progradational beach sequences are well-documented in the Late Cretaceous Western Interior (Curry, 1976a and 1976b, Harms, et al. 1975, Masters, 1965 and 1967, Merewether, Cobban, and Spencer, 1976, and Sabins, 1965), and illustrate progressive sedimentation from offshore through backshore beach environments upward. Stratification sequences indicative of the shoreface beach zone are identical with those of the Eagle Formation. However, trace fossils, which are not abundant in the Eagle Formation are more abundant in the lower Judith River. The following types were identified by the writer in shoreface sediments.

1. Simple snail trails confined in occurrence to the upper six inches of sand bodies, concentrated along bedding surfaces (plate 8A-B).
2. A straight, vertical burrow identified as <u>Asterosoma</u> (Howard, 1972) which occurs on bedding surfaces as a cup-shaped depression (plate 8C).
3. Straight, smooth, unbranching vertical burrows about one quarter inch in diameter with no identifyable internal marking.
4. <u>Ohiomorphia</u> <u>sp.</u> (plates 8D-E).

These forms occur within the laminated siltstones and interbedded sands characteristic of the lower shoreface environment and have been described by Howard (1972) for the Upper Cretaceous Book Cliffs Formation, and by Curry (1976b) from the Teapot Sandstone in the Powder River Basin. These trace fossils define a water depth zonation described by Howard (1972) in offshore to foreshore beach environments successively higher in a progradational beach sequence.

Vertical and horizontal branching burrows known as Ophiomorpha sp. are thought to represent the burrowing of a decapod crusacean similar to the modern form Calianassa major. These forms are very common beach indicators in the Upper Cretaceous progradational shoreline sequences (Curry, 1976b, Hoyt and Weimer, 1965, Lewis, 1961, MacKenzie, 1975, and Toots, 1961). Although Ophiomorpha occurs in many morphological types, it is identifiable by its typical corn cob structure (plate 8E). In the study area, this burrow consists of vertical and horizontal branching burrows, 0.25 inches in diameter, characteristically forming a U-shaped pattern (plate 8D).

Submarine Bar Zone

Submarine bars are asymmetric migrating megaripples composed of steeply dipping landward beds, interbedded with gently dipping seaward strata (Davidson and Greenwood, 1976 and Masters, 1965). Bedform generation is controlled by waves breaking on the bar, with the interaction of waves with currents flowing across the bar. The unique stratification of the bar zone is the result of plunging waves that build and cause migration, and associated rip currents which drain and dissect the back-bar area (Davidson and Greenwood, 1976 and Masters, 1965). Rip currents are caused by plunging waves which pile water up behind bars, creating localized currents parallel to the bar trend. Where two opposing rip currents meet, they are forced to cut a seaward channel through the bar which may migrate laterally, similar to unidirectional currents in meandering stream channels. Sedimentary structure sequences associated with submarine bars can occur in shoreface and foreshore areas, thus may be found in various locations in a progradational beach sequence.

Bedding sequences consist of interbedded large scale, high angle planar or trough cross bed sets formed on the lee side of bars, interbedded with large scale, low angle seaward dipping planar cross beds. Occasionally, small scale trough sets associated with ripple surfaces are present at the top of the submarine bar sequence and are formed by sedimentation on the bar crest in periods of low wave energy (Davidson and Greenwood, 1976).

Bedding sequences indicative of rip currents occur in the upward portions of submarine bars, consisting of landward dipping large scale trough cross beds overlain by type A or B ripple-drift lamination, or bi-polar ripple drift sets (plates 7C-E). These bedding types indicate seaward expansion of rip currents over the breaker zone, coupled with the rapid decrease in current energy and finer material remaining in suspension. Bi-polar ripple-drift sets may be the result of the influence of tidal currents over the bar crest area.

Foreshore

The foreshore portion of the beach is the area influenced by daily tidal cycles, and is developed as a result of swash and backswash of the surf. The foreshore slope is controlled by wave characteristics, associated with the energy of the swash-backwash currents, and by the grain-size of the material present. Any change in these parameters causes slope changes in the foreshore, and if the effectiveness of the swash is increased relative to backswash, a swash bar may be constructed on the foreshore. Migration of the swash bar is a result of daily tidal fluctuations, and causes interbedded steep, lee side cross bedding and low angle seaward dipping cross bedding (Thompson, 1937, Masters, 1965).

Foreshore bedding sequences in basal Lower and Upper Judith River sediments consists of horizontally bedded sands, overlain by large scale bi-polar sets of planar or trough cross beds with landward cross bed sets dipping at steep angles than seaward cross bed sets (fig. 17B). Frequently, thick bedsets of these sequences 10 to 25 feet thick overlie submarine bar and shoreface beds, and are termed "swash stratification" (Harms, et al. 1975). Sequences of this bedding types associated with foreshore sedimentation have been recognized in Upper Cretaceous strata by (Harms, et al. 1975, Howard, 1972, Masters, 1965 and 1967, Houston and Murphy, 1977, and Rieneck and Singh, 1973).

Barrier Island Environments

The origin of Barrier island complexes is a continuing controversy with a total understanding of hydrodynamic interrelationships not yet fully achieved. Johnson (1919) initially proposed that waves erode sediments from offshore areas and deposit them in the breaker zone, forming bars, which grow upward above sea level by addition of sand into barrier islands. Hoyt (1967) cites evidence against the accretionary model in that marine sediments are absent landward of barriers in ancient and modern examples, and has proposed that submergence and drowning of beach ridges and dune fields may be the cause of many barrier islands.

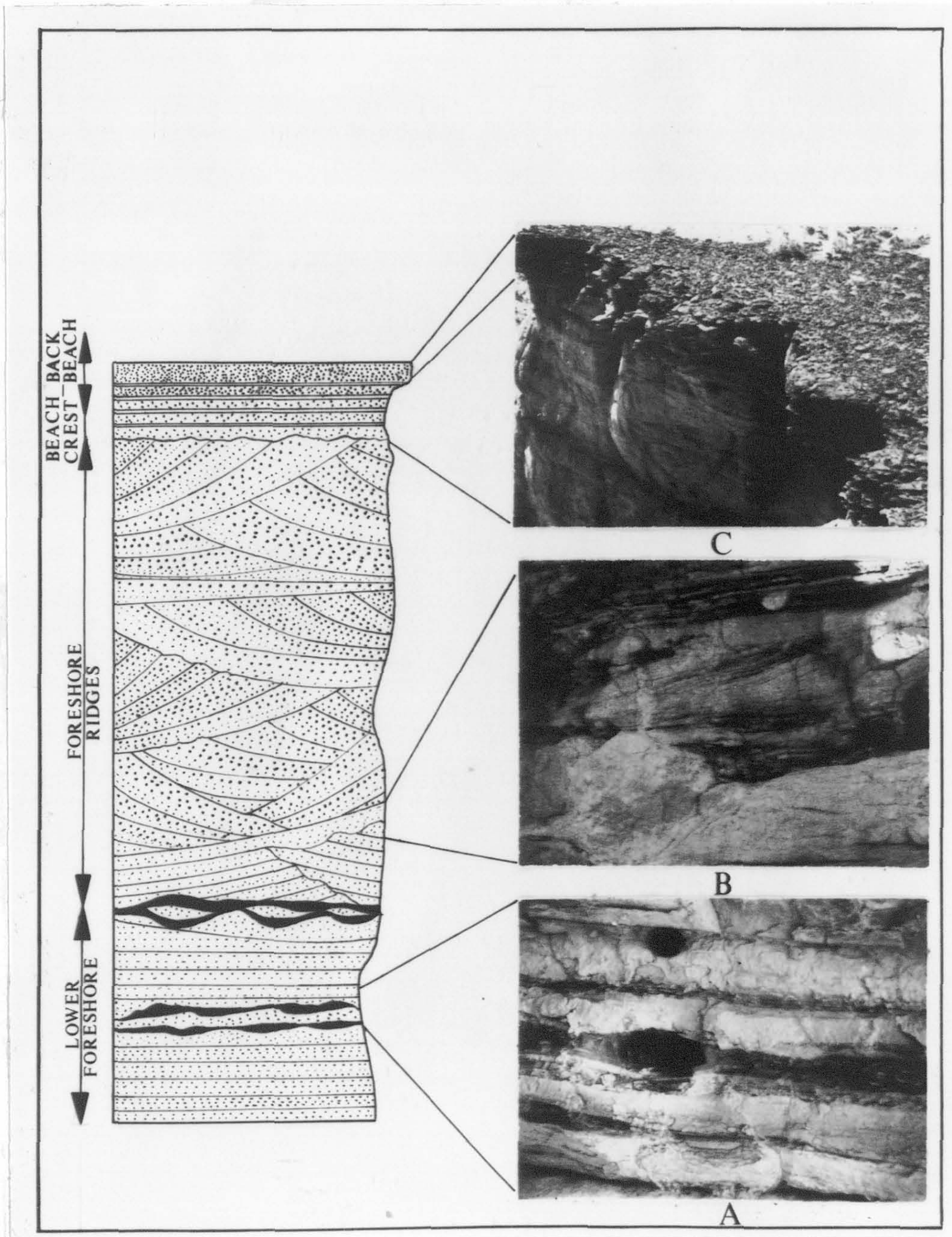
Regardless of the origin, modern and ancient barrier bars are recognized mainly by their relationships with sequences of primary structures and with overlying washover fans, tidal flats, and lagoonal deposits. Primary depositional features of progradational Upper Cretaceous interdeltaic, barrier bars have been documented by many authors, including Berg and Davies (1968), Davies, et al. (1971), Dickenson, et al. (1972), Houston and Murphy (1977), Peterson (1969), Ryer (1977), Shelton (1963 and 1965), Jacka (1965), and Weimer (1966 and 1975). Modern analogues with similar sequences of primary structures include Galveston, Padre, and Sapelo Islands (Bernard, LeBlanc, and Major, 1962, Dickenson, Berryhill, and Holmes, 1972, Friedman, 1967, Davies, et al. 1971, Dickenson, et al. 1972, Hoyt, 1967, Bernard, Major, and Parrot, 1970, Hoyt and Henry, 1967). Bedding sequences described by the writer are comparable to sequences of primary sedimentary structures described by all of the above authors.

Black Sandstone deposits

Titaniferous black sand deposits are associated with the upper portion of the lower barrier sequence in the Lower Judith River and are considered the most reliable environmental indicators and shoreline markers in the Mesaverde Group. Internal stratification features of sandstone bodies immediately subjacent to black sand deposits have been described by Houston and Murphy (1977) are are identical to those described by the writer.

The sequence of primary structures preserved below the backbeach concentration of heavy minerals consists of a four part sequence, and depicts from base to top, stratification typical of lower foreshore to back beach zones. As shown by figure 18A-C, the basal sandstone consists of horizontally bedded to large scale, low angle, planar cross bed sets, interbedded with single wavy bedded mudstones deposited in the lower foreshore zone (fig. 18A). The overlying beds are characterized by large scale, high angle, wedge shaped tough crossbed sets, with both landward and seaward dip directions preserved in individual sets (fig. 18B). Basal contacts of individual cross sets are highly erosional with internal laminae consisting of fine-grained, heavy mineral rich laminae, overlain by laminae rich in coarser-grains and quartz. This feature is described as reverse textural grading (Clifton, 1969) is characteristic of swash zones in which rapid segregation of coarse and

Figure 18. Bedform sequence diagnostic of regressive beach sequences underlying fossil beach placer deposits.


BEACH CREST
BACK BEACH
FORESHORE RIDGES
LOWER FORESHORE
A
B
C

fine and light and heavy minerals occurs within a moving layer of sand. Hoyt (1962) has noted in studies of Sapelo Island, Georgia barriers, that landward slopes of upper foreshore ridges may exceed 30 degrees, while seaward slopes frequently are less than 10 degrees, and are defined by laminae of black heavy minerals alternating with layers of quartz. The landward dipping units in zone 2 of the placer deposit are therefore interpreted not as beach crest ridges, but foreshore ridges on the basis of its position within the overall sequence and by similarities with Sapelo Island sediments. The overlying units contain large scale, low angle planar crossbeds with abundant reverse textural grading, associated with laminae containing over 60 percent heavy minerals (fig. 18C) deposited in the uppermost foreshore areas. The uppermost, thick, black, heavy mineral concentrates are considered to be storm generated black beach deposits (Houston and Murphy, 1962 and 1977). High energy storm waves frequently are capable of destroying or flattening the upper portions of the foreshore and berm zones, which may account for the absence of high angle crossbeds, typical of berm crest beach zones.

The black sandstones are the most reliable shoreline indicator in Mesaverde equivalent rocks throughout the Western Interior region. If exposed in three dimensions, and if the top of the deposit is uneroded, the thickest portion of the deposit is considered to accumulate landward, and the long axis of the deposit to be parallel to the general shoreline trend.

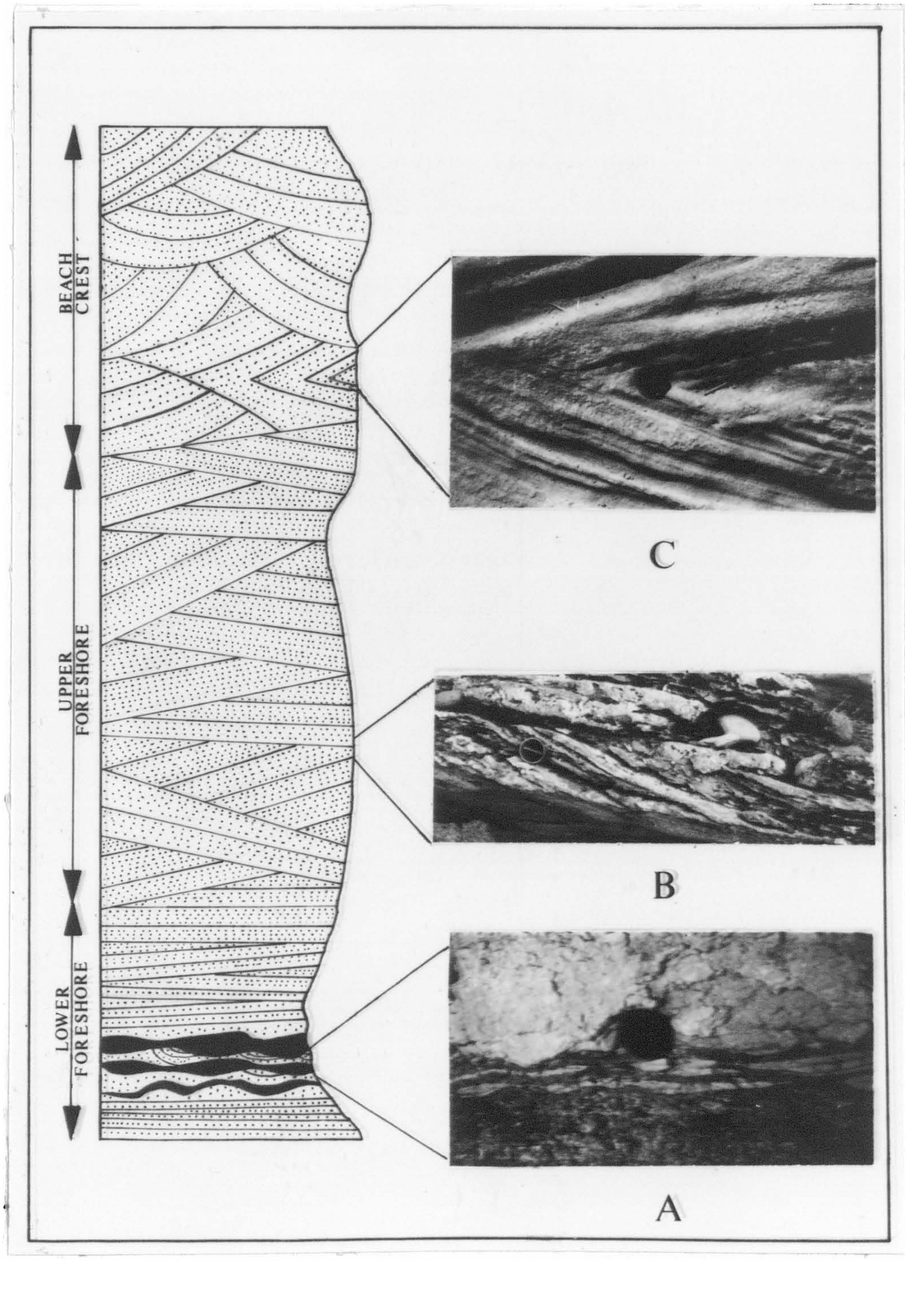
The form and shape of the deposits suggest that they might give clues as to the prevailing storm wind directions. Gardner (1955) and Houston and Murphy (1977) point out that in Australian black sandstone beaches, the thickest heavy mineral deposits occur on beaches that terminate against a natural barrier in the direction of striking storm generated waves. Using this analogue, Houston and Murphy (1977) have demonstrated prevailing wind and current directions in the southeastern portions of the Big Horn Basin, and in the study areas as being predominately to the north, since the placer deposits thicken in this direction.

Barrier Island sequences

Sequences of primary structures indicative of barrier bars are comparable in the lower and upper barrier units and are depicted in figures 19 and 20. The lower barrier sequence is distinctive in that it is always overlain by thin units of carbonaceous shale, deposited in trough zones behind the submerged bar, while the upper bar sequence records a seaward progradation, and is overlain by thick sequences of oyster concentrates, interbedded with thin coquina beds of lagoon origin.

The tripartite upper barrier sequence consists of a basal zone of interbedded large scale planar crossbedding and wavy bedded units, forming bedsets 5 to 10 feet thick, and is indicative of lower and middle foreshore zones (Harms et al., 1975) (fig. 19A). The middle zone consists of massive beds with internal large scale, low angle planar

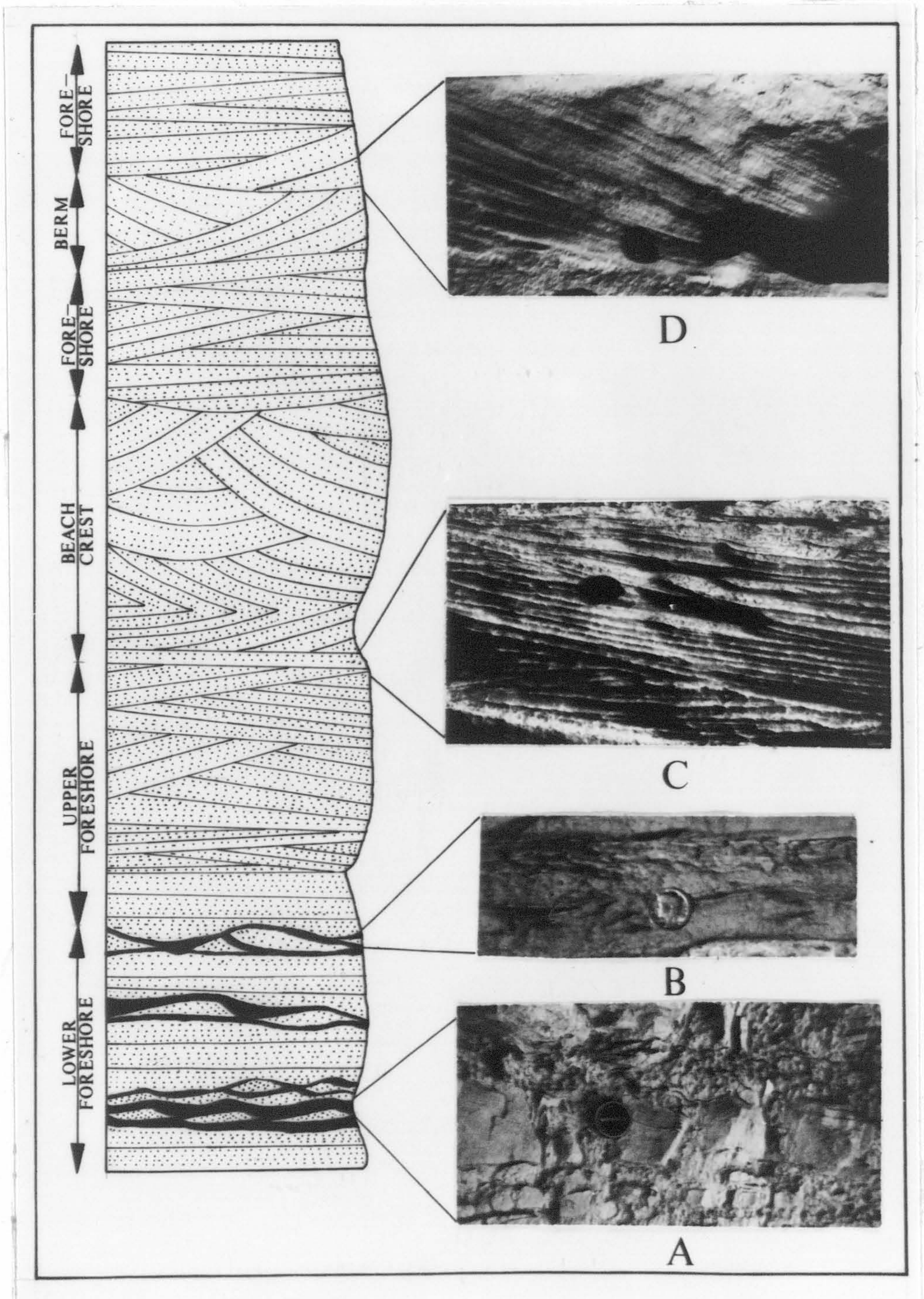
Figure 19. Tripartite Barrier Island sequence illustrating lower foreshore through back beach dune deposition.


BEACH CREST
UPPER FORESHORE
LOWER FORESHORE
C
B
A

crossbeds sets dipping both landward and seaward, characteristic of upper foreshore deposition (Jacka, 1965) (fig. 18B). The upper zone displays distinctive sets of very large scale high angle tangential trough crossbedding, with individual sets three to 10 feet thick. Rare herringbone crossbed sets (fig. 18C) may reflect tidal influences in the beach crest area. These crossbeds display erosional basal contacts and show high variability in trough axis orientation, which may be indicative of backbeach dune processes (Houston and Murphy, 1977). A unique characteristic of the lower barrier sequence is the repetition of middle and upper sequences, with uppermost units containing bedding indicative of the foreshore zone (fig. 19). This reflects increasing submergence of the barrier, and deposition in deeper water, lower energy conditions, as reflected decreasing grain-size near the top of the unit. Overlying thin carbonaceous beds may be associated with accumulation of organic material in a trough zone immediately landward of the bar, similar to those described by Davidson and Greenwood (1976).

Separating the lower and upper barrier bar units are thick, repetitive sequences containing horizontal to low angle planar crossbed sets, interbedded with ripple bedding. Plane bed-ripple bedsets are repeated up to 10 times within individual sand bodies, with sand increasing in thickness upwards. Sequences of this type are interpreted by the writer as representing cyclic deposition in a shoreface and foreshore zone similar to that described above for Eagle sediments.

Figure 20. Lower barrier bar bedform sequence, illustrating repetition of foreshore and back beach sediments.


FORE-SHORE
BERM
FORE-SHORE
BEACH CREST
UPPER FORESHORE
LOWER FORESHORE
D
C
B
A

Lagoonal Environments

Lagoonal sediments include a complex of subenvironments which include lagoon pond, tidal inlets and deltas, and salt marsh. A typical sequence of lagoonal rocks in the middle Judith River consists of interbedded carbonaceous mudstones and black shales, with lag concentrates of coquina, interbedded with oyster-bearing siltstones and sandstone beds. The high variation in lithology and bedding type are in themselves typical of the lagoonal setting, reflecting chemical and energy states associated with each subenvironment. Commonly rocks of all subenvironments are not all exposed in a single outcrop, hence the recognition of the lagoonal complex is dependent upon the documentation of one or more of the component environments.

Lagoon pond

Interbedded black, carbonaceous shales and laminated siltstones and sandstones typify sediments deposited in low energy lagoon pond environments. Thin shale and siltstone laminae may be deformed by penecontemporaneous lead flowage (Masters, 1965) during sand deposition. Where superposition of sufficiently thick sandstone layers are present, psuedonodules may form if the underlying muddy sediments are sufficiently plastic to flow around overlying sand beds (plate 7F-G). Masters (1965) has noted comparable rheotropic deformational structures in lagoon pond sediments of the Castor Member of the Iles Formation. Thin oyster beds are occasionally enclosed by lagoon pond sediments reflecting the high variations in salinity associated with this environment.

Tidal inlets and deltas

Tidal inlets are bodies of water connecting the backbar lagoon with the open marine environment. These features are closely associated with tidal deltas, which form adjacent to inlets cut through the barrier where channel velocities are suddenly lowered by the standing water of the lagoon pond.

Sandstone beds of tidal inlet channel origin are found interbedded and overlying sediments of the lagoon pond, giving rise to interbedded shale and sandstone sequences. Inlet sands resting upon lagoon pond shales display sharp, erosional basal contacts with small channeled areas, containing carbonaceous rip-up clasts. Tidal channels are recorded also by thin layers of coquina deposited and concentrated as lag deposits near the base of channels migrating across lagoon pond sediments. Bedding sequences are representative of high to low flow regime conditions accompanying lateral stream channel migration (Allen, 1965b). Basal sands contain horizontal to large scale, low angle planar crossbedding, overlain by superimposed ripple laminated units, with occasional sets of wavy bedded sandstones and mudstones. Mudstone interbeds are associated with upper portions of inlet sequences and reflect channel inlet migration over previously deposited lagoon pond sediments. Tidal inlet beds grade laterally into or under tidal delta sediments which display ripple-drift lamination formed during waning current conditions associated with delta formation. Support for tidal delta origin lies in the close lateral association with tidal inlet and lagoon pond sediments, and the presence of bi-polar ripple drift sets, which may indicate the influence of alternating tidal flow directions (Rieneck and Singh, 1973).

Salt marsh

Interbedded with other lagoonal sediments are carbonaceous shales and mudstones of the salt marsh environment. Thick marsh grasses typically grow and accumulate in shallow water, eventually forming a turf which builds vertically, keeping pace with subsidence. In Mesaverde strata, this turf becomes a carbonaceous mudstone, with abundant fossil plant debris (Masters, 1965 and 1967). These beds range from 10 to 15 feet in thickness and may reflect relatively long periods of shoreline equilibrium during which vertical buildup and subsidence processes could operate.

Teapot Sandstone Member Depositional Environments

The origin of the Teapot Sandstone Member within the Big Horn Basin, and its lateral equivalents in adjacent basins has been, and is still a controversial issue. As with the rest of the Mesaverde Group in the Big Horn Basin, the Teapot has not been studied in detail with regard to the determination of depositional environments. Environmental interpretations based solely on general lithofacies patterns is probably the crux of the controversy, as little attention has been paid to more obvious paleoenvironmental indicators. Gill and Cobban (1966) report the Teapot in the southern Big Horn Basin to be nonmarine, while Severn (1961) has described the same unit in the same location as combined beach, lagoonal, and fluvial. MacKenzie (1975) has recognized fining-upwards cycles, associated with changing high to low flow regime conditions, similar to those described in modern stream channels by Allen (1965). Obviously much more detailed paleoenvironmental work needs to be done before conclusive evidence can be accumulated in these areas. Within the study area the Teapot Sandstone Member is interpreted by the writer as representing overlapping meandering stream channel complexes, which were part of a larger coastal or delta plain setting. Bedding sequences are indicative of stream channel bedform migration under changing high to low flow regime conditions as described by Harms and Fahnstock (1965), Allen (1964, 1965a,b,c,d and 1970), Allen and Friend (1968), Simons and Richardson (1961 and 1962), and Visher (1972).

Bedding sequences preserved in Teapot sediments are comparable to fining-upwards cycles described by the above authors. These repetitive, fining-upwards sequences are typically one to three feet thick, and are characterized by a highly erosional basal contact (plate 5D). Associated with the scour contacts are concentrations of large carbonaceous mud galls, and unorientated rip-up clasts, often exceeding 50 percent of the sediment volume. Scoured surfaces are associated with large scale trough crossbed sets, averaging one foot in thickness. These sets are similar to the Pi cross stratification described by Allen (1965a,b,c,d), and have scooped shaped erosional bounding surfaces when viewed in three dimensions. The overlying finer-grained units are plane bedded with bedset thicknesses reaching 10 feet at times. Plane beds are usually found with lenses and stringers of carbonaceous mudstones and coals (plate 6A) comparable to those described by Reineck and Singh (1973) and may represent a channel-floodplain association. Fining-upwards cycles are repeated several times throughout the Teapot with scoured contacts truncating both trough and planar crossbed sets.

Bedding sequences comparable to those described above have been examined by Allen (1965b), Bernard and Major (1963), Leopold and Wolman (1957), McGowen and Garner (1970), and Miall (1977a and 1977b) and are considered to be produced during a transition from high to low flow regime in stream channels. Complex stream channel migration and down-cutting upon older fluvial sediments is indicated by erosional basal contacts and the repetitive nature of the individual fining-upwards units within the Teapot. However, the identification of point bar systems as described by the above workers is dependent upon the recognition of lateral accretionary bedform sequences, which are not evident in Teapot sediments. Similar fining-upwards cycles have been observed in modern and ancient tidal flat complexes (Reineck and Singh, 1973). These sequences are distinguishable from those formed by migrating channel complexes by the use of primary structure sequences within the beds themselves, and by those of the enclosing beds viewed within the overall sedimentary sequence.

PALEOGEOGRAPHIC MODEL

Paleogeographic models for the Upper Cretaceous strata in the Western Interior have been presented by several authors in the last 15 years. Cretaceous deltaic systems have been presented by Asquith 1974), Rea and Barlow (1975), Curry (1976a), Hubert, Butera, and Rice (1972), Isbell, Spencer, and Sietz (1976), and MacKenzie (1975). Barrier bar settings inbetween deltaic centers have been described by Jacka (1965), Masters (1965 and 1967), Ryer (1977), Shelton (1963 and 1965), and Weimer (1966 and 1975)

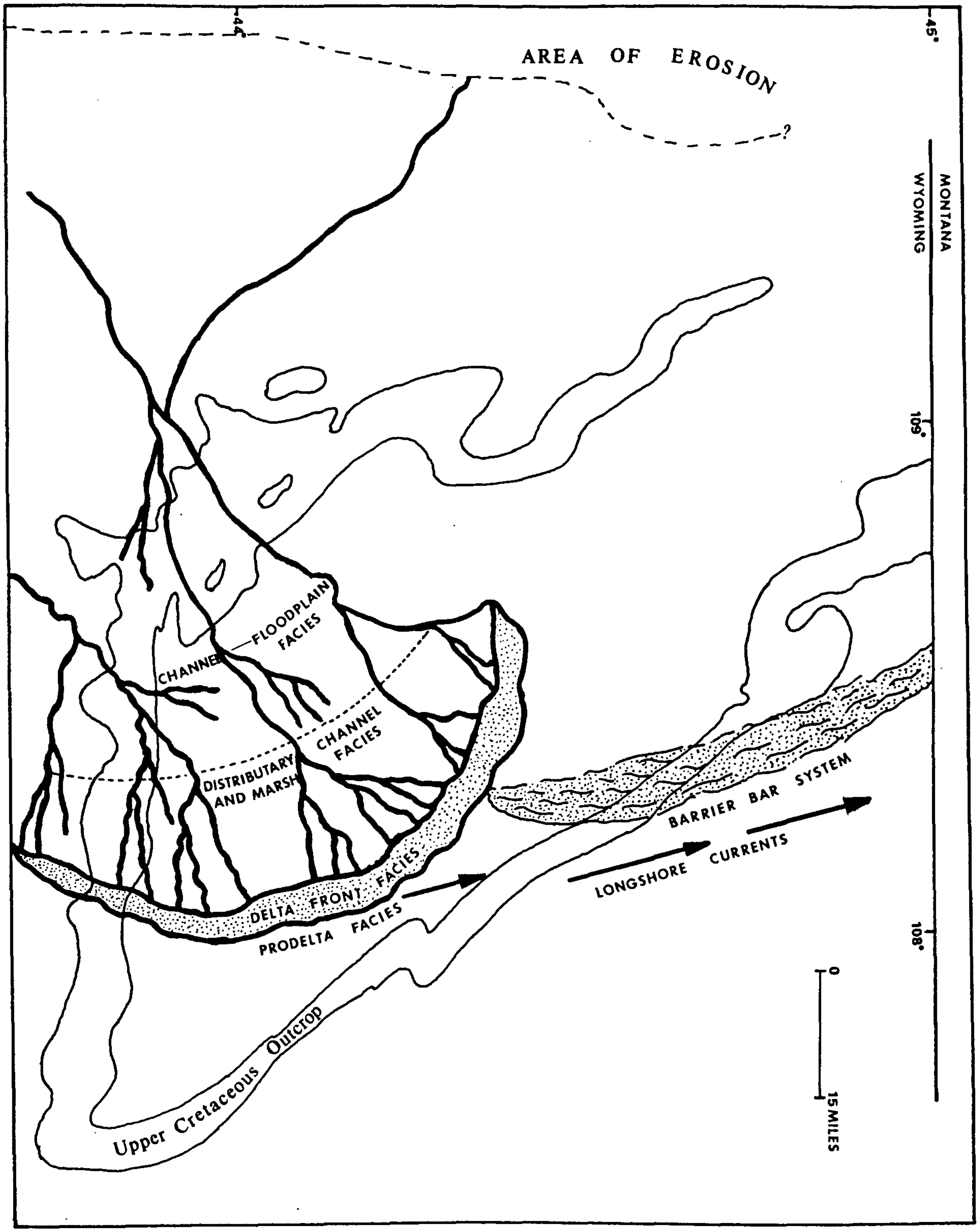
Although deltaic and interdeltaic complexes have been documented separately by the above authors, descriptions of ancient deltaic and barrier bar settings preserved in close proximity are not common in the Upper Cretaceous Western Interior. If wave dominated, high destructive type deltas similar to that presented by Isbell, Spencer, and Sietz (1976) did exist, barrier bar systems would be expected downcurrent from adjacent deltaic systems, provided that an adequate sediment supply existed.

Regional and local lithofacies associations and stratigraphic and paleoenvironmental evidence at this time point to the existence of a high destructive, wave dominated Niger type deltaic complex as illustrated by Allen (1964 and 1965a). Prodeltaic sediments were deposited on an easterly dipping paleoslope in the southern and western areas of the Big Horn Basin (Asquith, 1974, Barlow and Haun, 1975, and MacKenzie, 1975). Throughout the deposition of the Mesaverde, terriginous materials were dispersed downcurrent from the depocenter by longshore currents which resulted in the formation of the barrier island and lagoonal sequence described above (fig. 21).

Within the southwestern and southern part of the Big Horn Basin, the areal distribution of channel-dominated fluvial environments described by MacKenzie (1975) as upper and lower delta plain display arcurate patterns. This pattern of sedimentary environments is comparable to the distribution of fluvial and transitional marine units present in the Erickson Sandstone in southwestern Wyoming (Asquith, 1974). This type of horseshoe shaped plan view of paleoenvironmental patterns is indicative of deltaic sequences, and is thought to be present in the Big Horn Basin area (Asquith, 1974, MacKenzie, 1975). In both the Mesaverde and the Erickson, the arcurate distribution of fluvial sandstones may define the location of the delta more clearly than the seaward buldge in Upper Cretaceous strandlines noted by Gill and Cobban (1973).

Associated with the seaward margins of the delta are prodeltaic and delta front shales and sandstones of the Upper Cody Shale and Eagle formations, (Asquith, 1974, Rea and Barlow, 1975). Subsurface evidence, specifically correlation of time stratigraphic bentonite markers in the upper part of the Cody Shale at Neiber Anticline in the southeastern part of the basin, indicate deposition on an eastward dipping paleoslope. Isopach maps of the upper portion of the Cody and the lower

Figure 21. Paleogeographic setting, _Baculites perplexus_, early form time in the southeastern Big Horn Basin area.


AREA OF EROSION
45°
MONTANA
WYOMING
109°
CHANNEL
FLOODPLAIN FACIES
CHANNEL FACIES
DISTRIBUTARY AND MARSH
BARRIER BAR SYSTEM
LONGSHORE CURRENTS
DELTA FRONT FACIES
PRODELTA FACIES
108°
0
15 MILES
Upper Cretaceous Outcrop

Eagle interval in the same locality show abrupt thickening of time stratigraphic units and basinward thinning, which point towards a prodeltaic sequence. Rapid pinchouts of littoral sandstones with marine shales at Neiber may again be indicative of a prodeltaic slope in the area.

Comparable complex intertonguing relationships between forest and bottomset deltaic sequences have been described by Weimer (1966 and 1975) in the Lewis Shale-Fox hills interval in the Wamsutter Arch area. Complex shoreline trends are shown to be associated with deltaic embayments, and have resulted in reversals of normal Upper Cretaceous lithofacies patterns. Prodeltaic shales thin rapidly eastward, and interfinger with transitional marine sediments along the inferred depositional strike.

Comparable rapid pinchouts and facies changes are present in the Cody-Eagle interval in the study area along the depositional strike. These can be explained by local shoreline embayments associated with a deltaic center to the south. Depositional slopes associated with prodelta areas can easily account for rapid pinchouts in the upper Cody-lower Eagle interval.

Paleocurrent evidence was collected by the writer and by Houston and Murphy (1962 and 1977) along the eastern flank of the Big Horn Basin based upon the thickness and geometry of fossil beach placer sands. These paleo-strandline features indicate a dominate northerly active or storm generated sediment transport direction, parallel to the shoreline trend. These currents are indicated by the presence of the placer deposits themselves, which are formed by storm generated wave processes. These wave and current processes may have been responsible for the northerly dispersal of sand and carbonaceous debris, forming a composite deltaic-barrier island, lagoonal setting, similar to that described for the Almond Formation in the Rock Springs uplift area (Jacka, 1965). Comparable modern deltaic, barrier settings are typified by the Niger River delta where longshore currents have formed a barrier bar down current from the delta (LeBlanc, 1972, Allen, 1965 a,b,c,d and 1970).

It must be stressed at this point that the model proposed here is tentative, and is based on all available evidence at this point in time. Little or no work has been done on the Mesaverde in the northeastern half of the Big Horn Basin, and at present the area is largely unstudied. The existence of deltaic deposits in the area cannot be fully substantiated on the basis of sequences of primary sedimentary structures alone. Lithologic associations typical of prodeltaic, delta front, and delta plain environments of a high destructive type delta are very similar to progradational beach deposits described above and often cannot be differentiated on the basis of bedding sequences (Weimer, 1975). Extensive subsurface correlations and isopach maps of time stratigraphic intervals, are needed to firmly establish a three dimensional framework, from which future interpretations and paleoenvironmental predictions can be made.

SUMMARY

Detailed descriptions of rock body geometry, sequences of primary structures, lithology and fossils have enabled the writer to reconstruct the Campanian paleogeographic setting and paleoenvironments recorded by the Mesaverde Group in the study area. Mesaverde sediments were the site of a laterally extensive mainland and barrier beach complex throughout Campanian time. A high destructive, Niger type deltaic complex located to the south of the research area supplied terrigineous materials which were transported northward by active storm generated longshore currents, forming the adjacent barrier bar association.

A major Upper Cretaceous regressive episode in the Big Horn Basin area initiated the deposition of the Eagle Formation, during which sediments were deposited in a shoreface and offshore-shoreface transition zone. Cyclic shoreline migrations during this time period are recorded as evidenced by complex intertonguing between littoral Eagle sands and Cody shelf shales. These materials were deposited on an easterly dipping prodeltaic paleoslope, which resulted in complex interfingerings between offshore shales and beach sandstones within the upper Cody Shale, Eagle, and Claggett sediments along the inferred depositional strike. Early Campanian cyclic shoreline migrations were interrupted by a rapid transgression of the Late Cretaceous epieric sea, documented by shales and marine fauna contained in the Claggett Formation.

Judith River sediments record a complete regressive sequence with offshore and transitional marine deposits overlain by channel dominated lower deltaic or coastal plain environments. Continuous deposition resulted in the preservation of complete regressional, coarsening upward mainland and barrier beach and lagoonal sequences contained in the lower and middle Judith River Formations. Evidence presented by the writer indicates no major unconformity is present within the Judith River section, throughout the study area. The concept of depositional topography is probably the best explanation for the presence of mutually inclined time surfaces within a conformable stratigraphic sequence.

A brief transgressive phase within the study area is documented by upper Judith River shoreface and foreshore environments preserved in these sediments. Fluvial Teapot sediments record the final regressive episode within the Mesaverde Group, during which complex meandering stream associations dominated sedimentation patterns.

Although a complete regressive beach cycle is preserved in the Group, detailed sampling of key beds within the section reveal few significant vertical or lateral changes in sandstone composition. Shoreface and shelf sediments however, can be differentiated from other sediments in the Mesaverde by the presence of glauconite and dolomite grains.

BIBLIOGRAPHY

Allen, J. R. L. 1964. Sedimentation of the modern delta of the river Niger, West Africa, In Developments in Sedimentology, Deltaic and Shallow Marine Deposits, Elsevier, Amsterdam. 1:27-54.

Allen, J. R. L. 1965a. Coastal geomorphology of eastern Nigeria. Geol. Mijnbouw. 44:1-21.

Allen, J. R. L. 1965b. A review of the origin and characteristics of recent fluvial deposits. Sedimentology. 5:89-191.

Allen, J. R. L. 1965c. Late Quaternary Niger Delta and adjacent area: Sedimentary environments and lithofacies. Am. Assoc. Pet. Geol. Bull. 49:547-600.

Allen, J. R. L. 1965d. Coastal geomorphology of eastern Nigeria--beach-ridge barrier islands and vegetated tidal flats. Geol. Mijnbouw 27:1-21.

Allen, J. R. L. 1970. Sediments of the modern Niger Delta: Summary and review. S.E.P.M. Spec. Pub. 15: In Morgan, J. P. (eds.) Deltaic Sedimentation Soc. Econ. Palentologists and Mineralogists, Tulsa, Oklahoma. 138-151.

Allen, J. R. L., and P. F. Friend. 1968. Deposition of the Catskill facies, Appalachian region, with notes on some other Old Red Sandstone basins. Geol. Soc. Am. Spec. Pap. 106:27-74.

Armstrong, F. C., and S. S. Oriel. 1965. Tectonic development of Idaho-Wyoming thrust belt. Am. Assoc. Pet. Geol. Bull. 4:1184-1224.

Asquith, D. O. 1974. Sedimentary models, cycles, and deltas, Upper Cretaceous, Wyoming. Am. Assoc. Pet. Geol. Bull. 58:2274-2283.

Barnett, V. H. 1913. Possibilities of oil in the Big Muddy Dome, Converse and Natrona counties, Wyoming. U. S. Geol. Surv., Bull. 581:105-117.

Berg, R. T. 1975. Depositional environment of the Upper Cretaceous Sussex Sandstone, Horse Creek Field, Wyoming. Am. Assoc. Pet. Geol. Bull. 7:2099-2110.

Berg, R. G. and D. K. Davies. 1968. Origin of Lower Cretaceous Muddy Sandstone at Bell Creek Field, Montana. Am. Assoc. Pet. Geol. Bull. 52:1888-1898.

Bernard, H. A., C. F. Major, and B. S. Parrot. 1970. The Galveston barrier island, and environs-a model for predicting reservoir occurrence and trend. Gulf Coast Assoc. Geol. Soc., Trans. 9:221-224.

Bernard, H. A., R. J. LeBlanc, and C. F. Major. 1962. Recent and Pliestocene geology of southeast Texas, field excursion no. 3. Pages 175-224 In Geology of Gulf Coast and Central Texas and Guidebook of Excursions. Geol. Soc. Am., Houston, Texas.

Bernard, H. A., and C. F. Major, Jr. 1963. Recent meander belt deposits of the Brazos River: An alluvial sand model. Am. Assoc. Pet. Geol. 47:350-351.

Bowen, C. F. 1919. Gradations from continental to marine conditions in central Montana during the Eagle and Judith River Epochs. U. S. Geol. Surv., Prof. Pap. 125:11-21.

Cameron, B., and V. S. Hardarshar. 1977. Skewness as a paleoenvironmental indicator. J. Sediment. Petrol. 47:784-793.

Campbell, C. V. 1968. Lamina, laminaset, bed, and bedset. Sedimentology 8:7-26.

Clifton, H. E. 1969. Beach lamination-nature and origin. Mar. Geol. 7:553-559.

Clifton, H. E. 1976. Wave-formed sedimentary structures, a conceptual model. S. E. P. M. Spec. Pub. 24. In R. A. Davis and K. L. Ethington, (eds.). Beach and Nearshore Sedimentation. Soc. Econ. Paleontologists and Mineralogists, Tulsa, Oklahoma.

Clifton, H. E., R. E. Hunter, and R. L. Phillips. 1971. Depositional structures and processes in the nonbarred high-energy nearshore. J. Sediment. Petrol. 41:651-670.

Cobban, W. A. 1969. The Late Cretaceous ammonite *Scaphites hippocrepis* (DeKay) in the Western Interior of the United States. U. S. Geol. Surv., Prof. Pap. 619:29p.

Cobban, W. A., and J. B. Reeside, Jr. 1952. Correlation of the Cretaceous formations of the Western Interior of the United States. Geol. Soc. Am. Bull. 63:1011-1044.

Curry, W. H., III. 1976a. Late Cretaceous Teapot Delta of southern Powder River Basin, Wyoming. Wyo. Geol. Assoc. Guidebook. Ann. Field Conf. 28:21-28.

Curry, W. H., III. 1976b. Type section of the Teapot Sandstone. Wyo. Geol. Assoc. Guidebook, Ann. Field Conf. 28:29-32.

Curry, W. H., and G. C. Crews. 1976. Field trip road log. Upper Cretaceous sandstones Casper-Salt Creek-Tisdale Mountain. Wyo. Geol. Assoc. Guidebook. Ann. Field Conf. 28:312-322.

Davidson, R. D., and B. Greenwood. 1974. Bedforms and structures associated with bar topography in shallow water wave environment, Kouchibougac Bay, New Brunswich, Canada. J. Sediment. Petrol. 44:698-704.

Davidson, R. D., and B. Greenwood. 1976. Facies relationships on a barred coast, Kouchibougac Bay, New Brunswick, Canada. S. E. P. M. Spec. Pub. 24. In Davis, R. A. and K. L. Ethington, (eds.). Beach and Nearshore Sedimentation. Soc. Econ. Paleontologists and Mineralogists, Tulsa, Oklahoma.

Davies, D. K., F. G. Ethridge, and R. R. Berg. 1971. Recognition of barrier environments. Am. Assoc. Pet. Geol. Bull. 55:550-565.

Davis, T. M. J. 1976. An environmental interpretation of the Upper Cretaceous Shannon Sandstone, Heldt Draw Field, Wyoming. Wyo. Geol. Assoc. Guidebook. Ann Field Conf. 28:125-139.

Dickenson, K. A., H. L. Berryhill, and C. W. Holmes. 1972. Criteria for recognizing ancient barrier coastlines. S. E. P. M. Spec. Pub. 16. In Davis, R. A. and R. L. Ethington, (eds.). Recognition of Ancient Sedimentary Environments. Soc. Econ. Paleontologists and Mineralogists, Tulsa, Oklahoma.

Douglass, W. B., Jr., and T. R. Blazzard. 1961. Facies relationships of Blair, Rock Springs, and Erickson Formations of the Rock Springs Uplift and Washakie Basin. Wyo. Geol. Assoc. Guidebook. Ann. Field Conf. 16:81-86.

Eldridge, G. H. 1889. Some suggestions for the methods of grouping the formations of the middle Cretaceous and the employment of an additional term in its nomenclature. Am. Jour. Sci. 38:313-321.

Fairbridge, R. W. 1967. Phases of Diagenesis and Authigenesis. In Developments in Sedimentology. Elsevier, Amsterdam 8:19-89.

Fisher, C. A. 1908. Southern extension of the Kootenai and Montana coal-bearing formations in Northern Montana. Econ. Geol. 3:77-99.

Folk, R. L. 1954. The distinction between grain size and mineral composition in sedimentary rock nomenclature. J. Geol. 62:344-359.

Folk, R. L. 1974. Petrology of Sedimentary Rocks. Hemphill Publishing Co., Austin, Texas.

Folk, R. L., and W. C. Ward. 1957. Brazos River bar: A study in the significance of grain-size parameters. J. Sediment. Petrol. 27:3-26.

Friedman, G. M. 1961. Distinction between dune, beach, and river sands from their textural characteristics. J. Sediment. Petrol. 31:514-529.

Friedman, G. M. 1967. Dynamic processes and statistical parameters compared for size frequency distribution of beach and river sands. J. Sediment. Petrol. 37:327-354.

Gagliano, S. M., and W. G. McIntire. 1968. Delta components in recent and ancient deltaic sedimentation: A comparison. Louisiana State Univ. Coastal Studies Inst.

Gardner, D. E. 1955. Beach-sand heavy mineral deposits of eastern Australia. Aust., Bur. Miner. Resour., Geol. Geophys., Bull. 28:103p.

Gill, J. R., and W. A. Cobban. 1966. Regional unconformity in Late Cretaceous, Wyoming. U. S. Geol. Surv. Prof. Pap. 550:20-27.

Gill, J. R., and W. A. Cobban. 1973. Stratigraphy and geologic history of the Montana Group and equivalent rocks, Montana, Wyoming, and North and South Dakota. U. S. Geol. Surv., Prof. Pap. 776:37p.

Harms, J. C., and R. K. Fahnstock. 1965. Stratification, bed forms, and flow phenomena. Pp. 84-115. In G. V. Middleton (ed.). Primary sedimentary structures. Spec. Pub. 12. Soc. Economic Paleontologists and Mineralogists, Tulsa, Oklahoma.

Harms, J. C., J. B. Southard, D. R. Spearing, and R. G. Walder. 1975. Depositional environments as interpreted form primary sedimentary structures and stratification sequences. Soc. Econ. Paleontologists and Mineralogists, Short Course No. 2, Dallas.

Hewett, D. F. 1914. The Shoshone River section, Wyoming. U. S. Geol. Surv., Bull. 541:89-113.

Hewett, D. F. 1926. Geology and oil and coal resources of the Oregon Basin, Meeteetse, and Grass Creek Basin Quadrangles, Wyoming. U. S. Geol. Surv., Prof. Pap. 145:111p.

Holmes, W. H. 1877. Report on the San Juan District, Colorado. U. S. Geol. and Geog. Surv. Terr. 9th Ann. Rept., 237-276.

Houston, R. S., and J. F. Murphy. 1962. Titaniferous black sandstone deposits of Wyoming. Wyo. Geol. Surv. Bull. 49:120p.

Houston, R. S., and J. F. Murphy. 1977. Depositional environment of Upper Cretaceous black sandstones of the Western Interior. U. S. Geol. Surv., Prof. Pap. 994A:29p.

Howard, J. D. 1972. Trace fossils as criteria for recognizing shorelines in the stratigraphic record. S. E. P. M. Spec. Pub. 16. In Davis, R. A. and R. L. Ethington, (eds.). Beach and Nearshore Sedimentation. Soc. Econ. Paleontologists, Tulsa, Oklahoma.

Hoyt, J. H. 1962. High-angle beach stratification, Sapelo Island, Georgia. J. Sediment. Petrol. 32:309-311.

Hoyt, J. H. 1967. Barrier island formation. Geol. Soc. Am. Bull. 78:1125-1136.

Hoyt, J. H., and V. J. Henry. 1967. Influence of island migration on barrier island sedimentation. Geol. Soc. Am. Bull. 78:77-86.

Hoyt, J. H., and R. J. Weimer. 1965. Origin and significance of Ophiomorpha (Halymites) in the Cretaceous of the Western Interior. Wyo. Geol. Assoc. Guidebook. Ann. Field Conf. 19:203-208.

Hubert, J. F., J. G. Butera, and R. F. Rice. 1972. Sedimentology of the Upper Cretaceous Cody-Parkman Delta southwestern Powder River Basin, Wyoming. Geol. Soc. Am. Bull. 83:1649-1670.

Isbell, E. B., C. W. Spencer, and T. Sietz. 1976. Petroleum geology of the Well Draw Field, Converse County, Wyoming. Wyo. Geol. Assoc. Guidebook Ann. Field Conf. 28:165-175.

Jacka, A. D. 1965. Depositional dynamics of the Almond Formation, Rock Springs Uplift, Wyoming. Wyo. Geol. Assoc. Guidebook. Ann. Field Conf. 19:81-100.

Johnson, D. W. 1919. Shore Processes and Shoreline Development. Wiley and Sons, New York.

Jopling, A. W., and R. G. Walker. 1968. Morphology and origin of ripple-drift cross-lamination with examples from the Pleistocene of Massachusetts. J. Sediment. Petrol. 38:971-984.

LeBlanc, R. J. 1972. Geometry of sandstone reservoir bodies. Am. Assoc. Pet. Geol. Mem. 18:133-190. In Cook, T. D. (ed.) Underground Waste Management and Environmental Implications. Mem. 18. Am. Assoc. Pet. Geol., Houston, Texas.

Leopold, L. B., and Wolman. 1957. River channel patterns: Braided, meandering, and straight. U. S. Geol. Surv., Prof. Pap. 282B.

Lewis, J. L. 1961. The stratigraphy and depositional history of the Almond Formation in the Great Divide Basin, Sweetwater County, Wyoming. Wyo. Geol. Assoc. Guidebook. Ann. Field Conf. 16:67-75.

Lupton, C. T. 1916. Oil and gas near Basin, Big Horn County, Wyoming. U. S. Geol. Surv., Bull. 621:157-190.

MacKenzie, M. 1975. Stratigraphy of the Mesaverde Group, south part of the Big Horn Basin. Ph.D. thesis. Tulane University, New Orleans.

McGookey, D. P. 1972. Cretaceous System. Pages 190-328 In Geologic Atlas of the Rocky Mountain Region, United States. Rocky Mt. Assoc. Geol., Denver.

McGowen, H. H., and Garner. 1970. Physiographic features and stratification types of coarse-grained point bars: Modern and ancient examples. Sedimentology 14:77-111.

Mason, C. C., and R. L. Folk. 1958. Differentiation of beach, dune, and aeolian flat environments by size analysis, Mustang Island, Texas. J. Sediment. Petrol. 28:211-226.

Masters, C. D. 1965. Sedimentology of the Mesaverde Group and the upper part of the Mancos Formation, northwestern Colorado. Ph.D. thesis. Yale University.

Masters, C. D. 1967. Use of sedimentary structures in determination of depositional environments, Mesaverde Formation, Williams Fork Mountains, Colorado. Am. Assoc. Pet. Geol. Bull. 51:2033-2043.

Meek, F. B., and V. Hayden. 1856. Descriptions of a new fossil species of Mollusca; together with a complete catalogue of all the remains of invertebrata hitherto described and identified from the Cretaceous and Tertiary formations of Nebraska Terr. Phila. Acad. Nat. Sci. Proc. 8:265-286.

Merewether, E. A., W. A. Cobban, and C. W. Spencer. 1976. The Upper Cretaceous Frontier Formation in the Kaycee-Tisdale Mountain area, Johnson County, Wyoming. Wyo. Geol. Assoc. Guidebook. Ann. Field Conf. 28:33-45.

Miall, J. D. 1977A. A review of the Braided River environment. Earth Sci. Rev. 13:1-62.

Miall, J. D. 1977b. Fluvial Sedimentology. Can. Soc. of Pet. Geol., Calgary, Alberta Canada.

Miller, D. N., J. A. Barlow, and J. D. Haun. 1965. Stratigraphy and petroleum potential of latest Cretaceous rocks, Big Horn Basin, Wyoming. Am. Assoc. Pet. Geol. Bull. 49:277-285.

Nordstrom, K. F. 1977. The use of grain size statistics to distinguish between high and moderate energy beach environments. J. Sediment. Petrol. 47:1287-1297.

Passega, R. 1957. Texture as a characteristic of clastic deposition. Am. Assoc. Pet. Geol. Bull. 41:1952-1984.

Passega, R. 1964. Grain-size represented by CM patterns as a geological tool. J. Sediment. Petrol. 34:830-847.

Passega, R., and Byramjee. 1969. Grain-size image of clastic deposits. Sedimentology. 13:233-252.

Peterson, F. 1969. Cretaceous sedimentation and tectonism in the southeastern Kasparoents region, Utah. U. S. Geol. Surv., open file rep. 259p.

Rea, D. B. and J. A. Barlow. 1975. Upper Cretaceous and Tertiary rocks, northern part of the Big Horn Basin, Wyoming and Montana. Wyo. Geol. Assoc. Guidebook Ann. Field Conf. 27:63-71.

Reineck, H. F. and Singh. 1973. Depositional Sedimentary Environments. Springer-Verlag, New York.

Reineck, H. E. and F. Wunderlich. 1968B. Classification and origin of flaser and lenticular bedding. Sedimentology. 11:99-104.

Rich, E. I. 1958. Stratigraphic relation of latest Cretaceous rocks in parts of the Powder River, Wind River, and Big Horn Basins, Wyoming. Am. Assoc. Pet. Geol. Bull. 42:2424-2443.

Ross, C. S., and D. F. Kerr. 1931. The Kaolin minerals. U. S. Geol. Surv., Prof. Pap. 165E:151-170.

Royse, C. F., Jr. 1968. Recognition of fluvial environments by particle-size characteristics. J. Sediment. Petrol. 38:1171-1177.

Ryer, J. A. 1977. Patterns of Cretaceous shallow-marine sedimentation. Coalville and Rockport areas, Utah. Geol. Soc. Am. Bull. 88:177-188.

Sabins, F. F. 1962. Grains of detritial, secondary, and primary dolomite from Cretaceous strata of the Western Interior. Geol. Soc. Am. Bull. 73:1183-1196.

Sabins, F. F. 1965. Review of petrographic criteria for Cretaceous depositional environments. Wyo. Geol. Assoc. Guidebook Ann. Field Conf. 19:27-34.

Schultz, L. G. 1962. Quantitative interpretation of mineralogical composition from X-ray and chemical data for the Pierre Shale. U. S. Geol. Surv., Prof. Pap. 391C:29p.

Scott, T. 1954. Sand movements by waves. U. S. Beach Erosion Board Tech. Memo. 48:37p.

Sears, A. D., C. B. Hunt, and T. A. Hendricks. 1941. Transgressive and regressive Cretaceous deposits in southern San Juan Basin, New Mexico. U. S. Geol. Surv., Prof. Pap. 193F:101-122.

Severn, W. P. 1961. General stratigraphy of the Mesaverde Group, Big Horn Basin, Wyoming. Wyo. Geol. Assoc. Guidebook Ann. Field Conf. 16:195-199.

Shelton, J. W. 1963. Measured sections of sandstones and shales. Am. Assoc. Pet. Geol. Bull. 47:2047-2051.

Shelton, J. W. 1964. Authigenic Kaolinite in sandstones. J. Sediment. Petrol. 34:102-111.

Shelton, J. W. 1965. Trend and genesis of the lowermost sandstone unit of the Eagle Sandstone at Billings, Montana. Am. Assoc. Pet. Geol. Bull. 49:1385-1397.

Simons, D. B., and E. V. Richardson. 1961. Forms of bed roughness in alluvial channels. Am. Soc. Civ. Eng., Proc. 87:87-105.

Simons, D. B., and E. V. Richardson. 1962. Discussion of forms of bed roughness in alluvial channels. Am. Soc. Civ. Eng., Proc. 88:23-243.

Simons, D. B., E. V. Richardson, and C. F. Nordin, Jr. 1965. Sedimentary structures generated by flow in alluvial channels. S. E. P. M. Spec. Pub. 12 In Middleton, G. V. (eds.) Primary Sedimentary Structures and their Hydrodynamic Interpretation. Soc. Econ. Paleontologists and Mineralogists, Tulsa, Oklahoma.

Sorby, H. C. 1859. On the structures produced by the currents present during deposition of stratified rocks. Geologist. 2:137-147.

Spieker, E. M. 1949. Sedimentary facies and associated diastrophism in the Upper Cretaceous of central and eastern Utah. Geol. Soc. Am. Bull. Mem. 39:55-82.

Stanton, T. M., and J. B. Hatcher. 1903. The stratigraphic position of the Judith River beds and their correlation with the Belly River beds. Science. 18:211-212.

Stanton, T. W., J. B. Hatcher, and J. H. Knowlton. 1905. Geology and paleontology of the Judith River beds. U. S. Geol. Surv., Bull. 257:128p.

Stebinger, E. 1914. Stratigraphy of the Montana Group (Upper Cretaceous) in northwestern Montana, and its relation to the Belly River beds and Judith River formation. J. Wash. Acad. Sci. 4:383-384 (abstr.).

Thompson, W. D. 1937. Original structures of beaches, bars, and dunes. Geol. Soc. Am. Bull. 48:723-752.

Toots, H. 1961. Beach indicators in the Mesaverde Formation. Wyo. Geol. Assoc. Guidebook Ann. Field Conf. 16:165-170.

Visher, G. S. 1972. Environmental Criteria - Their Use and Misuse. Trans. G. H. Coast Assoc. Geol. Soc. 22:67-72.

Washburne, C. W. 1907. Coal Fields of the northeastern side of the Big Horn Basin, Wyoming, and Bridger, Montana. U. S. Geol. Surv., Bull. 341:165-170.

Weaver, C. E. 1956. The distribution and identification of mixed-layer clays in sedimentary rock. Am. Mineral. 56:202-215.

Weaver, C. E. 1961. Clay mineralogy of the Late Cretaceous rocks of the Washakie Basin, Wyoming. Wyo. Geol. Assoc. Guidebook Ann. Field Conf. 16:148-154.

Weed, W. H. 1889. Description of the Fort Benton Quadrangle, Montana. U. S. Geol. Surv., Atlas Fort Benton Folio 55:7p.

Weimer, R. J. 1961. Uppermost Cretaceous rocks in central and southern Wyoming and northwestern Colorado. Wyo. Geol. Assoc. Guidebook Ann. Field Conf. 16:17-29.

Weimer, R. J. 1966. Time-stratigraphic analysis and petroleum accumulations, Patrick Draw Field, Sweetwater County, Wyoming. Am. Assoc. Pet. Geol. Bull. 50:2150-2175.

Weimer, R. J. 1970. Rates of deltaic sedimentation and intra-basin deformation, Upper Cretaceous of the Rocky Mountain region. S. E. P. M. Spec. Pub. 15. In Morgan, J. F. (ed.). Deltaic Sedimentation Modern and Ancient. Soc. Econ. Paleontologists and Mineralogists, Tulsa, Oklahoma.

Weimer, R. J. 1975. Deltaic and Shallow Marine Sandstones: Sedimentation, Tectonics, and Petroleum Occurrences. Am. Assoc. Pet. Geol. Continuing Education Course Note Series 2. 167pp.

Weimer, R. J., and J. H. Hoyt. 1964. Burrows of Callainassa major (Say) geologic indicators of littoral and Shallow neitic environments. J. Paleontol. 38:711-767.

Williams, P. F., and B. R. Rust. 1969. The sedimentology of a braided river. J. Sediment. Petrol. 39:649-679.

Wilson, M. D., and E. D. Pittman. 1977. Authigenic clays in sandstones: Recognition and influence on reservoir properties and paleoenvironmental analysis. J. Sediment. Petrol. 47:3-32.

Woodruff, E. G. 1907. Coal fields of the southwest side of the Big Horn Basin, Wyoming. U. S. Geol. Surv., Bull. 341:200-219.

Zapp, A. D. 1962. Some late Cretaceous strandlines in southern Wyoming. U. S. Geol. Surv., Prof. Pap. 540D:52-55.

Zapp, A. D., and W. A. Cobban. 1960. Some Late Cretaceous strandlines in northwestern Colorado and northeastern Utah. U. S. Geol. Surv., Prof. Pap. 400B:246-249.

APPENDIX

Plate 1

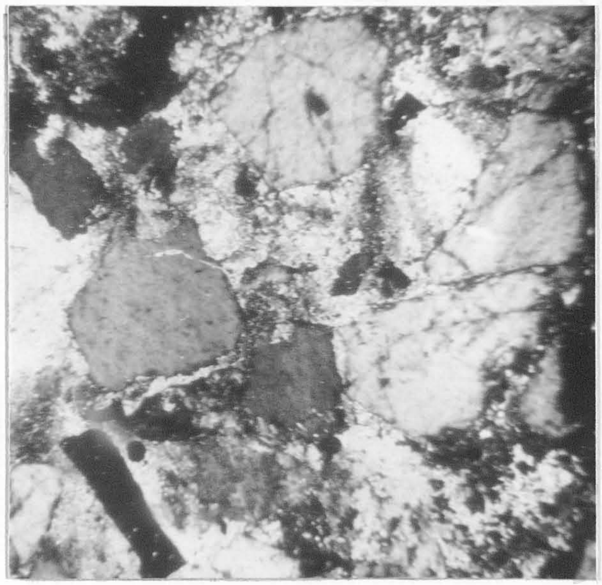
1 - Thin sections of Mesaverde sandstones.

1A - Replacement textures of framework grains by calcite.

1B - Detrital Dolomite - note aggregate of well-rounded grains.

1C - Grains of detrital and secondary dolomite.

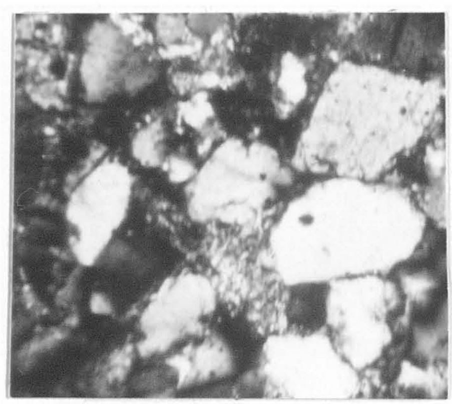
1D - Euhedral secondary dolomite rhombs.



A

B

C



D

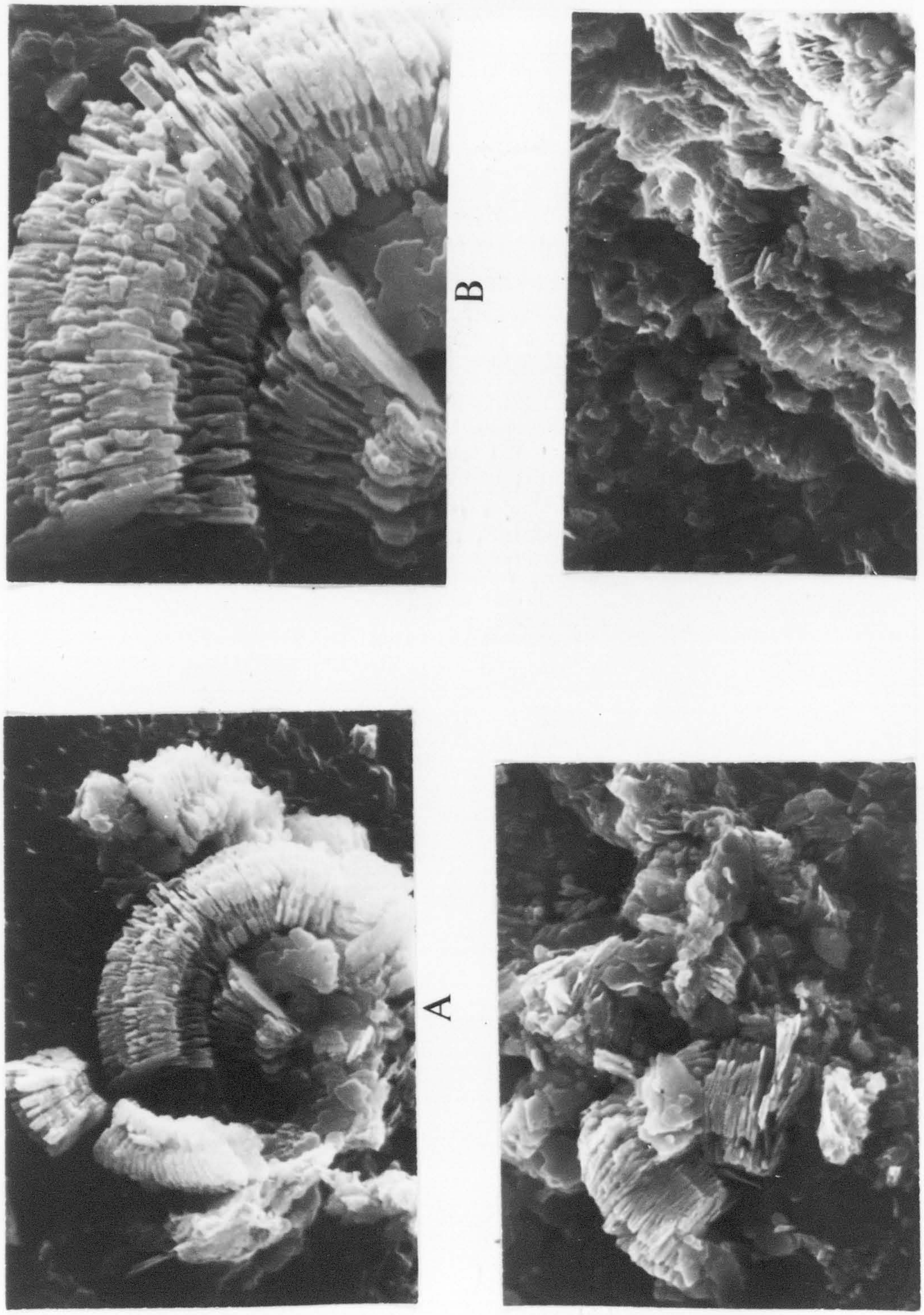
<u>Plates 2A through 2D</u>

<u>2A</u> - Delicate Vermicular, pseudohexagonal Kaolinite books, in pore space 4,500 x

<u>2D</u> - Close-up of plate 2A, 10,000 x -

<u>2C</u> - Broken authigenic kaolinite plates in sandstone pore - 3,000 x

<u>2D</u> - Authigenic Kaolinite books, (right), and clots of authigenic kaolinite filling + lining pore space (left) 3,000 x


A
B
C
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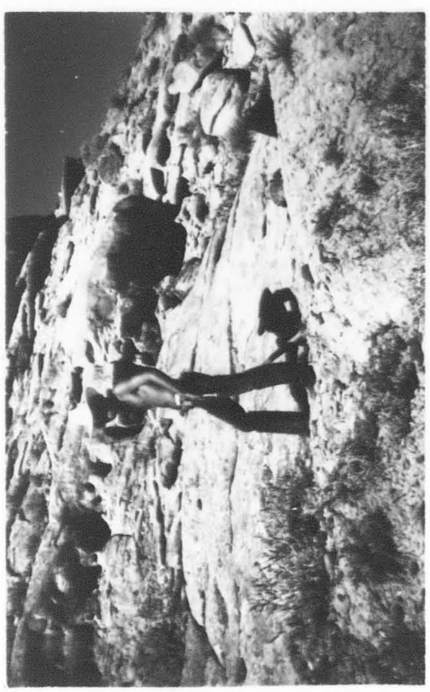
<u>Plates 3A through 3D</u>

<u>3A</u> - Resistant lower Eagle Formation Hogback. Exposed at Little Dry Creek, near Greybull.

<u>3B</u> - Erosional scour contact in upper Eagle Sands at Dry Creek.

<u>3C</u> - Contact between resistant Eagle Hogback and Claggett Shale Exposed west of Alkali Anticline along Alkali Creek.

<u>3D</u> - Contact between resistant rust-colored Teapot sediments and dark banded, carbonaceous Meeteetse sediments.



A

B

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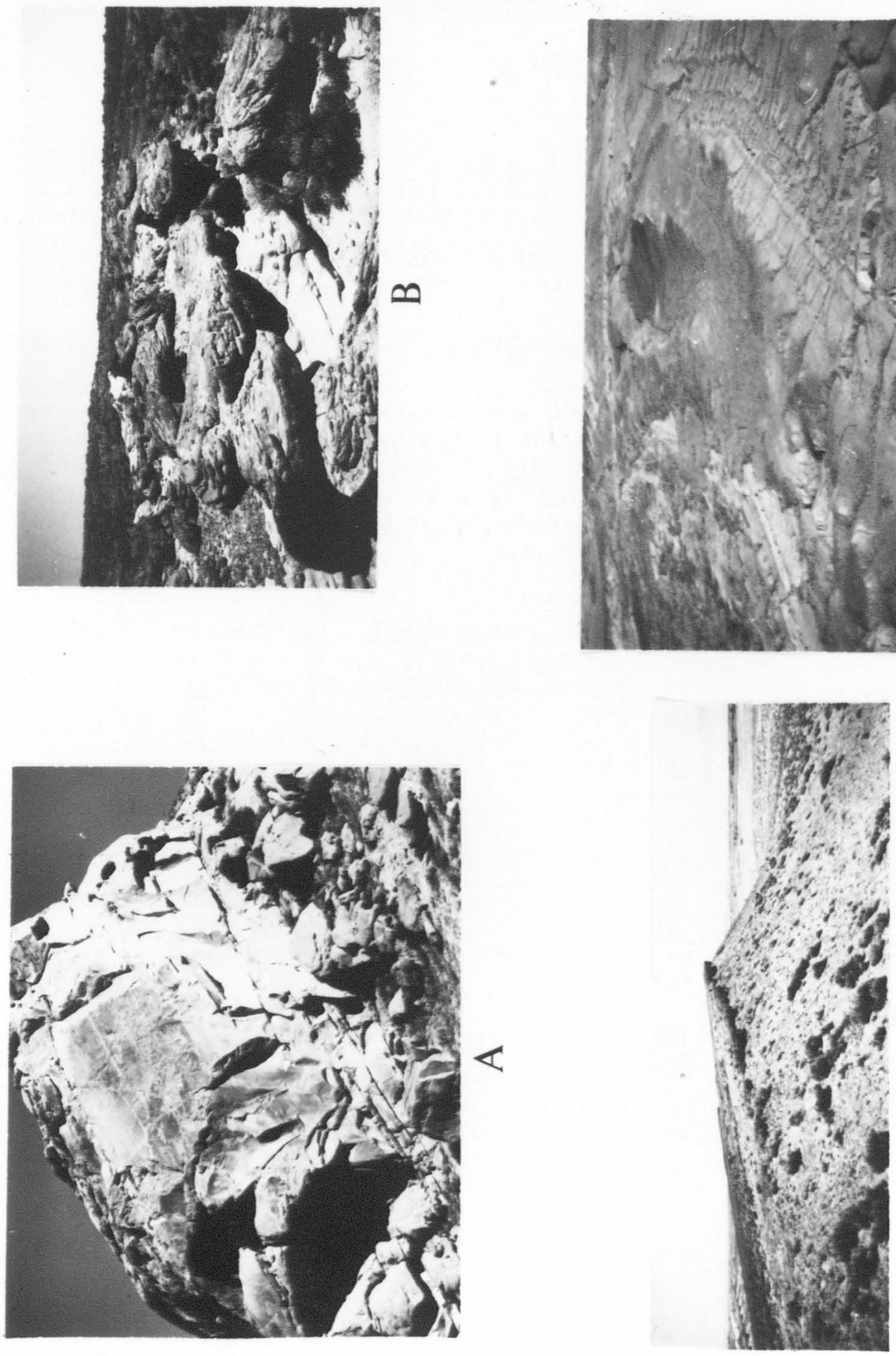
D

Plates 4A through 4D

4A - Massive 90 foot thick barrier is land sand unit in Lower Judith River formation.

4B - Ferriginous sandstone concretions. Exposed in the Teapot Sandstone Member.

4C and D - Titaniforous black sandstone deposit. Exposed six miles south of Lovell, Wyoming.


A
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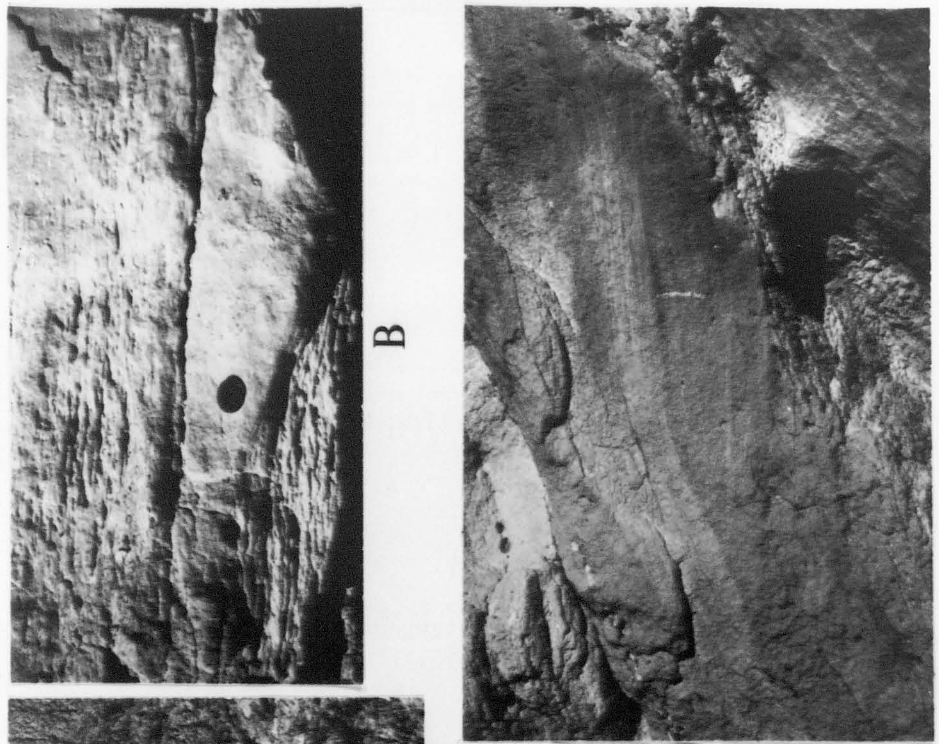
Plates 5A through 5D

5A - Interbedded carbonaceous mudstone and sandstones (light) within the Middle Judith River. Exposed along the Greybull River Road.

5B - Erosional contact of sandstone unit in Upper Judith River west of Alkali Anticline.

5C - Massive Teapot cliff (80 feet). Exposed along the Greybull River Road.

5D - Highly erosional scoured contact at base of trough cross bedded unit. Within the Teapot Sandstone Member.

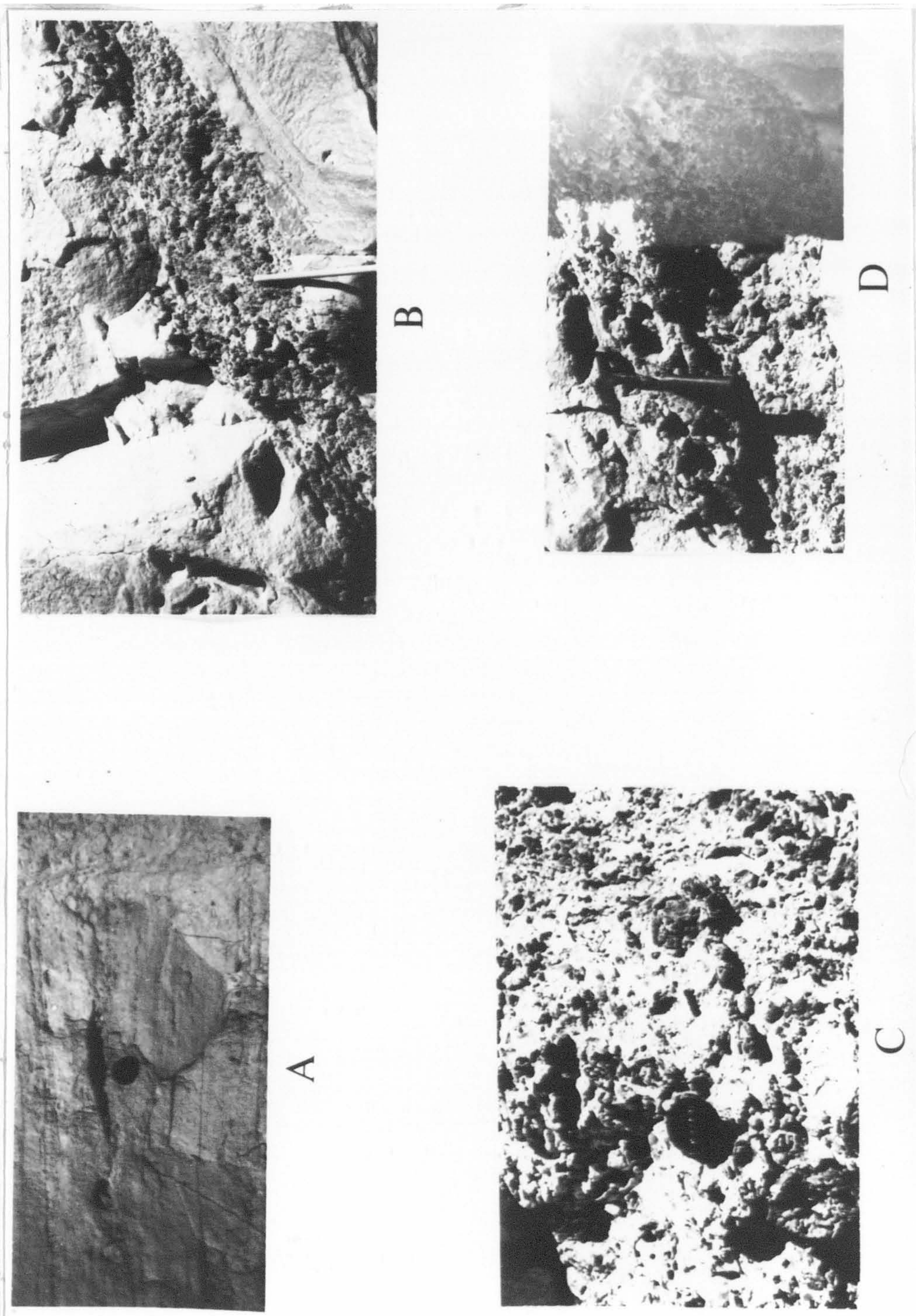


A B C D

Plates 6A through 6D

6A - Carbonaceous mudstone lenses in Teapot Sandstone Member.

6B-D - Lenses of Ironstone concretions contained within the Teapot; note scoured based contact in plate 6D.


A
B
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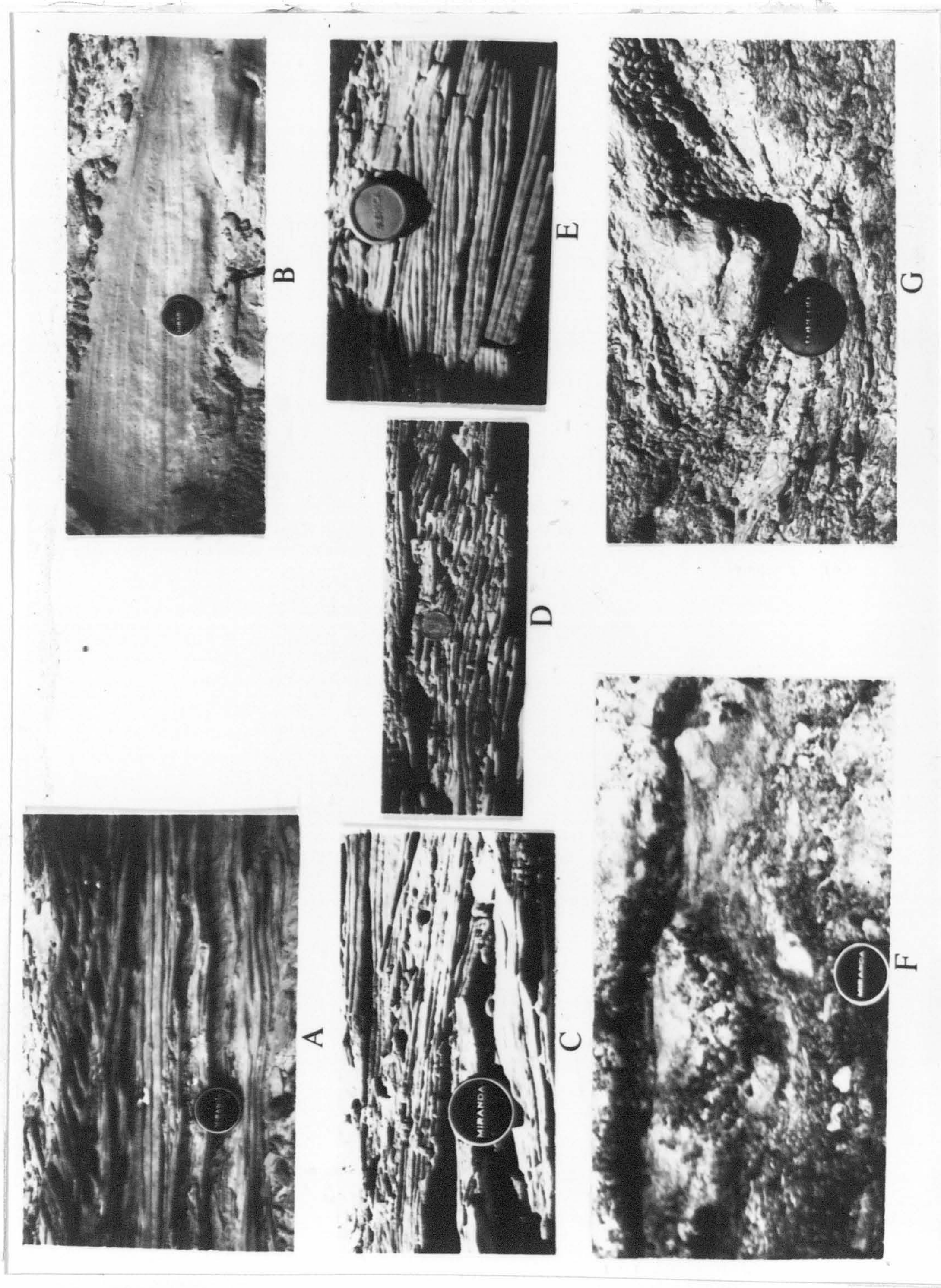
Plates 7A through 7G

7A - Ripple-drift sequences. Exposed in basal Eagle near the south nose of Alkali Anticline.

7B - Hummocky cross stratification in basal Judith River at Little Dry Creek.

7C - Bi-polar ripple-drift sets resulting from tidal currents moving over submarine bar.

7D and E - Non-erosional type ripple drift, the result of gentle rip currents moving over submarine bars.


A
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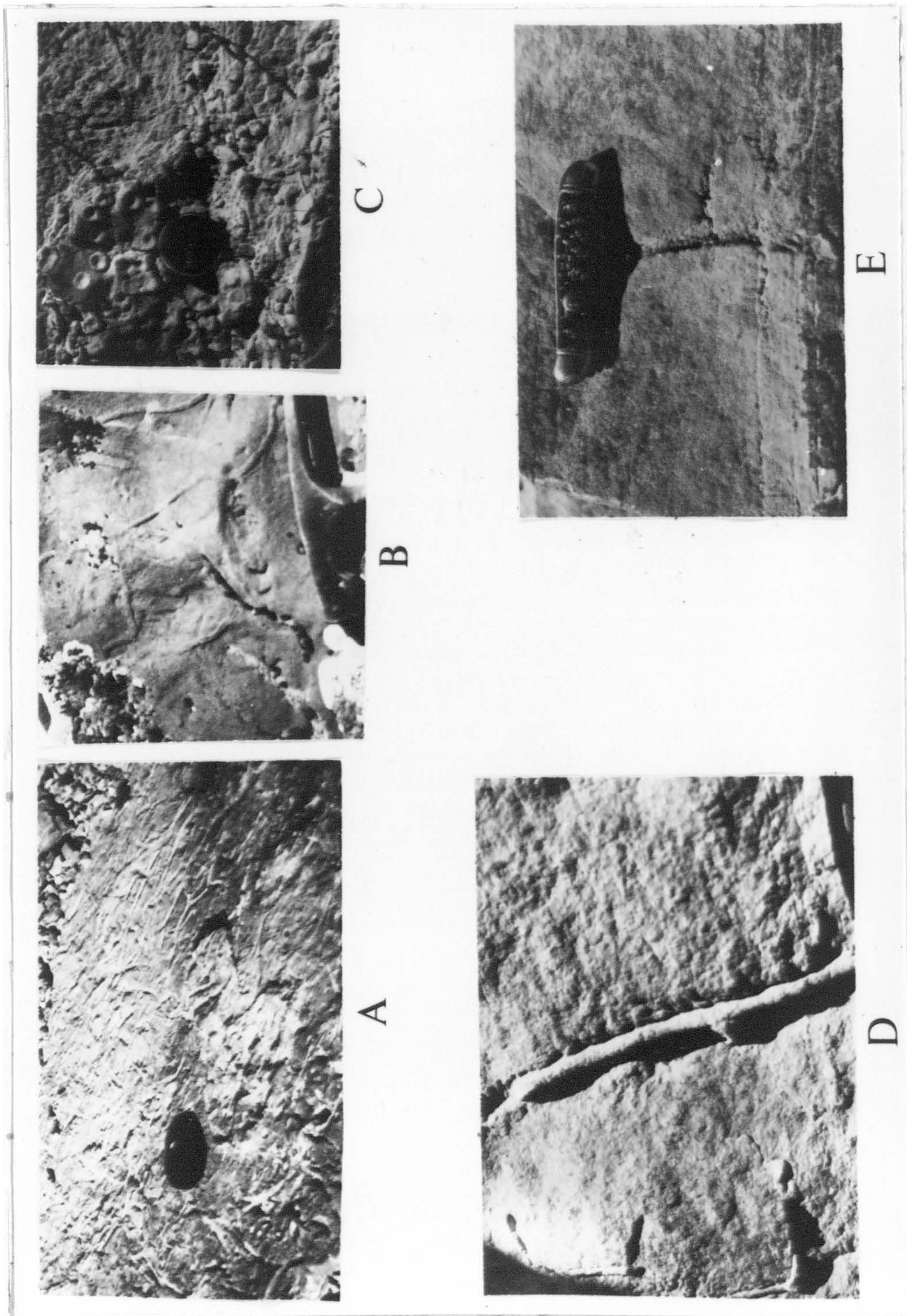
G

Plates 8A through 8C

8A and B - Snail trails concentrated along bedding planes.

8C - Bedding plane expression of Asterosoma. Exposed at Little Dry Creek.

8D and E - *Ophiomorpha sp*. note typical corn-cob structure and u-shaped branching burrows.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