The sedimentology of Camas Prairie basin and its significance to the Lake Missoula floods

James C. Lister

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THE SEDIMENTOLOGY OF CAMAS PRAIRIE BASIN AND ITS SIGNIFICANCE TO THE LAKE MISSOULA FLOODS

by

James C. Lister

B.S., University of South Carolina, 1975

Presented in partial fulfillment of the requirements for the degree of Master of Science

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Chairman, Board of Examiners

Dean, Graduate School

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ABSTRACT

Lister, James C., M.S., Spring, 1981 Geology

The Sedimentology of Camas Prairie Basin and its Significance to the Lake Missoula Floods

Director: Don Winston

The Camas Prairie, an intermontane basin of western Montana, is the site of extremely well preserved Glacial Lake Missoula flood deposits, offering an excellent opportunity to study not only the sedimentology of large flood deposits, but the stratigraphy of the basin as well. The sedimentology and morphology of the flood deposits were analyzed and the findings were compared to Baker's (1973) study of eastern Washington. The stratigraphy of the basin was studied to provide information regarding how many flood events Glacial Lake Missoula (Pardee, 1910) produced. All of the findings were used to explain the recorded flood event in terms of hydraulics.

The stratigraphy of the basin can be divided into four units. From the surface down, they are an uppermost youngest section of Pleistocene varves, an older flood-deposited unit of poorly sorted Pleistocene gravels, a still older section of Tertiary volcanic agglomerate, and finally Precambrian Prichard Formation and Ravalli Group argillites which form the bedrock of the basin.

Profiles of the large scale asymmetric bars were measured and the results were analyzed. The height and chord of the bars decreases downvalley as a polynomial function of distance. An estimated average rate of decrease for each is 0.7 m/km for height, and 15.0 m/km for chord. Their vertical form index (VFI) ranges from 7 to 44. Crossbedding is moderately high angle planar tabular at dips between 19 and 26 degrees. Sorting is poor and the sediments are immature.

The variability of chord to height of the bars is described by $H = 0.144V_0^{0.72}$ and is comparable to Allen's (1968) data on sand-sized largescale ripples. The present gravel of the bars decreases from nearly 100 percent near the source, to 42 percent 5 kilometers downstream.

Antidune deposits and evidence of kolking in each of the passes, succeeded by lower flow regime bars of decreasing size and percent gravel, all indicate steadily decreasing stream power. Cross strata intercalated with mud and silt near the downstream end of the basin complete the sequence, and are indicative of slack water, low energy conditions.

Finally, the Pleistocene stratigraphy of the basin indicates a minimum of two lake stands, separated by one major flood event. Several, earlier floods could have preceded the one recorded in the basin, but all must have been lesser because their deposits were completely swept away leaving no reactivation surfaces within the bars. Likewise, there could have been numerous smaller floods post dating this one.
ACKNOWLEDGMENTS

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Finally, I owe a great deal of gratitude to my wife, Jeannie, and to the staff of the Geology Department, including Don, Bob, Shirley and Katie, for their patience and encouragement.
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CHAPTER I
INTRODUCTION

Ever since Pardee (1942) first recognized that the hills of Camas Prairie represented giant ripples formed by the Glacial Lake Missoula Flood (Pardee, 1910), the Camas Prairie has been of great interest to geologists and hydrologists. Pardee believed that the giant bedforms of the valley floor and scour features in each of the four passes were the result of a lowering of the lake level over Camas Prairie by the sudden failure of an ice dam in Idaho which held Glacial Lake Missoula.

The failure first drained the arm of the lake nearest the dam, flooding the Channeled Scablands area of eastern Washington. Flow from the remainder of the lake was repeatedly impeded by a succession of narrows upstream (Fig. 1). At each location, flow constricted created steep gradients and strong currents through each. In this manner, the water level over Camas Prairie basin was eventually lowered initiating inflows and strong currents across the northern rim of the basin (Pardee, 1942).

Scour through the four passes (Figs. 10 and 20) mainly eroded closely jointed argillites of the Prichard Formation and Ravalli Group which form the bedrock of the northern rim. In addition to sweeping clean the blocky-weathering talus, eddying currents plucked bedrock from the passes producing several undrained depressions. The largest measures 305 meters across and is in the Wills Creek Pass (Fig. 2) but each of
FIG. 1 — Central part of Glacial Lake Missoula and the narrows that impeded its drainage (after Pardee 1942).
FIG. 2 — Topographic depressions formed by kolking action of flood waters through Wills Creek Pass (T21N R23W).
the other passes have several, yet smaller, scoured depressions. By studying the elevation of scoured features within the passes, Pardee estimated actively eroding currents to have been 9 to 12 meters deep in Duck Pond and Big Creek Passes, 46 meters deep in Markle Pass, and 76 meters deep in Wills Creek Pass. Since the channels are not filled by sediment from dwindling currents, Pardee believed the flows must have ceased abruptly.

The debris from both the fine shingle talus and scoured depressions was deposited in the lee of each pass to form what Pardee termed gravel bars. The largest of the bars is below Wills Creek Pass (Fig. 3) and extends about 1.6 kilometers downstream on the slope that descends to Camas basin (Pardee, 1942). The height of the bar at its downstream terminus is about 61 meters above the basin floor. A smaller bar at Markle Pass (Fig. 4), extends 1.0 km into the basin, and is 46 meters in height. He identified, but did not measure the bars at Duck Pond and Big Creek Passes.

Some of the sediment that was carried beyond the gravel bars was deposited downstream in a series of giant asymmetric ripple marks (Pardee, 1942). The ripple marks have the form, structure and arrangement of parallel-crested current ripples, but are so large that Pardee (p. 1587) recognized the inappropriateness of the term, ripple. Nevertheless, he retained it for lack of a better one, as did Baker (1973) in his study of Lake Missoula flooding in eastern Washington. In this paper, the bedforms shall be called transverse bars for reasons of
FIG. 3 — Areal extent of Wills Creek deltaic bar. Undulations of 3200 ft. contour correspond to transverse bars on delta surface.
FIG. 4 — Markle Pass deltaic bar and associated topographic depressions (black) created by kolking.
genesis and morphology which will be developed later. All of the bars are grouped in belts downstream from each of the four passes of the northern rim. Pardee did not include the bars beneath Big Creek Pass with the others because he felt they were not fully developed.

Pardee measured some of the bars and found that their amplitude varied from less than a meter to greater than 15 meters, but averaged between 4 and 9 meters. Pardee said that the wavelength of the bars reached a maximum of 152 meters, but averaged 76 meters. He stated that the general dimensions of the bars decreased from the center of the basin, where they are highest, to the lower downstream end where most of the bars eventually fade out completely. He did not, however, study the exact relationship of this change.

Additionally, Pardee generalized that the rest of the bars should have characteristics similar to the one he described from a gravel pit in the east center portion of the valley. He said that their largely sub-rounded pebbles, 8 to 20 mm in diameter, form loosely compacted, yet distinctly stratified crossbeds whose dip is oriented parallel to the lee side of the ridge. He also reported that the deposits contained little to no sandy matrix.

Pardee believed that the average grain size of the bars decreased downvalley, but he never measured this. Pardee attributed the postulated downvalley decrease in both average grain size and the size of the bars to steadily decreasing water velocity, but was unable to verify downstream decrease in water velocity because he did not measure grain
size elsewhere in the valley. However, by applying the relationship between maximum pebble size and stream velocity (Hjulstrom, 1935) he did estimate a water velocity of 2.1 meters per second for the single bar he studied. By application of a similar relationship (Leighly, 1934; Hjulstrom, 1935), Pardee also estimated an average water velocity of 19.8 meters per second for the currents which flowed through the passes.

Pardee believed that the current which deposited the bars was at least twice as deep as the bar heights or about 30 meters. He believed that the width of the channel may have been equal to the full width of the valley floor.

As the deposits were being formed, the water surface fell, shutting off flow through the passes of northern Camas basin (Pardee, 1942). The timing was determined by the elevation of each pass. Thus, Duck Pond and Big Creek Passes went dry after the water surface fell to about 1067 meters, followed by Markle Pass at about 1036 meters, and finally Wills Creek Pass at approximately 1006 meters (Pardee, 1942). A wide gap at the southern end of Camas basin emerged when the lake fell an additional 91 meters, marking the complete separation of the water between the basin and Perma Narrows.

Whether or not Glacial Lake Missoula drained completely is unknown (Pardee, 1942). The Camas basin did refill, however, as evidenced by two beach strand lines cut into flood channels at Markle and Wills Creek Passes (Pardee, 1942). The higher of the two is at Markle pass at an elevation of 1079 meters.
Varved silts were deposited in many of the valleys of western Montana during a refill of Lake Missoula, but according to Pardee (p. 1580) none were deposited in the Camas valley. He argued that none of the entrances to the basin were low enough to divert the silt-laden turbid waters from their source in the Mission basin into the Camas basin.
CHAPTER II

PURPOSE

The purpose of this study is to test Pardee's hypotheses and to compare the flood deposits of Camas Prairie with other glaciofluvial-glaciolacustrine studies. The two principal studies used for comparison are Baker's (1973) study of the glaciofluvial flood deposits of eastern Washington, and Bradley, and others' (1972) study of the flood deposits of Kink River, Alaska.

Pardee's suggestion that grain size and bedform size decrease down-valley was tested by plotting measured sediment samples and bedforms against distance. Pardee's grain size analysis of the gravel pit in S\(\frac{1}{2}\)E\(\frac{3}{4}\) Section 12 T20N R24W as representative of all the gravel ridges was evaluated by studying additional localities. Pardee's belief that no glaciolacustrine sediments were deposited in Camas basin following the flood event was investigated by studying the stratigraphy within the basin. Since Pardee recognized the inappropriateness of the term ripple mark, new information concerning size, shape, internal composition and depositional origin was used in an attempt to classify the bars. Finally, stratigraphic information was sought to provide data for answering the ongoing debate (Bretz, 1923; Pardee, 1942; Bretz and others, 1956; Alt and Chambers, 1970; Baker, 1973, 1978; Curry, 1977; Waitt, 1980) about how many flood events Glacial Lake Missoula produced.
CHAPTER III
TERTIARY GEOLOGY

Contrary to Pardee's belief (1942, p. 1588), a fine grained Tertiary volcanic agglomerate directly underlies the flood gravels along the valley floor of Camas Prairie. Pardee (p. 1577) thought the nearest Tertiary deposit was a small patch of gravel containing smooth quartzite cobbles located on a ridge west of Camas Creek about 3.2 kilometers north of Perma, Montana.

A contracted local backhoe operator struck Tertiary volcanic sediments at the base of the bars in two gravel pits studied (Fig. 5). Samples of the volcanic sediments have an overall tan color, but red, gray, and dark colored grains, believed to be xenoliths, impart a mottled appearance (Fig. 6). Red grains are interpreted to be iron rich xenoliths. The fragments are of two average grain sizes. The larger range from two to fifteen millimeters and the smaller from one to five millimeters.

Composition and texture of the fragments vary greatly. Some of the dark colored ones look like pieces of the locally occurring Precambrian Prichard and Ravalli rocks but verification requires thin section study. Others, definitely volcanic on the basis of their texture, range from massive, fine grained and gray to light and porphyritic to white with cryptocrystalline dark bands.

All the xenoliths have subangular to subround edges and rest in a
FIG. 5 — Location of gravel pits and water wells used to study the subsurface stratigraphy of Camas basin. (See Appendix B for cross section.)
FIG. 6 — Example of the texture and composition of the Camas Basin Tertiary volcanic agglomerate.
light colored, glassy groundmass that is microvesicular and porous. The almost spherical shape of the vesicles indicate very little compaction or flow. These rocks are very similar to the latite tuff of Hog Heaven Tertiary volcanics about 32 kilometers to the north. Shenon and Taylor (1936) described eleven isolated patches of volcanics overlying Precambrian Belt rock. They reported that the volcanics are overlain near their edges by Pleistocene lake beds (Fig. 7) interpreted to be deposits of Glacial Lake Missoula.

Shenon and Taylor named the volcanic rock latite tuff, which they described as highly altered and mottled. Phenocrysts and xenoliths are common and the most abundant alteration product is a reddish brown iron oxide. They also report that the rock contains somewhat rounded fragments of Belt Supergroup argillies.
FIG. 7 — Stratigraphic relationship of Glacial Lake Missoula silts to Hog Heaven volcanics (from Shenon & Taylor, 1936). Quaternary alluvium rests directly on the glacial silt deposits.
CHAPTER IV
PLEISTOCENE GEOLOGY

General

Pleistocene sediments of the Camas Basin can be divided into three units. The differences between each indicate a major change in environment and regime of sedimentation. The oldest is a basal lacustrine unit suspected to, where preserved, locally rest on Tertiary volcanics. Flood gravels produced by the strongest flow of one or more Glacial Lake Missoula flood events comprise the middle unit. The uppermost and youngest unit is a thin layer of silty varves. This sediment was deposited from the last stand of Lake Missoula.

Early Lacustrine Deposits

An earlier, preflood stage of Glacial Lake Missoula deposited light colored silty varves in the Camas Prairie Basin because clasts of these varves, measuring no larger than 3 centimeters (c-axis), occur within cross strata of the gravel ridges. The varves were never found resting directly on the Tertiary volcanics because erosion by the entrained bedload gravels during flood was probably great enough to remove most of the varves.

The varves are formed by tan silty layers and dark brown clay rich layers each about 3 millimeters thick. Due to the small clast size preserved, no more than three sets of laminae were ever observed in a
clast. All varves studied were normal graded, fining upward sequences of silt and clay.

**Flood Deposits**

**General**

The two types of flood deposits in the Camas Prairie basin are large transverse bars and deltaic bars. Both deposits consist primarily of graven (>2mm) and have been described by Pardee (1942). Since his only purpose in describing the features, however, was to support a flood deposit origin for them, the descriptions of the deposits are rather limited in scope. This paper describes more completely the deposits by stressing data pertinent to sediment hydraulics, such as percent gravel, bedform size, shape and distribution.

**Deltaic Bars**

The four deltaic bars below the four passes along the northern rim of Camas Prairie vary in size according to the duration and velocity of flow through each (Pardee, 1942). Their composition is colluvium and bedrock blocks derived from the passes. The changes in flow characteristics that occurred through each pass explain how the deposits were formed.

**Mechanism of Deposition**

In terms of hydraulics, each pass constricted water flow and increased head, which in turn steepened the hydraulic gradient. A steep gradient produced higher flow velocities through the area of constriction. Under such conditions, turbulent velocity produces secondary circulation, flow
separation and birth and decay of vorticity around obstacles and along irregular boundaries (Baker, 1978). An erosive form of such phenomena, which collectively are called marcoturbulence, is the kilk, (Matthes, 1946). Kilking very simply refers to erosion by hydraulic lift rather than abrasion. Bretz (1924, 1969), Bretz and others (1956) and Baker (1973, 1978) have very convincingly argued that the scablands of eastern Washington were eroded by plucking rather than by abrasion. They cite columnar jointing in the basalts of the scablands as facilitating the kilking mechanism. Although there is no columnar jointed, basalts around the rim of Camas basin, closely jointed argillites of the Prichard and Ravalli Group apparently served the same purpose.

Rocks, colluvium and other debris thus removed from the passes were transported downstream from the constrictions. There flow expanded in the y plane in response to increased channel width, and in the z plane because the valley deepened, producing flow separation (Fig. 8).

Flow expansion produced just the opposite general effect that flow constriction did. The hydraulic gradient was lowered, decreasing flow velocity so that the largest grains were deposited. Thus, most of these grains were deposited just past the lip of each pass in response to flow separation and expansion, forming the deltaic bars.

Transverse Bars

General

Part of the sediment not deposited as deltaic bars was transported downstream to form a system of transverse bars (or gravel ridges)
FIG. 8 — Diagram of flow expansion in the y and z planes through each pass.
(Pardee, 1942). These extend nearly across the valley floor, and in plan view they are arranged in straight transverse to transverse sinuous trains, except below each deltaic deposit where they form a system of nested arcs convex downstream (Fig. 20).

In a flood system with a tilted gradient such as described by Pardee (1942), there should be a progressive change in the hydraulic conditions reflecting a decrease in stream power between constrictions (passes). Figure 9, showing conditions parallel to flow, illustrates this idea. The conditions of stream flow digress from an energy high within the constriction where flow is turbulent and flow separation occurs, to a lower energy zone where flow reattachment and normal bedload transport is established, and end with the lowest energy level where slack water deposits (Patton, and others, 1979) may accumulate. In this picture, bedform size should decrease downstream as stream power does because decreasing water depth accompanies the decreasing energy conditions.

Method of Investigation

To test this hypothesis, a Dumpy level, Brunton compass and transit tape were used to measure the physical dimensions in the x and y directions (Allen, 1968, pp. 61-62) of the bedforms along an azimuth parallel to flow direction and perpendicular to axial trace (Fig. 10). (The profiles not figured in the body of this paper are included in the appendix.) The raw data were then tabulated and converted to vertical change with respect to a field datum corresponding to station one of each transect. Vertical and horizontal changes were then
FIG. 9 — Idealized interpretation of the progressive change in hydraulic conditions and related deposits of Camas basin.
FIG. 10 — Location of transects used in profiling the bedforms of Camas basin. (See Appendix A for profiles not used in text.)
plotted on graph paper to provide a visual display of the downvalley profiles of the bedforms. Graphed bedforms were compared to aerial photos to determine which bedforms were well defined and thus suitable for statistical treatment, resulting in 32 data points from the various profiles (Figs. 12-17).

The transects measured (Fig. 10) provide data from the two largest systems, Markle Pass and Wills Creek. Transects cover both the delta surface of each as well as the valley floor, from a point approximately equal to the zone of flow reattachment below the deltas (Fig. 9) to a distance downvalley of about 5 kilometers. Profile measurements reveal transverse bars on top of Wills Creek deltaic bar, but not on Markle Pass deltaic bar (Fig. 11). Therefore, no data from the surface of Markle Pass deltaic bar appears in the discussion below.

**Bedform Size Study Results**

The plot on arithmetic graph paper of bedform height (H) versus distance (Fig. 12) for all 32 data points shows a wide scatter of bar height from 0 to 3500 meters downstream, but a closer grouping thereafter with a negative slope. The best fit line through the data is a curve, thus the regression is not linear. Chorley (1966, pp. 340-348) states that almost all geomorphic relationships are non-linear. Even though an infinite number of non-linear forms exist, Chorley (p. 342) says that three functions (exponential, power and polynomial) commonly apply to geomorphic data. Straight line plots on semi-logarithmic paper express exponential functions, whereas straight line plots on logarithmic
FIG. II — Measured topographic profile of Wills Creek deltaic bar (a) and the top of Markle Pass deltaic bar (b).
FIG. 12 — Relation of 32 measurements of bedform height (H) to distance downvalley.
paper indicate power functions. The 32 data points for height-distance were plotted on each type paper, but the plots describe curves (Figs. 13, 14). This means that bar height versus distance is a polynomial function. Unfortunately, the scope of this study does not permit the solution of this more complicated regression.

I do believe, however, that I can explain the narrowing of the range of heights with distance. I think that near the passes, variations in channel roughness, geometry and current strength produced corresponding variations in bar sizes, thus the wide scatter of points on the left of each diagram. Farther downstream and away from these variations, a narrower range of bar sizes was produced in response to a more uniform, steady flow that prevailed.

Bedform chord (B) was similarly plotted as a function of distance and yielded much the same results (Fig. 15). It was thus concluded that additional plots on semi-logarithmic and logarithmic paper would produce no new results.

Since the exact rate of downvalley decrease of height and chord cannot be determined, an estimate can be prepared by averaging the values within an arbitrary interval of distance. This lessens the effect of the extreme values at the upstream end, but should not significantly alter the more uniform set of values farther downstream. To get such an estimate, I chose an arbitrary interval of 1000 meters and averaged height and chord measurements falling within each interval.

I plotted each average beginning from the delta base (800 m +), so as not to mix delta top data with downstream valley floor data. The reason
FIG. 13 — Relation to distance downvalley of bedform height (H) when plotted on semi-log paper.
FIG. 14 — Relation to distance downvalley of bedform height (H) when plotted on logarithmic paper.
FIG. 15 — Bedform chord (B) as a function of distance downvalley.
for separating the data was to allow for any differences in hydraulic conditions that may have existed between delta surface and the point of flow reattachment on the valley floor.

The results show straight line plots for bedform chord and height as a function of distance (Fig. 16). Each graph has a negative slope, demonstrating decreasing height and chord with downstream distance. Bedform chord decreases at a rate of about 15 m/km, while height decreases at about 0.7 m/km.

Baker (1973, pp. 50-53) compared the variability of 40 current ripple heights and chords from the channeled scablands to a relationship determined by Allen (1968, p. 71) for large scale asymmetrical ripples with a similar range in size. Baker found that the scabland data plotted within two standard deviations of Allen's plot, but that the scabland data had a much steeper regression line.

The 32 data points for the Camas Prairie were compared to Baker and Allen's plots and the results are shown in Figure 17. The regression line for the Camas Prairie is described by the equation

\[
\bar{H} = 0.144 \bar{B}^{0.72},
\]

where \(\bar{H}\) is the average height for a given bedform and \(\bar{B}\) is the average chord for the same sample. The regression line for the Camas Prairie compares very closely in slope to Allen's relationship, but has a higher y intercept (height value) than does Allen's plot. Although not shown in Figure 17, the data from this study fall within two standard deviations of Allen's data, as does Baker's.
FIG. 16 — Average bedform chord (a) and average bedform height (b) as a function of downvalley distance.
FIG. 17 — Variability of 32 large transverse bars of Camas Prairie Basin compared with data from Allen (1968, pg. 71) and Baker (1973, pg. 50-53). Dashed lines embrace one standard deviation of the data points for this study.
Baker attributes the steep slope of his plot to the coarse grain size of the scabland bedforms. He lists three possible reasons as given by Tanner (1967, p. 94) for the high VFI (F/H) values of the scablands:

1) Trough infilling may occur if significant amounts of suspended sediment settles out of the water.
2) Crestal material may infill troughs as falling water planes off bedform crests.
3) Bedform crests may be scoured if a change in flow regime occurs.

Grain size differences between the bedforms of the Camas Prairie and channeled scablands cannot be compared because Baker measured only intermediate particle diameters. This means that Baker's reason of coarse grain size as the cause for the steep slope of the scabland data cannot be compared with the Camas Prairie data.

The bedforms of the Prairie exhibit anomalously high VFI values as do the scabland data. The Camas values range from a low of 14 to a high of 36 for the 32 data points of Figure 1 and from 7 to 44 for all the bedforms measured.

**Sedimentology**

The internal structure of the bars is characterized by planar tabular cross bedding in angular contact with the underlying Tertiary volcanics. The average degree of dip of the cross stratification ranges from about 20° in the northern end of the valley, to about 19° near the southern end. Both average values are classified as moderately high angle.
Sorting is poor, and the fabric is an open work, grain-to-grain contact series of alternating coarser and finer cross strata. The thickness of the individual cross strata were measured at two locations. Near the middle of the downstream progression of bedform trains, the coarser cross strata ranged from 5 to 15 cm, and the fine units from 3 to 5 cm. Near the southern, downstream end, the coarse units measured between 2 and 3 cm and the fine units were 1 mm and less. Thus the thickness of the cross strata units decreases downvalley as does bedform size. Thinner cross strata units probably relate directly to the downvalley decrease in the percent gravel contained in the bars discussed in the next section.

Near the southern end of the valley, the cross strata were occasionally intercalated with irregular muddy or silty layers, apparently deposited contemporaneously by slack water (Patton and others, 1979) (Fig. 9).

The bars are composed chiefly of gray argillite and quarzite clasts of the Prichard Formation and Ravalli Group that crops out at each of the four passes. Clasts of a green, medium grained diorite which crops out on the east side of Wills Creek Pass infrequently occur within the bedforms. More of these easily recognizable clasts occur down the eastern side of the valley, indicating strong unidirectional flow and little cross valley mixing of clasts. Clasts of the underlying Tertiary volcanics also occur within the flood deposits. They were probably eroded by the traction load of boulders and cobbles as it bounded and gouged its way downvalley. The angularity of the clasts indicate very
short transport distances. Finally, clasts of varved lacustrine sediments were also found within the bars, indicating not only that a measurable amount of lacustrine sediment settled on the lake bottom prior to flooding, but that these clasts too must have been eroded as were the volcanics.

Rare grey, black and brown carbonate clasts and green and red argillite clasts were also found. These are thought to have been transported by ice to the Camas Valley. Erratics, reaching several meters in diameter, of the same exotic, non-local lithologies are found throughout the area.

Long, intermediate and short axes (L, I, S) of approximately 200 clasts were measured for plotting on Folk's sphericity-form diagram to characterize the particle shape of the clasts. Even though the data points were distinguished by lithology (Fig. 18) and L-axis increments (Fig. 19), there was no dominance of any one particle shape. There is, however, an obvious dominance of Belt Supergroup type lithology.

Immature sediments characteristically shown no dominance of particle shape; only a heterogeneity of shapes. Since the postulated sediment source for the bedforms is a mixture of talus, soil and jointed rock, the classification of immature is consistent with these observations.

Variations in Percent Gravel

General. The coarse grain size fraction of any current deposited bedform is the most important in terms of estimating the stream energy required to transport and sort (Hjulstrom, 1939; Visher, 1961). As a
FIG. 18 — Clast data for Camas basin transverse bars plotted on particle shape diagram and distinguished by lithology.
EXPLANATION:

CLAST SIZE (L-AXIS)
- ○ 10 - 39 cm.
- ■ 40 - 69 cm.
- x 70 cm.

FIG. 19 — Clast data for Camas basin transverse bars plotted on particle shape diagram and distinguished by L-axis size. (Data base is the same as that used in Fig.18)
result, some studies of fluvial deposits have specifically dealt with the gravel portion of the range of sediment sizes (Bradley, and others, 1972).

**Method of investigation.** Samples for a gravel grain size study were collected from road cut exposures of Montana Highway 382 which runs north-south through the valley nearly perpendicular to the axial trace of the bedforms (Fig. 20). Each sample weighed between 5 and 10 kilograms and was collected by shovel. The samples were collected from the lee side of each bedform about one meter below ground level. The samples were sieved to the sand-gravel size boundary using the facilities of Sorenson and Company, Engineers.

**Results.** The percent gravel was determined for 34 samples, and the results are plotted against downvalley distance in Figure 21. It shows that the average percent gravel decreases downstream, but the range of percent gravel values increases. The narrow range of percent gravel values of the left of the plot reflects the deposition near the passes of coarse, neavier grains that the water was incapable of transporting. The smaller sizes, however, were no doubt easier to transport and thus are the increasingly dominate size fraction downstream. The wide range of values downstream is explained by the poor sorting of the deposits. This explanation is supported by Figure 22 which shows that throughout the sampled section, the range of largest clast per sample remains wide.

**Antidune Deposits**

An antidune deposit is exposed in a gravel pit on the north side of
FIG. 20 — Crestal trace map of the transverse bars of Camas basin. Bedforms were sampled where they were cut by Highway 382.
FIG. 21 — Variation in percent gravel (>2mm) of transverse bars in relation to downvalley transport.
FIG. 22 — Coarsest clast size per sample versus downvalley distance. Data points are the same as those in Fig. 21.
Markle Pass. The deposit is arranged in a three part stratigraphic sequence consisting of a basal unit which is horizontally bedded, a middle unit which contains low angle cross strata dipping upstream, and an upper structureless unit. The upper, massive unit truncates the underlying antidune cross bedded layer (Fig. 23).

The type of generalized sequence described above has been reported from both laboratory studies (Simons and others, 1965, pp. 37, 40, 41, 44) and field studies (Harms and Fahnestock, 1965, pp. 84, 92, 97, plate 1) to represent the deposit that results when flow passes from transition into upper regime. As stream power increases, antidunes form but may be partially destroyed when standing waves break throwing much of the bed material into suspension. At that point, transport virtually halts and some of the sediment is deposited on the truncated surface (Simons, and others, 1965, pp. 37, 44, 52, Fig. 21).

Big Creek Pass is the only other pass where possible upper flow regime deposits were found. There a road cut about five by one meters across exposes a structureless deposit of matrix supported grains comparable in size to the deposit at Markle Pass.

**Subsequent Lacustrine Deposits**

Glacial lake varves deposited from one or more subsequent stands of water lie on top of the transverse bars in the basin. The thickness of these sediments varies from about two meters near the southern, lowest part of the valley to only a few centimeters in the northern, higher end.
Fig. 23. Photo of the three part stratigraphic sequence believed to be an upper flow regime antidune deposit. Location is north side of Markle Pass.
In every northern location studied, the varves occupy only the troughs of the transverse bars. Near the southern end of the valley, however, the varves completely cover the flood deposits.

Supporting evidence for post-flood lake stands are incised shorelines along the faces of Big Creek and Wills Creek deltaic bars. The highest identifiable along Big Creek deltaic bar stands at an elevation of about 1024 meters.
CHAPTER V
DISCUSSION

The flood deposits of the channeled scablands of eastern Washington have been extensively studied (Bretz, 1923, 1928, 1929, 1930, 1969; Flint, 1938; Baker, 1973, 1978; Waitt, 1980). From these studies, workers have suggested at least several episodes of flooding (Bretz, and others, 1956), and perhaps 40 or more (Waitt, 1980). Flood water drainage through valleys above the dam has received comparatively little study (Pardee, 1942). As a result, deposits of the two regions have not been correlated.

What this study has proven is that there was a minimum of two lake stands, separated by one major flood. There could have been several, earlier floods that preceded the one recorded in Camas basin, but all must have been lesser because there is no evidence of reactivation surfaces within the bedforms. Likewise, there could have been numerous smaller floods post dating this one.

In this study, two findings about the stratigraphy of the Camas basin are contrary to what Pardee (1942) postulated, and thus important to subsequent workers who may address the correlation problem. First, up to two meters of glacial varves overlie the flood deposits, thus the water level of the refill probably was high enough to connect the Camas basin with the sediment source areas of the Little Bitterroot and
Mission basins to the east. Secondly, the transverse bars do not rest on Precambrian Prichard and Ravalli Group bedrock. Instead, they overlie fine grained, tan volcanics that are probably Tertiary in age and similar to the nearby Hog Heaven volcanics described by Shenon and Taylor (1936).

This study also demonstrates a clear difference between the bedforms of the Camas basin and those that Baker studied in eastern Washington. The regression equation for height and chord variability of the Camas basin bedforms has a higher Y-intercept value and more gentle slope than does Baker's plot. The higher intercept simply indicates that the Camas bedforms tend to be higher at small chord values than does the system Baker studied. The much steeper slope of Baker's plot indicates that with increasing chord values, the scabland bedforms increase in height at a much faster rate than those of the Camas basin. Baker attributed the steep slope of his plot to the coarse grain size of the scabland bedforms. Since most of his values for intermediate diameter of largest grains range up to 6 meters (whereas the largest grain found in Camas basin measured less than one meter), his conclusion is consistent. This means that the regression equation for the Camas basin is more comparable to the sand sized grain data of Allen (1968).

The high VFI values of similar range to the scabland data are due primarily to trough infilling by later glacial lake varves. This kind of trough infilling is obviously very different from any of Baker's (1973, p. 52) three reasons suggested as the cause of high VFI values of the scablands.
The bars are lower flow regime ($F < 1$) features because when Pardee's (1942, p. 1597) current velocity estimate, 2.1 m/sec., and the minimum flow depth, 3 meters (bar height), are placed in the Froude number equation,

$$ F = \frac{V}{\sqrt{g \cdot h}} $$

where

$V =$ flow velocity

$g =$ acceleration due to gravity (9.8 m/sec$^2$)

$h =$ depth of flow,

the results are

$$ F = \frac{2.1 \text{ m/sec}}{\sqrt{9.8 \text{ m/sec}^2 \cdot 3 \text{ m}}} = 0.39. $$

All sedimentological data seem to indicate rapidly decreasing flow strength of the flood waters through the Camas basin. Antidune deposits and evidence of kolking in the passes are succeeded by lower regime bar deposits. The height and chord of the deposits decreases downstream at a rate of 0.7 m/km and 15 m/km, respectively. The percent gravel contained in the bars decreases from nearly 100 percent near the source to as low as 42 percent only 5 kilometers downstream. Whether this reduction in stream power is due to dwindling amounts of water, ponding by constriction downstream, or increasing flow cross section is unknown.
The problem of nomenclature for the Camas flood deposits was first addressed by Pardee (1942, p. 1587). The general problem of universally applicable bedform classification schemes has been present for years (Middleton, 1964, p. 2). Sand bedforms are most common and are thus the most widely studied. Examples of unusually large bedforms deposited from high energy flow conditions are rare. It is perhaps for these two reasons that the bedforms of this study, Baker's (1973), and many others do not neatly fit into any classification scheme based solely on size.

The classification system developed by Jackson (1975) and supported by Miall (1977) is not based on size alone, but on genesis as well (Table 1). Under this system, the bedforms of the Camas basin are mesoform features. They are formed during a high energy event, the breadth of the bars approach the width of the valley, the wavelength and height is measured in meters, and Curry, and others (1977) believe the features to have been formed in relatively shallow waters.

The bedforms named as examples in the table are based on previous classifications and do not describe the bedforms of the Camas basin. In comparison, however, the bedforms of this study are more like transverse bars for two reasons. A transverse bar is by definition oriented at right angles to the banks of the river channel, and grows by downcurrent additions to the slip face margins. The only other bedform which is similar due to its physical structure is the sand wave. However, Allen (1980) has demonstrated that sand waves are generated under
Table 1. A Genetic Classification of Bedform Hierarchy
(from Miall, 1977)

<table>
<thead>
<tr>
<th>Microform</th>
<th>Mesoform</th>
<th>Macroform</th>
</tr>
</thead>
<tbody>
<tr>
<td>Example</td>
<td></td>
<td>Point bar, lateral bar, meander belt</td>
</tr>
<tr>
<td>height</td>
<td>( \approx d_b )</td>
<td>( \leq d, &gt; d_b )</td>
</tr>
<tr>
<td>length</td>
<td>ripple wavelength (centimetre scale)</td>
<td>dune or bar wavelength (metre scale)</td>
</tr>
<tr>
<td>breadth</td>
<td>( \ll Z )</td>
<td>( &lt; Z )</td>
</tr>
<tr>
<td>duration (assuming maintenance of equilibrium)</td>
<td>( \ll T ) (seconds to hours)</td>
<td>( \approx T ) (months to few years)</td>
</tr>
<tr>
<td>genetic control</td>
<td>local boundary turbulence</td>
<td>turbulence during high energy dynamic event</td>
</tr>
</tbody>
</table>

Symbols: \( d \): depth of flow; \( d_b \): thickness of turbulent boundary layer; \( Z \): width of flow field (channel width); \( T \): periodicity of dynamic event (storm, seasonal flood etc.).
fundamentally different dynamics and kinematics. He argues that the sand wave is restricted to tidal environments and is distinguished from large transverse bedforms of fluvial environments by the differences in which the bed material is transported. The former has unique mass transport currents attributable to the wave motion itself, and the bed curvature in the presence of the oscillatory flow.

T, the periodicity of the dynamic event which created the deposits is unknown (Table 1). The dynamic event could have been as frequent as yearly (Marcus, 1960; Bradley, and others, 1972), or of a much longer period. But whatever the timing, it is no doubt ultimately controlled by variations in climate.
CHAPTER VI
CONCLUSIONS

The stratigraphy of the Camas Prairie basin consists of a basal Tertiary volcanic fill of unknown maximum thickness but which probably rests on Precambrian Prichard and Ravalli Group argillite bedrock. Pleistocene flood gravel deposits and perhaps locally a pre-flood section of glacial lake varves unconformably overlie the Tertiary deposits. The flood gravels are in turn overlain by a subsequent unit of varved sediments which varies from a maximum of two meters in the south, to a feather edge in the north.

The floor deposits consist of deltaic bars appended to each of the valley's four passes, and large scale transverse asymmetric bars. The bars decrease downvalley in height and chord at rates of 0.7 m/km and 15.0 m/km, respectively. Their vertical form index (VFI) ranges from 7 to 44. Crossbedding is moderately high angle planar tabular at dips of between 19 and 26 degrees. Sorting is poor and the sediments are immature.

The variability of chord to height of the bars takes the form of \( \bar{H} = 0.144 \bar{B}^{0.72} \) and is comparable to Allen's (1968) data on sand-sized bedforms. The percent gravel of the bars decreases from nearly 100 percent near the source, to as low as 42 percent 5 kilometers downstream. Intermediate diameter of largest particle does not follow any relationship downstream and thus reflects poor sorting.
Antidunes and evidence of kolking in each of the passes, succeeded by lower flow regime bedforms of decreasing size, and decreasing percent gravel, all indicate steadily decreasing stream power. Whether the decrease is due to a dwindling supply of water, ponding by constriction downstream, or increasing flow cross section is uncertain.
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APPENDIX A

Topographic profiles not used in text of paper. See Fig. 10 for location of profiles. Downstream is to the right in all cases.
Profile 2
Detail of a single bar along Profile 4. Solid line represents actual surface, dashed line is an estimate of the original depositional surface.
Detail of a single bar along Profile 7.
APPENDIX B

Subsurface geologic cross section as interpreted from water well log sample descriptions.