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Chapter 5

Large-Scale Erosional and Depositional Features of the Channeled Scabland

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ABSTRACT

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The Channeled Scabland is a great anastomosing complex of highly overfit stream channels eroded into the basalt bedrock and overlying sediments of the Columbia Plateau. Both the erosional and depositional bed forms in these channels can be described according to a simple hierarchical classification. The catastrophic flood flows produced macroforms (scale controlled by channel width) through the erosion of rock and sediment and by deposition (bars). Mesoforms (scale controlled by channel depth) are also erosional and depositional. Microforms (scale controlled by the inner part of the turbulent boundary layer) are not discussed.

Large-scale bedrock macroforms are crudely developed, especially in the Cheney-Palouse scabland tract. The streamlined residual forms in loess are the most striking macroforms. Their characteristic elongation of 3 times their maximum width results from the balancing of skin resistance and pressure drag factors to create an equilibrium landform. The scabland bars are the characteristic depositional macroforms. Pendant bars, the most common type, accumulated wherever large flow separations were generated by various flow obstructions or diversions.

The bedrock mesoforms include longitudinal grooves, potholes, rock basins, inner channels, and cataracts. A sequence of bed form development probably operated for bedrock erosion in the Channeled Scabland. With time the forms changed from (1) longitudinal grooves, to (2) rock basins and potholes, (3) inner channels with recessional cataracts at their heads, and finally to deeply incised inner channels.

The depositional mesoforms for major channel flows were giant current ripples varying from 18 to 130m in chord length and from 0.5 to 7m in height and composed predominantly of gravel. The bed forms may be empirically related to depth-slope, mean velocity, and stream power. Correlation coefficients for the relationships are all greater than 0.9. Nevertheless, the prediction of these hydraulic parameters from ripple dimensions applies only to the narrow range of flow conditions which characterized the Missoula Flood through the reaches containing the bed forms.

REGIONAL CHANNEL PATTERNS

J. H. Mackin has been quoted as saying, "to understand the scabland, one must throw away textbook treatments of river work" (Bretz and others, 1956, p. 960). Certainly a failing of Bretz' critics in the Spokane Flood debate was their insistence that the Channeled Scabland conform to "established" geomorphic processes. The scale of the problem was key, and a completely new frame of reference was required. Bretz (1932a, p. 28) provided the required viewpoint: "Channeled Scabland is river bottom topography magnified to the proportion of river-valley topography."

The Spokane Flood debate might have been resolved more easily if the participants could have viewed modern orbital photography of the region (Fig. 5.1). At a glance one can appreciate Bretz'

observation. Clearly the Channeled Scabland is a plexus of *channels* rather than a network of valleys.

Channel Anastomosis

The general pattern of the Channeled Scabland is large channels eroded in loess and underlying basalt. The channels form locally anastomosing complexes with individual channels that have relatively low sinuosity. The term "anastomosis" should not be confused with "braiding" (as was done in the A. G. I. *Glossary of Geology*). "Braiding" refers to branching and rejoining around alluvial islands or bars. Braided streams are part of a continuous series of fluvial forms that develop in quasi-equilibrium with external controls on the river systems. "Anastomosis" has no genetic connotation. It refers to channel morphology whether in alluvial streams ("braided")



Figure 5.1. Anastomosing channel pattern in the Telford-Crab Creek scabland complex. This LANDSAT image depicts a scene 70 x 150 km (LANDSAT E-1039-18143-5, 31 August 1972).

or bedrock streams. The scabland anastomosis is deeply cut into rock. Anastomosis occurs in the Channeled Scabland because pre-flood valleys did not have the capacity to convey the Missoula Flood discharges without spilling over pre-flood divides into adjacent valleys. This crossing of divides produces the effect of channels dividing and rejoining. Before the era of aerial photographs and adequate topographic maps, Bretz (1928b) used field surveys to show over 100 channel ways in the scablands, 50 of which bifurcated in a downstream direction. Modern data sources now show that anastomosis occurs on a variety of scales. The regional scale (Fig. 5.2) is controlled by pre-flood topography. Individual channels may also be controlled by relatively straight geologic structures, such as High Hill anticline in the Channeled Scabland. Small-scale anastomosis includes the minor divide crossings used for reconstruction of the flood high-water surface.

STREAM OVERFITNESS

Misfit streams are streams that are either too small or too large for the valleys in which they flow (Dury, 1958, 1964). The underfit variety is relatively common, and such streams often show smaller channel widths and meander wavelengths than the winding valleys in which they flow. The disparity between river and valley size is explained by a reduction in stream discharge, either by capture (Davis, 1913) or by climatic change (Dury, 1965). Although there is considerable debate over the validity of the climatic explanation of underfit streams, the concept does seem to apply to alluvial valleys (Baker and Pentead-Orellana, 1977).

An overfit stream is too large for the valley in which it flows. Dury (1964) considered overfit streams in the context of sudden increases in discharge with rapid channel enlargement. Although such a condition would not persist long in an alluvial valley, the erosion of bedrock probably provides an opportunity to preserve overfit stream relationships.

The Crab Creek area, near the town of Wilson Creek in the Channeled Scabland, provides an excellent example of overfit relationships introduced

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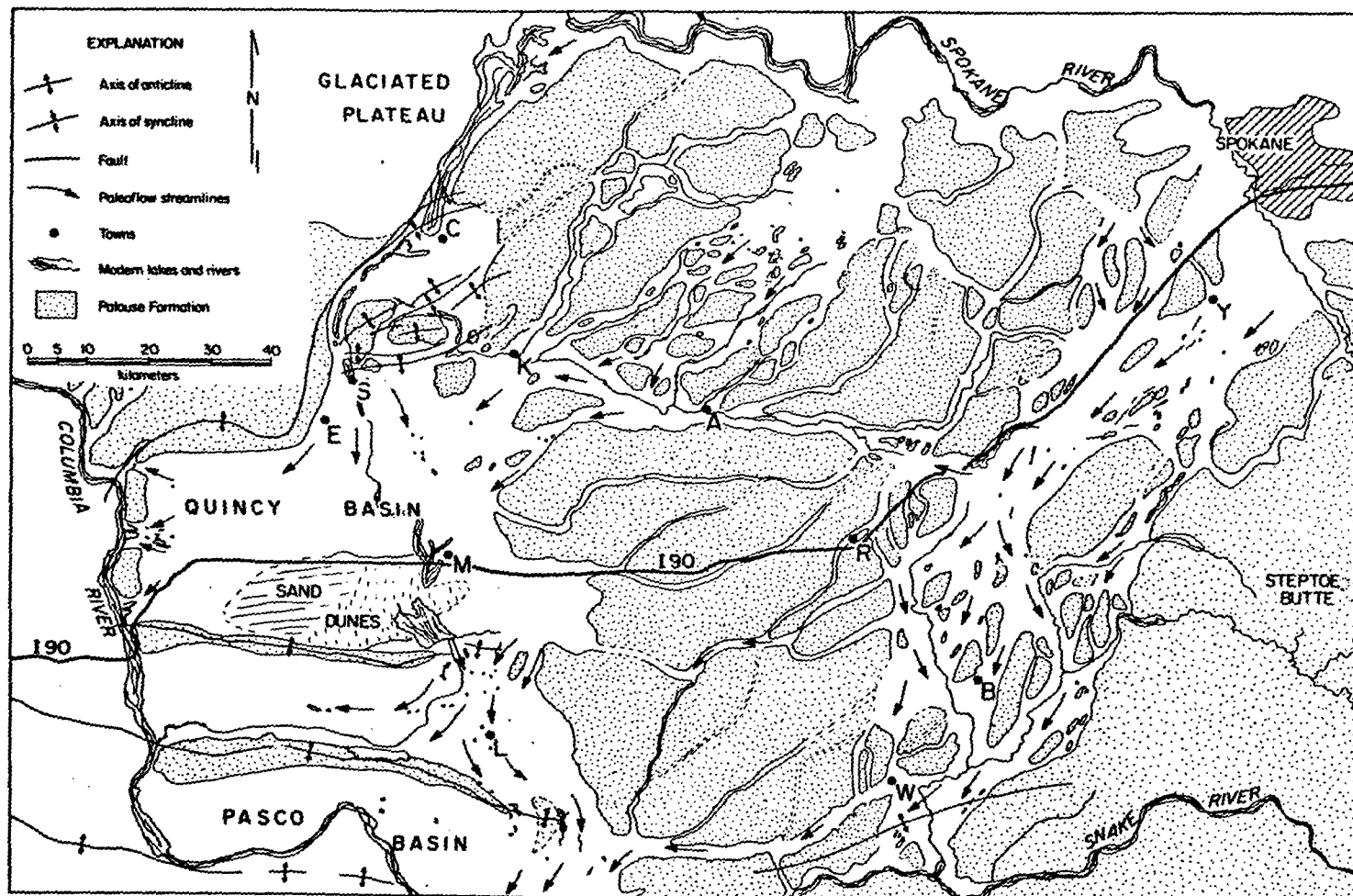


Figure 5.2. Regional pattern of the Channeled Scabland as mapped from LANDSAT imagery (E-1039-1843-5 and E-1004-18201-7). (Y) Cheny, (R) Ritzville, (B) Benge, (W) Wash Tucna, (M) Moses Lake, (E) Ephrata, (S) Soap Lake, (C) Coulee City.

by catastrophic scabland flooding. (See Bretz and others, 1956, their Plate 8; Baker, 1973a, his Fig. 22.) Prior to the catastrophic flooding, the topography near Wilson Creek was probably very similar to that of the modern Palouse Hills region near Pullman, Washington, with the interstream divides thickly mantled by the Palouse loess. The major streams had been superimposed onto the basalt from this loessal cover. The streams flowed in relatively narrow valleys. The well-formed valley meanders had a wavelength of 2000m. Using empirical relationships that characterize most rivers, the normal bankfull discharge of Crab Creek was approximately 850 m³/sec.

The last major scabland flood completely filled the valley of Crab Creek and adjacent streams. Water spilled over many of the divides between the stream valleys. The Missoula flood flows in this area where approximately 2,800,000 m³/sec (Baker, 1973a, his Plate 1). Bretz (1928b) noted that the flooding could not tolerate the leisurely pre-flood curves of the incised stream. Slip-off slopes were vigorously attacked producing what he called "trenched spur buttes". Huge streamlined bars were deposited downstream from the former valley bends. Many of these have giant current ripples on their upper surfaces (Fig. 5.3). The pre-flood topography was reduced to mere bottom roughness elements by flooding 3 to 4 orders of magnitude greater than any flooding these streams had ever experienced. In the upper parts of the Wilson Creek drainage, the scabland erosion presents a striking contrast to the adjacent loess-mantled terrain (Fig. 5.4).

A HIERARCHY OF SCABLAND BED FORMS

A fluvial bed form is defined (Am. Soc. Civil Engrs., 1966) as follows: "any deviation from a plane bed that is readily detectable by eye or higher than the largest sediment size present in the parent bed material". Most of the scientific interest in bed forms has focused on small primary forms in sand, e.g. Allen (1968). However, the Channeled Scabland affords a unique opportunity for the study of large-scale forms that are either composed of gravel (Bretz, 1928b; Baker, 1973a) or eroded into rock (Baker, 1973b). Two major scales of bed forms have remarkable pres-

ervation throughout the scablands. The larger features are scaled to channel width and consist either of depositional "bars" or various kinds of erosional residuals in rock or sediment. The smaller features are scaled to the phenomenal scabland flow depths.

Jackson (1975) has developed a hierarchical classification of bed forms generated by fluid shear and composed of cohesionless granular material. The classification relates to bed form size and to the time span of existence for various bed configurations. The bed form groups each relate to different formative processes (Jackson, 1977). Macroforms for rivers include point bars, scroll bars, alternate bars, and pool-and-riffle sequences. These bedforms do not relate to local flow conditions. They rather respond to long-term hydrologic and geomorphic factors. Mesoforms include large-scale ripples (dunes), antidunes, and large-scale lineation. The spacing of mesoforms depends on the outer zone of the turbulent boundary layer as the flow varies through a dynamic event such as a flood. In rivers the boundary layer control is approximated by flow depth. Microforms include current lineation and small-scale ripples. Microforms respond to flow structure in the inner part of the turbulent boundary layer, and their lifetime is much shorter than the periodicity of dynamic events.

Jackson's classification is especially interesting because of its genetic implications. Heretofore most hydrodynamic bed form studies have focused on the unstable motion of water over a rough boundary generating variable shear stress. Flow properties are perturbed either into longitudinal vortices that produce flow separation or into transverse roller vortices that produce alternating separation and reattachment of streamlines along the boundary (Allen, 1971a, 1971b). Until recently there has been little correspondence between experiments and theory. Jackson (1976, 1977) advocates a resolution to this dilemma through the concept of flow structures.

Recent fluid mechanics work (Laufer, 1975) indicates that turbulent shear flows contain an inherent structure consisting of discrete secondary flow patterns superimposed on the prevailing unidirectional mean flow. For the mesoscale, these structures consist of the bursting phenomenon (Offen and Kline, 1975), longitudinal vortices (Karcz, 1967), in-phase waves, and possible



Figure 5.3. Oblique aerial photograph of Bar 4 in the great scabland channel that was carved from the pre-flood valley of Crab Creek. The bar is developed for 3 km downstream of the trenched spur buttes (T) that mark the pre-flood slip-off slope of Crab Creek's meandering valley. Giant current ripples (R) mark the bar surface. These ripples average 3.3 m in height with chords of 66 m. The flood flow maximum depth here was 90 m.

transverse roller vortices. Jackson (1977) speculates that fluvial dunes (a variety of large-scale asymmetrical ripple) are produced by bursting; large-scale lineation (Coleman, 1969) produced by longitudinal roller vortices; and antidunes are produced by in-phase waves. Although this scheme requires further experimental verification, it nevertheless is a useful working hypothesis for resolving the enigma of bed forms.

Although much attention has focused on depositional forms (ripples, dunes, sand waves, etc.) the Channeled Scabland requires consideration of erosional forms as well. Allen (1971b) discusses the two theories generally applied to explain erosional bed forms. The passive bed theory holds that the fluid flow imparts its properties (bursting, vortices, etc.) on a passive bed through hydraulic character of the responsible flow. The defect theory, in contrast, emphasizes defects or irregularities on the bed which disturb the flow to generate turbulent flow separations. Actually the two theories are difficult to resolve

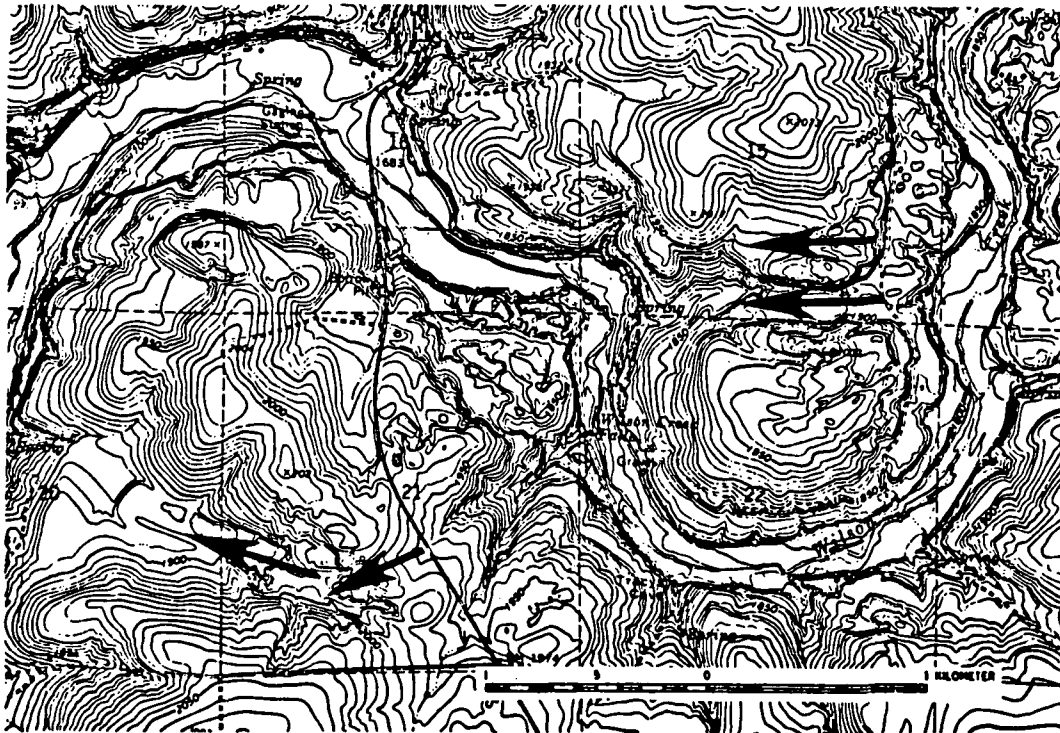


Figure 5.4. Topographic map of the upstream part of Wilson Creek (U.S.G.S. Almira, Wash., 7.5-minute quad-

angle). Contour interval is 3 m (10 feet). Arrows show major divide crossings.

Table 5.1. Important Bed Forms In The Channeled Scabland

| | Scoured in Rock | Scoured in Sediment | Depositional |
|--|---|---|---|
| MACROFORMS (Scale Controlled by Channel Width) | Pool-and-Riffle Sequence. Quadrilateral Residual Forms in Channel. Anastomosis. | Large-scale Streamlined Residual Forms. | Longitudinal Bars. (a) Pendant Bars. (b) Alternate Bars. (c) Expansion Bars. Eddy Bars. |
| MESOFORMS (Scale Controlled by Channel Depth) | Longitudinal Grooves. Potholes. Inner Channels. Cataracts. | Scour Marks. | Large-scale Transverse Ripples (Giant Current Ripples). |
| MICROFORMS | Scallop Pits. | Not Preserved. | Small-scale Ripple Stratification. (Restricted to Slackwater Facies.) |



Figure 5.5. Excellent association of depositional macroform (bar) with superimposed mesoforms (giant current ripples). The bar occurs along the Snake River just downstream from its junction with the Palouse River (top center). The Snake River is impounded by a dam just downstream from the reach depicted in this oblique aerial photograph.

because the effects of one tend to enhance the operation of the other. In the basalt bedrock of the Channeled Scabland it is mainly structural irregularity of the rock that provides defects which then perturb the flow hydrodynamics to create distinctive bed forms.

A modification of Jackson's (1975) classification will be used in subsequent discussions of bed forms in the Channeled Scabland (Table 5.1). A rather fortuitous aspect of the depositional morphology in the Channeled Scabland is the excellent preservation of both macroforms and their superimposed mesoforms (Fig. 5.5). This association probably results from the nature of the

hydrograph for flood flows in the Channeled Scabland. In most rivers the hydrograph shows a long recession. Depositional bed forms that are stable at high stage (mesoforms) are washed out, and postflood surfaces show only the highly stable macroforms such as alternate bars. In the Scablands an abrupt cessation of flood discharge resulted in preservation of many of the mesoforms, especially those located on the higher bar surfaces.

BEDROCK MACROFORMS

Quadrilateral Residual Forms

The eastern part of the Channeled Scabland is generally called the Cheney-Palouse Scabland tract (see Patton and Baker, Ch. 6, this volume). Its overall pattern is a complex of channel ways and interchannel divides. The pattern is especially pronounced on orbital photography (Figs. 5.1 and 5.2) because the loess on interchannel divides contrasts sharply with the eroded basalt scabland on the channel floors.

In overall pattern the Cheney-Palouse scabland tract resembles a braided stream. Instead of bars of sediment laid down by high discharges, however, the "islands" in the Cheney-Palouse are erosional residuals of basalt and loess. The largest of these have a crude quadrilateral shape in plan, often forming diamonds or parallelogram shapes. Morphologically these are similar in appearance

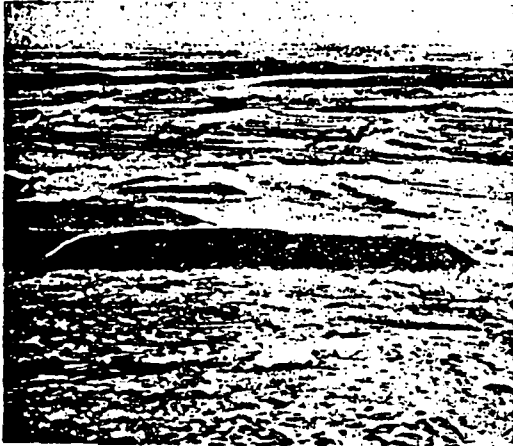


Figure 5.6. Oblique aerial photograph of the upstream end of a quadrilateral loess residual on the Palouse-Snake divide. Note the steep scarps that were eroded by flood water streaming over the divide. Washtucna Coulee is visible in the background.

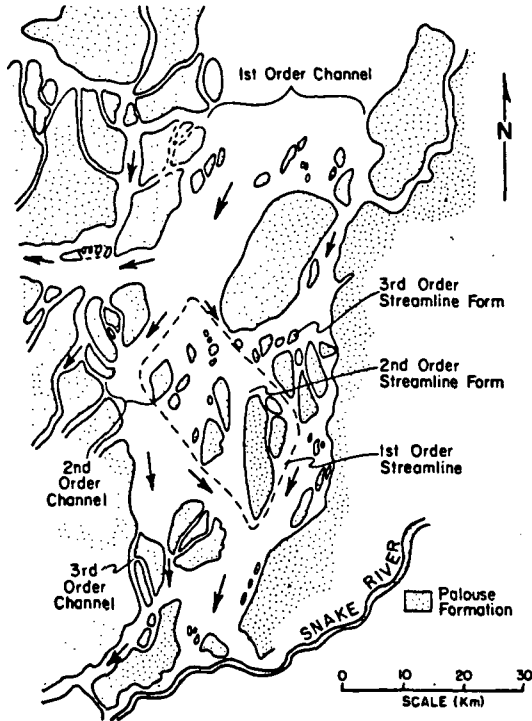


Figure 5.7. Sketch map of the Cheney-Palouse scabland showing the hierarchical relationship of channels and interchannel residual areas (mantled by Palouse Formation). An arbitrary section has been ordered to show these relationships.

to eroded bar remnants that are common in braided gravel streams. The flow split around these residual elements, eroding their upstream ends (Fig. 5.6). The flow then reconverged on the downstream ends of the residual forms. Many of the residuals were further modified by relatively shallow flood water flowing obliquely across their surfaces.

The successive division of channels and of the interchannel residual elements results in a kind of hierarchy for these erosional macroforms. Both the channels and the residuals can be ordered in the same way that Williams and Rust (1969) have ordered depositional bars in braided outwash (Fig. 5.7).

Pools and Riffles

Pools are topographically low areas produced by scour on river beds. They seem to have a very regular spacing in stream channels relative to topographically high areas called riffles. As pointed out by Leopold and others (1964, p. 203), pools and riffles tend to be spaced at 5 to 7 times the channel width. Keller and Melhorn (1973) have pointed out the significance of convergent and divergent flow for the development of pools and riffles. Pools tend to occur at locations where convergent flow increases the bottom shear stress at flood stages. In alluvial rivers, the riffles also develop at high stage when divergent flow induces deposition. In normal rivers the usual low flow condition produces a velocity reversal; the water moves more rapidly over the riffles, eroding them and transporting sediment into the relatively tranquil pools (Keller, 1971).

Because the Channeled Scabland contains some pronounced convergent and divergent flow patterns (Fig. 5.2), we would expect that flood scour might produce some incipient pool-and-riffle development. Detailed high-water profiles of the Cheney-Palouse scabland tract (Patton and Baker, Ch. 6, this volume) indicate the presence of a crude pool-and-riffle sequence. Although spacing is somewhat irregular, an average spacing of 10-15 km exists between successive riffles (steep areas on profiles). It is unclear whether this represents a pool-and-riffle sequence associated with a meandering thalweg or whether it is simply an effect caused by varying channel widths along the flood route. It is possible that both fac-

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tors are interacting. The erosion of pools into bedrock requires further study.

STREAMLINED RESIDUAL FORMS

Bretz (1923b, p. 624-626) first recognized that the hundreds of isolated loess hills of the eastern scablands possessed remarkably steep, unguilted marginal hillslopes. These slopes converge to form definite prows that point up the local scabland gradient (Fig. 5.8). Although Bretz and others (1956) interpreted these hills as fluvially-eroded loess "islands", high-water mark reconstruction (Baker, 1973a) has shown that many of the hills were eroded subfluvially. There is a complete transition in scale and complexity between the small and simple streamlined hills and the large, complex zones of less modified regions of loess topography (quadrilateral forms) that separate the major anastomosing channels.

The best developed streamlined forms (Fig. 5.9) often show several or all of the following characteristics: (a) flow obstacles that localized the resistant landform, (b) upstream crescent-like scour marks, (c) downstream tapering streamlines on the adjacent channel floor, (e) oblique channels cutting through small divides at the crest of the streamlined form. The striking contrast between preserved residuals of relatively soft loess and adjacent tracts of scoured, potholed rock appears at first to be a physical paradox. Why did the harder rock erode, while the loess was preserved?

Streamlined forms are common in a variety of geomorphic phenomena. Drumlins, yardangs, and "island hills" are the characteristic forms for moving glacial ice, wind and water respectively. The water forms generally do not occur at a large scale on earth because most terrestrial rivers are too shallow. Again the deep, swift scabland flood flows account for another remarkable type of landform.

The scabland streamlined loess hills were studied by morphometric analysis of large-scale topographic maps. To simplify the analysis we considered the shapes of the hills as projections onto a plane surface parallel to the geoid (or general ground surface). Thus we only considered the shape of the forms in plan. Analysis of

the streamlined shapes included three physical measurements: length, l (km) — measured parallel to suggested flow direction; width, w (km) — taken as the maximum width of the streamlined form measured perpendicular to the implied flow direction; and area, A_s (km²) — measured with a grid emplaced over the feature. From these physical measurements a dimensionless parameter (K) can be calculated:

$$K = \frac{l^2 \pi}{4 A_s} \quad (5-1)$$

Chorley (1959) used this parameter to show that drumlins have a close resemblance to airfoils. Through the use of the lemniscate loop equation:

$$P = 1 \cos K\theta, \quad (5-2)$$

where P and θ are polar coordinates, an equation for the streamlined form can be derived and plotted on polar coordinates (Fig. 5.10). The shapes strongly support Bretz' contention that the loess hills were streamlined by a rapidly flowing fluid. Water velocities averaged 12-15 m/sec for depths of 30-60 m in these areas.

The width (w) vs. length (l) ratio serves as a measure of the Index of Elongation (Fig. 5.11). Greater values of the ratio l/w (more gentle slopes) imply greater elongation of the form. The curve fitted to the point scatter (via regression analysis) shows that the streamlined forms exhibit nearly parallel relationships. There is little change in the l/w ratio with increasing size. The index is very close to 3.

The relationship of length to area (Fig. 5.12) measures the narrowness of streamlined forms. It is a measure of the relative percentage of the area contributed by the length component. Higher l/A_s values indicate increasing narrowness of streamlined forms.

The relationship of width to area (Fig. 5.13) measures the degree of broadness, i.e. how much of the total area is contributed by the width component. As the w/A_s increases, the tendency toward broader forms will increase.

These detailed shape analyses further document the average geometry for streamlining of the scabland loess hills. If the hills had rectangular shapes, then their total areas A_t would be

$$A_t = lw. \quad (5-3)$$

However, from Figures 5.12 and 5.13 we have:

$$l = 1.9 \sqrt{A_s}, \quad (5-4)$$

$$w = 0.66 \sqrt{A_s}, \quad (5-5)$$

and

$$A_s = \frac{3.19}{4 A_t}. \quad (5-6)$$

That this area ratio derives from producing a streamlined shape is seen in the analogy of producing the minimum perimeter shape that has the same length and width as a square. For the latter problem, $l = w$ is the diameter of a circle with area A_c :

$$A_t = l^2.$$

$$A_c = \frac{\pi}{4} l^2.$$

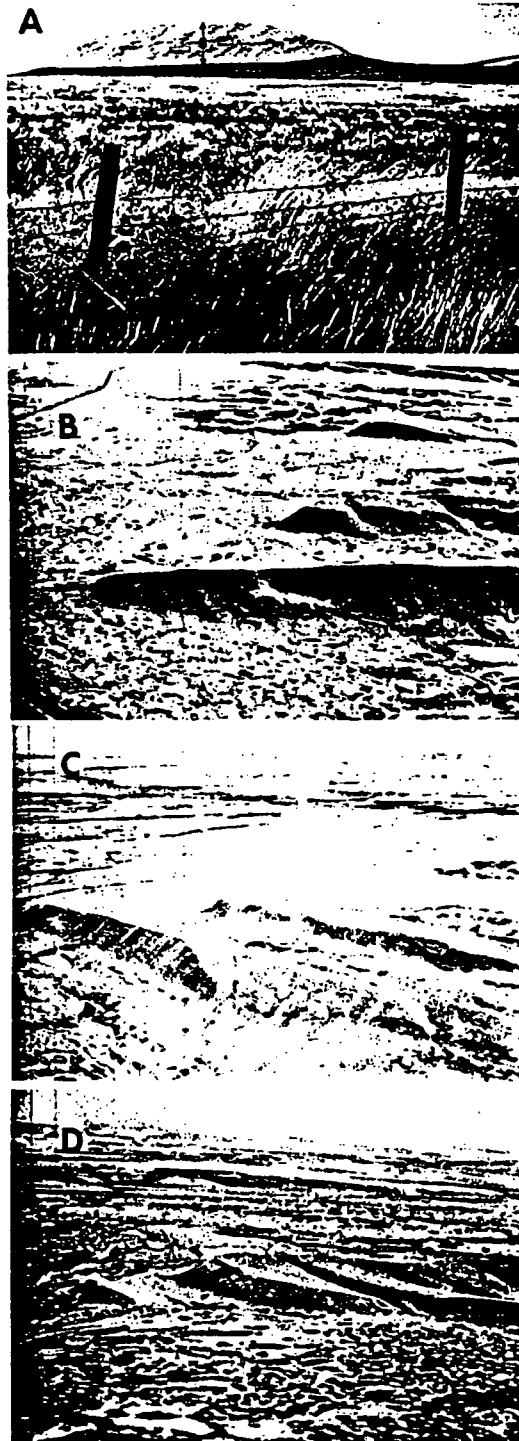
$$A_c = \frac{\pi}{4} A_t. \quad (5-7)$$

which is nearly identical to the result produced empirically in equation (5-6). We can further extend this theoretical exercise to calculate an ideal K factor:

$$K = \frac{l^2 \pi}{4 (\pi (l/2)^2)} = 1. \quad (5-8)$$

By minimizing K one is essentially minimizing the skin resistance of an object with a given area. However, skin resistance is only one of two components of the drag that develops around any submerged object. The other component is pressure drag, which derives from the turbulent wake developed behind an obstacle. This factor is

Figure 5.8. Streamlined residual forms in the Cheney-Palouse scabland: A. Ground-level view showing height(s) of the flood-eroded loess scarp. B. "Upstream" end of the loess hills on the Palouse-Snake crossing. Note the prominent prows formed by convergent hillslopes pointing upstream for the flood flows. The scarps average 40 m in height. Maximum flood flow depths averaged 60 m at this location. C. "Downstream" ends of same loess hills. D. View of the channel cutting obliquely through the "snout" of a stream-lined loess hill. Note the surrounding scabland and unmodified loess topography in the far distance.



minimized by increasing the elongation of the obstacle so that less width component is presented transverse to the flow. The scabland forms are elongated to about three times their maximum width (Fig. 5.11). It is apparent that the streamlined hills are equilibrium forms, elongated sufficiently to reduce pressure drag, but not so long that they create excessive skin resistance. They owe their preservation in the high-velocity flood flows to this marvelous streamlining. Whereas the adjacent rock outcrops generated destructive macroturbulence, the loess hills fostered a smooth bending of streamlines around them.

Some slight tendency for increased streamlining (increased K factors) appears to occur with in-

creasing Reynold's number (Fig. 5.14). The Reynold's numbers (R) used are the maximum values achieved in the flood flows and were computed from the expression:

$$R = \frac{QR\rho}{A\mu}, \quad (5-9)$$

where Q is the maximum discharge of a scabland reach (computed by slope-area procedures described by Baker, 1973a), R is the hydraulic radius of the reach, ρ is the fluid density, A is the cross-sectional area, and μ is the dynamic viscosity. These maximum flow Reynold's numbers varied from 2×10^8 to 2×10^9 in most scabland reaches.

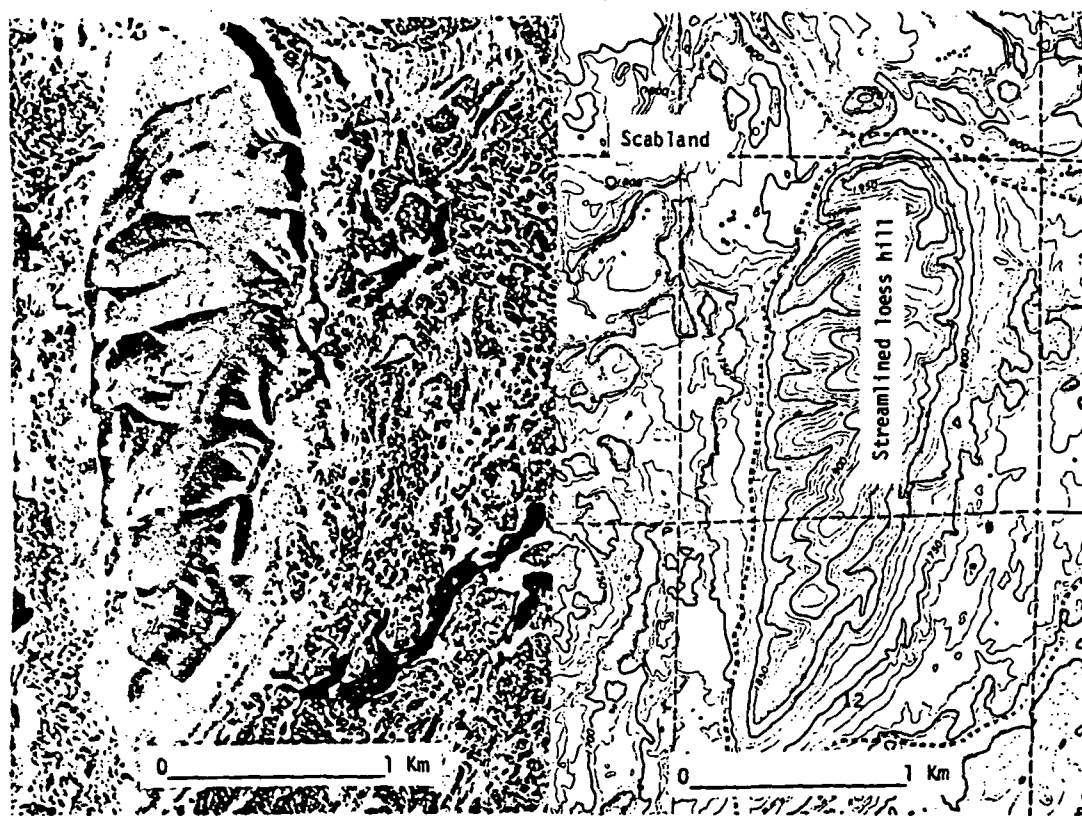


Figure 5.9. Residual loess hill streamlined by flood erosion. A small cataract, heading an inner channel, has worked its way around the blunt upstream end of the hill (on map). Longitudinal grooves and butte-and-basin topography can be seen in the marginal scablands. Water depths and velocities averaged 12 m/sec for depths of

30-40 m in this area during the flood maximum. Map contour interval is 10 feet (~ 3 m). This hill is located in the Cheney-Palouse scabland tract in sections 1 and 12, T. 18N., R. 37E. The same hill is viewed obliquely in Fig. 5.8D.

BARS OF THE CHANNELED SCABLAND

The term "bar" is used for all large-scale depositional forms in streams and rivers. Bars originally denoted impediments to navigation. Here the term is applied to all depositional macroforms in scabland channels. Unfortunately there is no generally accepted classification of fluvial bars. Indeed, a hard and fast classification of bars is probably impossible for the following reasons:

(1) no classification can satisfy all three major purposes of bar studies, morphologic, hydrodynamic, and paleohydraulic; and (2) many bar forms are ephemeral members of evolutionary sequences, complexly related to initial conditions and transitory flow conditions.

The classification used here for scabland bars is limited to the Channeled Scabland. It is developed from the relationship of bars to the large-scale flow pattern in a local scabland reach.

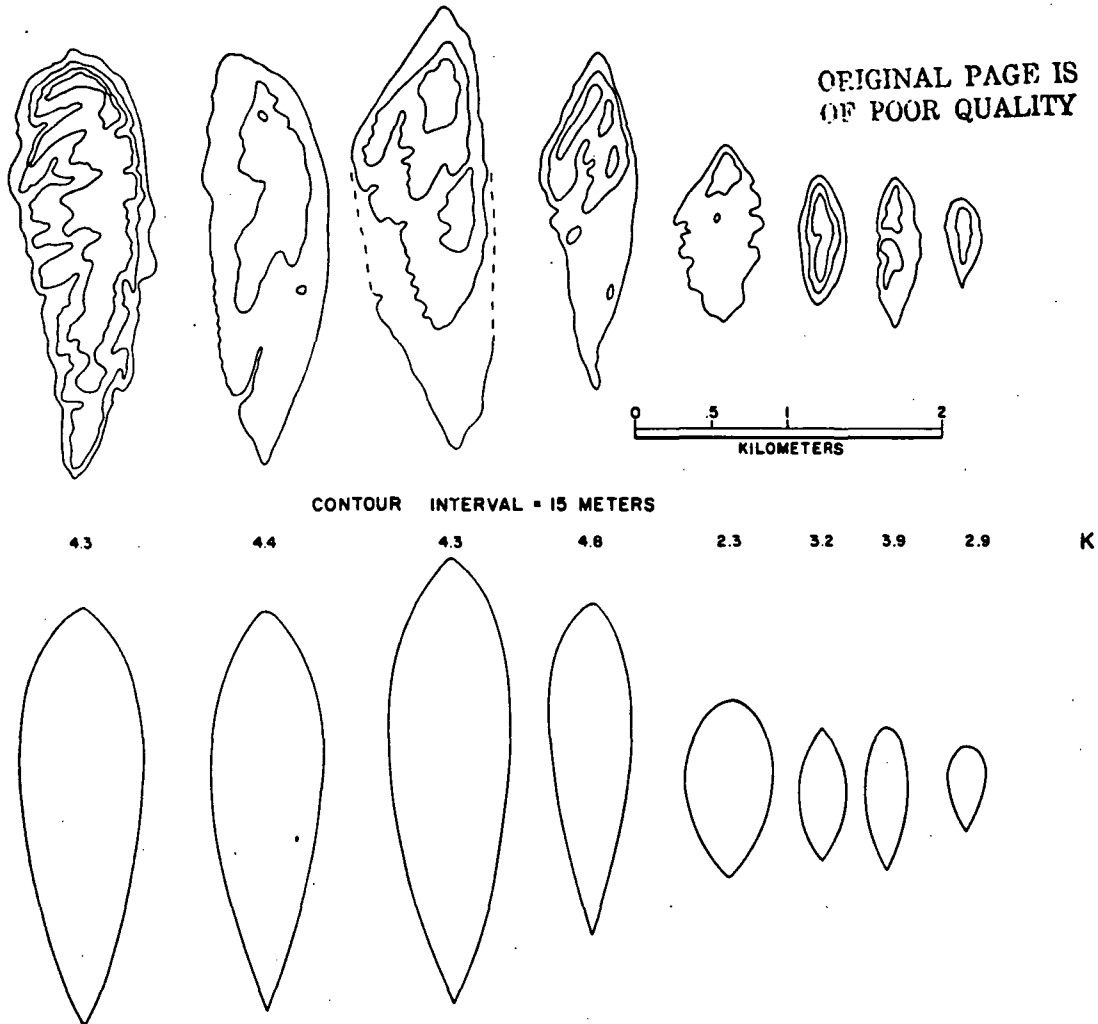


Figure 5.10. Comparison of typical streamlined scabland loess hills to equivalent lemniscate loops as calculated according to Chorley's (1959) analysis of drumlin shapes. Water flowing over and around the easily scoured loess

caused the streamlined shapes, permitting less resistance to flow. Subsequent gulleying of the soft loess has somewhat modified the streamlined shapes.

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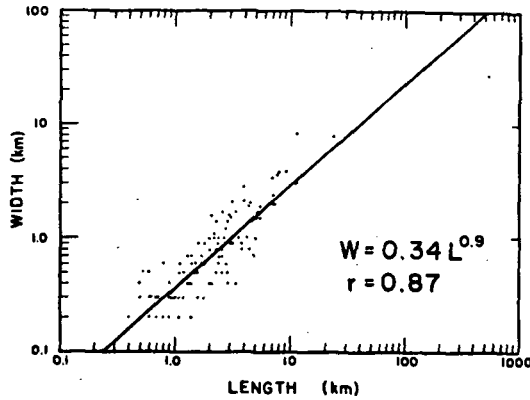


Figure 5.11. Maximum width of scabland streamlined forms versus the length of the form.

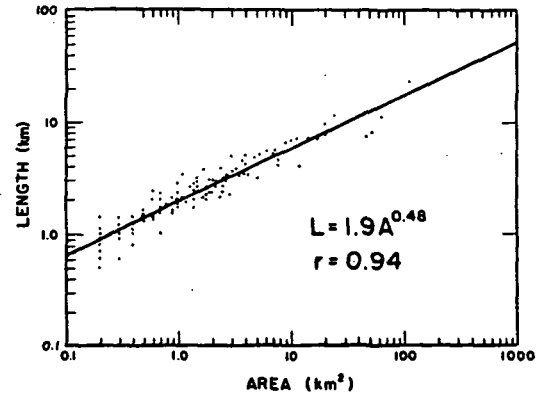


Figure 5.12. Length of scabland streamlined forms versus form area.

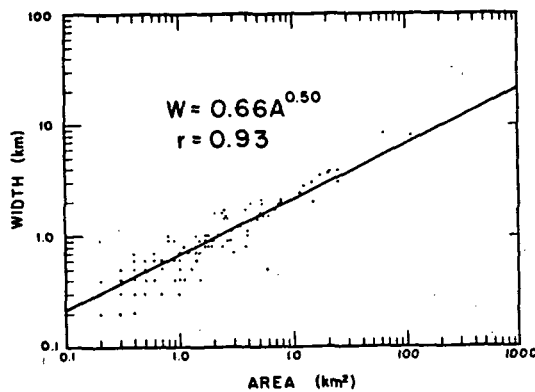


Figure 5.13. Maximum widths of scabland streamlined forms versus form area.

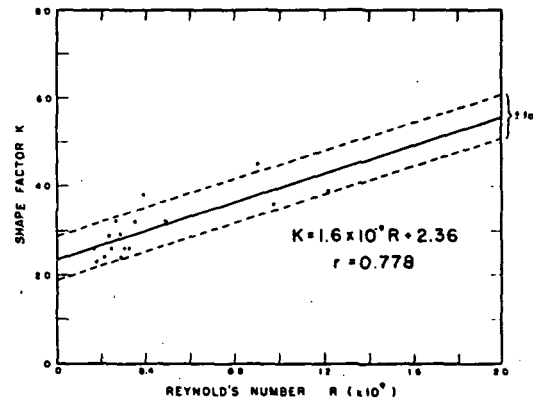


Figure 5.14. Streamlined shape factor K versus the Reynolds number for maximum scabland flood flows associated with scabland streamlined forms. The dashed lines indicate one standard error.

Longitudinal Bars

These bars are elongated parallel to the flow direction. They are characteristic of relatively uniform scabland reaches, lacking abrupt expansions and constrictions. In braided gravel rivers, longitudinal bars tend to be broad, low forms with massive bedding or crude horizontal structure within. In the Channeled Scabland, however, the longitudinal bars are mounded, streamlined forms tens of meters thick. Stratification is dominated by foreset beds that were accreted to avalanche faces on the downstream margin of the bar.

Malde (1968) introduced the term "pendant bar" to refer to streamlined mounds of Bonneville Flood gravel that occur downstream from bedrock projections on scabland channel floors. Baker (1973a) found that this was the most

common type of bar in the Channeled Scabland. The locus for bar initiation may be a knob of basalt (Fig. 5.15) or the bend of a pre-flood meandering valley (Fig. 5.16). Bar deposition was apparently initiated by gravel deposition in flow separations that developed downstream from a variety of flow obstructions (Fig. 5.17). Additional material was then added as huge foresets on the downstream margins of the bars. By this downstream accretion, scabland bars maintained a zone of flow separation that induced deposition from flows that otherwise would be competent to transport even coarse boulders.

Rather than being purely depositional forms, pendant bars are best viewed as the consequences

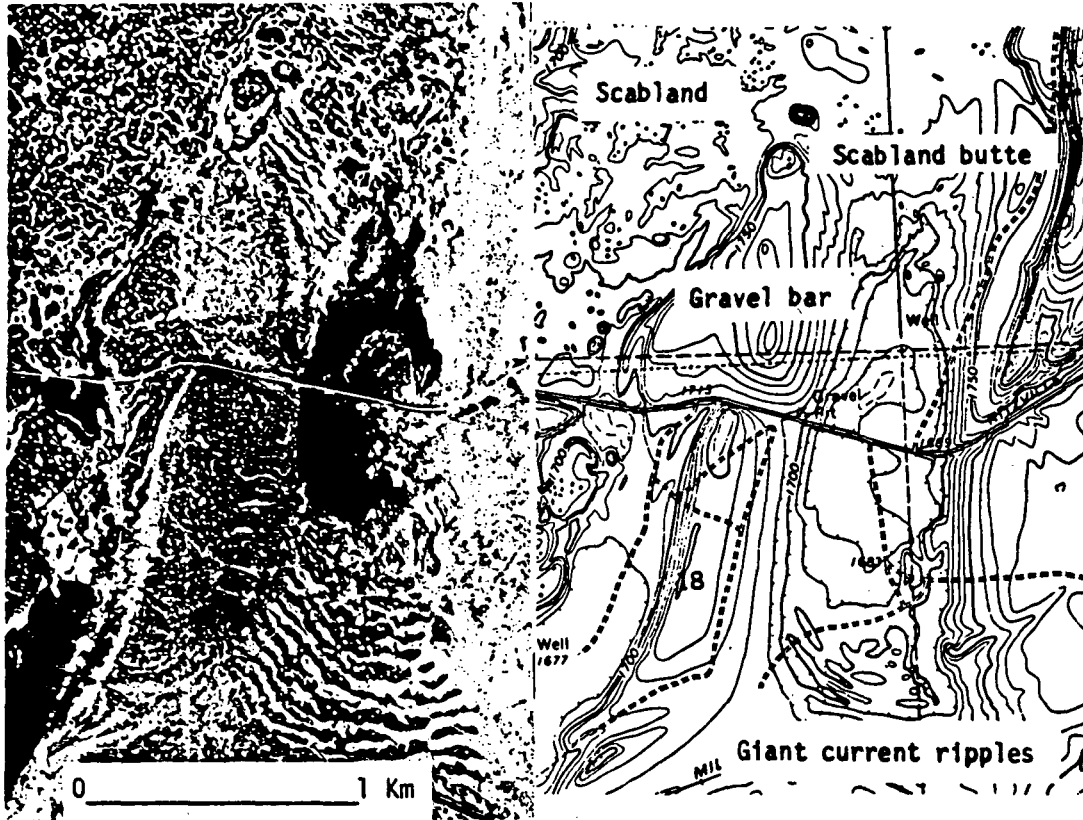


Figure 5.15. Vertical aerial photograph and topographic map of a small pendant bar in the Cheney-Palouse scabland near Macall, Washington. Contour interval is ap-

proximately 3 m (10 feet). The bar accumulated downstream from a residual butte of basalt. The bar occurs in sections 7 and 18, T. 18N., R. 38E.

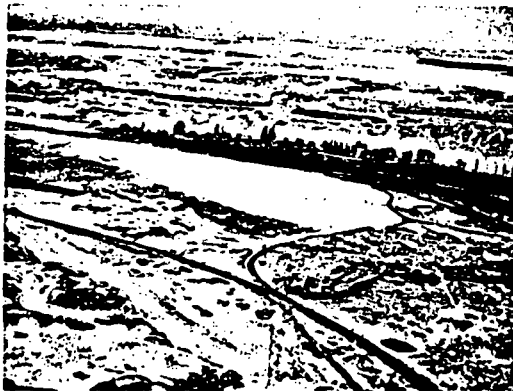


Figure 5.16. Oblique aerial photograph of Bar 2 near Wilson Creek, Washington. This is a relatively small bar about 1 km in length.

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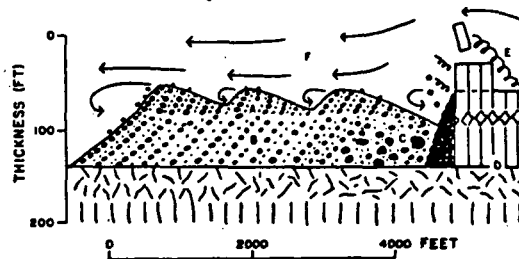


Figure 5.17. Schematic development of a hypothetical pendant bar. Bedload is transported across the surface of the bar by giant ripples and deposited at the downstream end of the bar. Explanation: A—foreset bedding associated with giant current ripples, B—foreset bedding within the bar, C—chaotically deposited flood gravel, D—basalt bedrock, E—erosion of basalt columns by kolks, F—flow directions.



Figure 5.18. Pendant bar deposits exposed near Coulee City, Washington. Note the well-sorted foresets of open-work, cobble gravel.



Figure 5.19. Large column of basalt in flood gravel comprising Bar number 3 near Wilson Creek, Washington (see description in Bretz and others, 1956, p. 977-979). Note percussion marks to right of rock hammer.

of special flow conditions that locally reduced the fantastic competency of the peak flood discharges. The transport rate of flood gravel into the separation zone downstream from the bar was simply greater than the transport rate out of that zone. The relationship of the bars to the flow conditions is evinced by the fact that they are streamlined to present minimum resistance to the flood water. Moreover, in curving reaches pendant bars never abut on the channel walls, but are separated from the walls by depression which Bretz and others (1956) termed "fosses."

Measurements of scabland bars show that, like the streamlined residual hills, they are elongated to no more than about three times their width. Thus, the pendant bars are also equilibrium forms, balancing the shear drag from skin resistance (which increases with the length of a streamlined form) and the form drag (which decreases with the width of a streamlined form). The bars grew downstream no more than three times their maximum width because the increased skin resistance produced by greater elongation would have resulted in greater transport rates away from the bar.

Pendant bars probably accumulated a rather good sample of the sediment load carried by the highly turbulent flood flows. Their internal stratification is characterized by foreset bedding (Fig. 5.18). Rapid deposition of recently eroded boulders is indicated by the inclusion of basalt

columns in the bar sediments. These columns are sometimes marked only by a few percussion impacts (Fig. 5.19).

In some scabland channel reaches, pendant bars occur in groups. Near Odessa, the resistant basaltic ring structures form resistant knobs. Many of these knobs have pendant bars trailing from their downcurrent ends (Fig. 5.20).

Other Bar Forms

The absence of linguoid and transverse bars from the Channeled Scabland has been discussed by Rust (1975, p. 246). Unlike most braided rivers, in which channels are exceptionally wide and shallow, the Missoula and Bonneville Flood channels were relatively constricted. The exceptionally deep flood water allowed bars to develop prominent slip-faces. These were the stable bedforms under extreme flood conditions. The major form of modification during falling stage has been the concentration of large boulders on the bar surface by winnowing. Most braid bars, in contrast, form by deposition on subhorizontal surfaces. Foresets are rare in gravel braided streams.

Rust (1975) believes that gravel braid bars are like the scabland bars in that they form initially as primary bedforms. They are stable under the flood flows in which all bed material is in motion. In contrast, however, many sand bed and

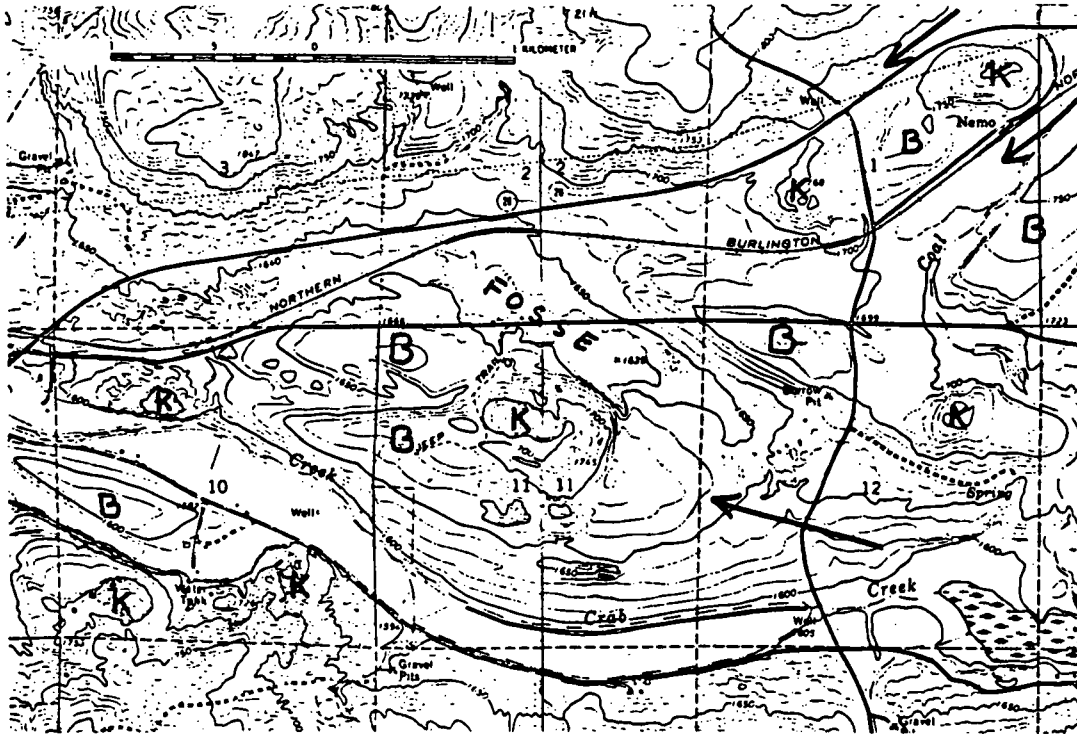


Figure 5.20. Topographic map of pendant bars (B) near Odessa, Washington. The prominent basalt knobs (K) result from the resistance of circular dikes intruded into

the jointed lava flows. Contour interval is 3 m (10 feet). Topography is from the Odessa and Sylvan Lake, Washington, 7.5-minute quadrangles.

some gravel bed rivers show extensive bar modification during changing river stages. These complex, modified bar forms may characterize the great expansion bar complex of the northern Quincy Basin (Baker, 1973a, p. 39-42).

Eddy Bars

These bars occur at the mouths of alcoves or valleys that were tributary to valleys invaded by catastrophic flood flows. They are particularly well-developed along the eastern margins of the Cheney-Palouse scabland tract. Nearly every drainage entering the scabland tract is blocked at its mouth by an eddy bar. Excellent examples occur at Willow Creek (described by Bretz and others, 1956, p. 1031-1033) and at the mouth of the Tucannon River.

The contrast between the internal structures of pendant and eddy bars can be seen by comparing Figures 5.18 and 5.21. Pendant bars (Fig. 5.18)



Figure 5.21. Eddy bar deposits at the mouth of the Tucannon River. The lowest layer of horizontally bedded sand is overlain by about 60 cm of laminated silt. This is overlain by poorly sorted, angular to subangular flood gravel containing percussion flaked boulders and cobbles. The prominent boulders of silt are identical to the underlying laminated silt layer. They may be "rip-off" clasts such as commonly found in turbidites.

show well developed foreset bedding. Individual foresets contain well sorted subrounded boulders and cobbles with an open-work structure. Other foresets are entirely made up of granules and pebbles. Eddy bars (Fig. 5.21) contain a variety of grain sizes and structures. Interfingering occurs between poorly sorted boulder gravel, laminated silts, cross-bedded granule gravel, and graded sand-silt layers. The boulders and cobbles are marked by percussion flake scars. Many are "broken rounds," as described by Bretz (1929).

Foreset bedding in the pendant bars almost always dips downstream. Bedding in the eddy bars indicates varying directions of sediment transport. Crude foresets in the boulder gravels usually dip away from the main scabland channel. Frequently, however, the smaller foresets in the granule gravels dip back toward the main scabland channel. This pattern may be the result of the swirling eddies which deposited the bars. The stronger currents carried the coarsest flood debris up the tributary valley. Weaker back flow currents then deposited the finer granule gravels.

The swirling currents which resulted in eddy bars are another manifestation of macroturbulent phenomena as described by Matthes (1947). As suggested by Krumbein (1942), extremely high turbulence may prevent the equilibrium conditions that serve to distinguish bedload and suspended load. This may be the origin of the extremely poor sorting of eddy bar deposits when compared to pendant bar deposits. Another result may be the fact that giant current ripples are never associated with eddy bars. Because macroturbulent phenomena are not well understood a precise origin of eddy bars cannot be described.

The sedimentologic features noted for eddy bars may be seen for many kilometers upstream along scabland tributaries. However, the abundance of coarse materials decreases in an upstream direction relative to sand and silt. These deposits upstream from the eddy bar blockade are termed "slackwater deposits." Slackwater deposits are distinct from the main channel deposits, but it must be remembered that there is a complete lateral gradation between the two.

BEDROCK MESOFORMS

The bed forms eroded by the macroturbulent flood flows on bare basalt surfaces are of immense variety and unique size in the Channeled Scabland. Indeed these are the features most characteristic of "scabland." Bretz' studies of these bizarre erosional forms were largely descriptive. It is now apparent, however, that the rock bedforms exist in an evolutionary sequence that is related both to the flood flow hydrodynamics and to the resistant characteristics of the jointed basalts. Here the forms will first be described, and then their evolutionary sequence will be discussed.

Cataracts and Inner Channels

The most impressive erosional forms created by the Missoula floods are probably the abandoned cataracts (Fig. 5.22). Most famous of these is Dry Falls (Fig. 5.23), 5.5 km wide and 120 m high. Bretz and others (1956, p. 1029) recognized that many of the cataracts were formed subfluentially rather than by the plunge-pool undercutting classically illustrated by Niagara Falls. This contention is supported by the high-water mark evidence left by the early Pinedale (late Wisconsin) flood (Baker, 1973a; his Fig. 7). Bretz (1932a) described in detail the initiation of a 250 m cataract near Coulee City, Washington, and its 32 km upstream recession to create the upper Grand Coulee (Fig. 5.24). The unique capacity of vertically jointed basalt to maintain the lip of a recessional cataract was held to be the primary consideration in this type of erosion.

All scabland cataracts show multiple horse-shoe-shaped headcuts. Potholes and Frenchman Springs have two parallel headcuts or alcoves (Fig. 5.25). At the base of each alcove is a large closed depression. Even with post flood modification, the closure in these depressions is as much as 35 m. Probably the cataract headcuts acted as efficient funnels during the maximum flood flows. The water surface was sharply drawn down over the cataract producing intense macroturbulent scour beneath the locally steep water-surface gradient. Plucking erosion was concentrated in the columnar-jointed zones, and large blocks of entablature were undermined at the cataract lip.

The headward recession of scabland cataracts produced distinct inner channels. Field mapping shows that the margins of these channels and the

cataract lips are held up by relatively resistant basalt entablature (Fig. 5.26 A and B). Bretz and others (1956) envisioned powerful kolks at the plunge pool locations, undermining the cataract lip. Sediment, usually basalt columns and large blocks of entablature, were transported away from the cataract lip in a state of quasi-suspension, buoyed by the intense macroturbulence (Baker, 1973a, p. 26-29). The competency of the kolks is difficult to estimate theoretically. Boulders up to 30 m in diameter (estimated from aerial photography) were transported from the lip of West Potholes cataract. Certainly the vertical vortices (kolks) were exceedingly powerful agents of lift and transport.

In the upper Grand Coulee (Fig. 5.24) cataract recession was initiated at the structural step provided by the Coulee Monocline. The undermining and recession process proceeded rapidly enough for the cataract to recede 32 km to the gorge of the Columbia River on the northern margin of the Columbia Plateau. On the western rim of the Quincy Basin (Fig. 5.25) cataract recession was initiated at the canyon walls carved by the Columbia River.

Many scabland channels were excavated from preflood stream valleys. At Lenore Canyon and Moses Coulee (Fig. 5.27) the preflood tributaries now form hanging valleys that drain Palouse loess topography unmodified by catastrophic flooding. The floors of these valleys were deepened by flooding, often with the production of distinct inner channels (Fig. 5.28).

Potholes and Buttes

Perhaps the most prevalent topographic form in eroded rock of the scabland tracts is butte-and-basin topography. The usual development is small anastomosing channels and rock basins surrounding buttes and mesas, with a total relief of 30-100 m (Fig. 5.29). The rock basins range in size from shallow saucers to the scale of Rock Lake,

Figure 5.22. Oblique aerial photographs of scabland cataracts and inner channels. A. West Potholes cataract on the western rim of the Quincy Basin. B. Frenchman Springs cataract also on the western rim of the Quincy Basin. C. Hudson cataract at the head of Hudson Coulee in the Hartline Basin. D. Palouse Falls, a small cataract in the inner channel eroded through the Palouse-Snake divide crossing (the Palouse River now occupies this flood channel).



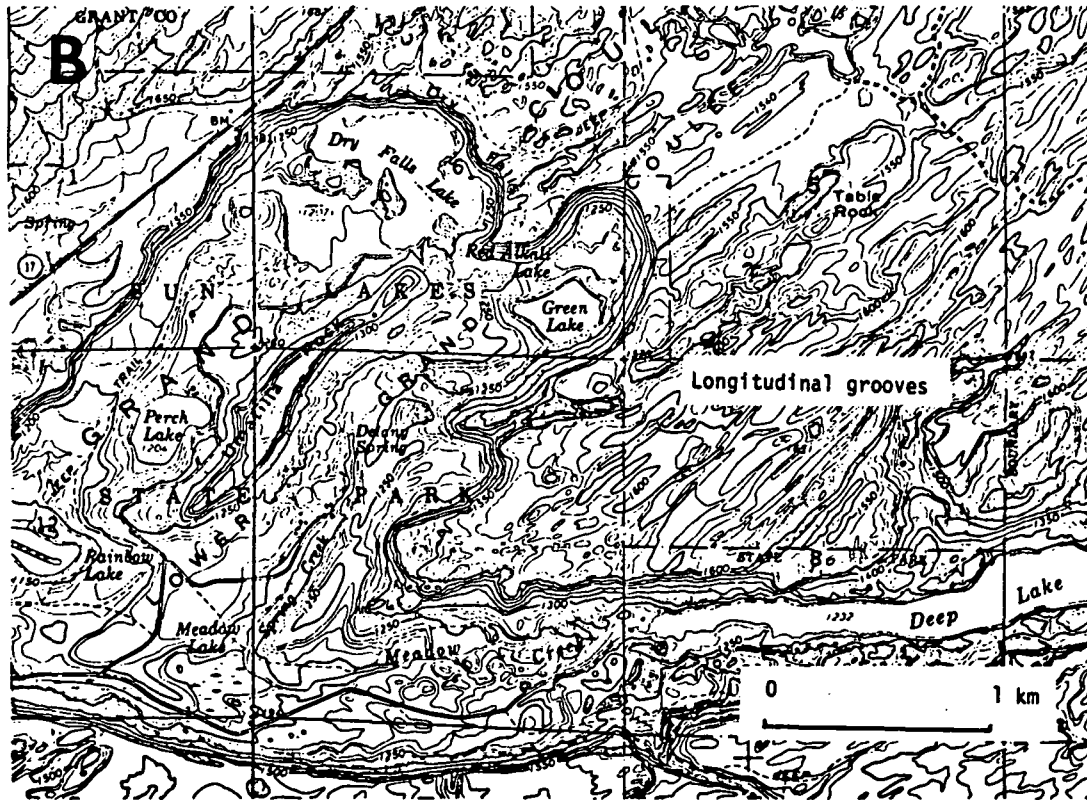
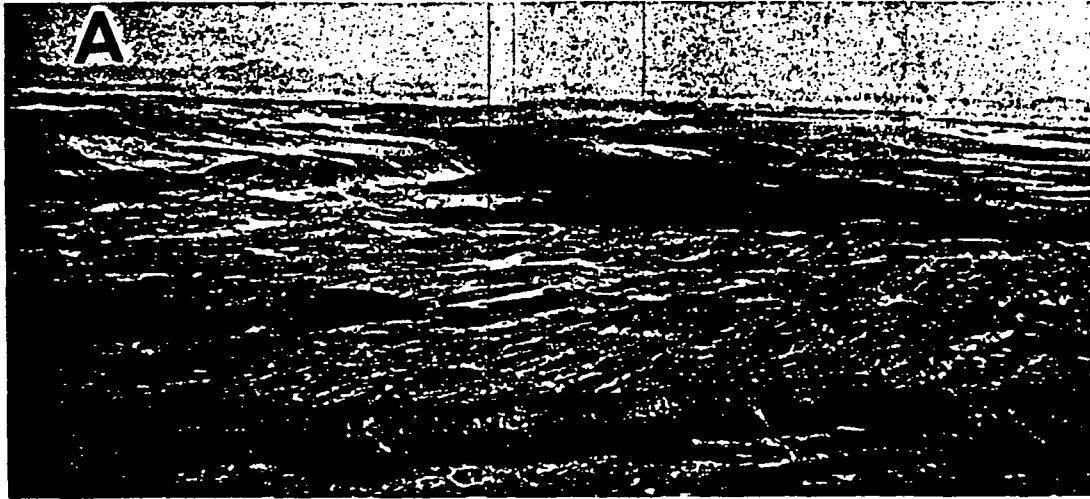


Figure 5.23. A. Oblique aerial photograph of the Dry Falls cataract group. The cataract is 120 m high and 5.5 km wide. Longitudinal grooves are visible just upstream from the cataract head, and the upper Grand Coulee (containing Banks Lake) extends to the horizon.

B. Topographic detail of the Dry Falls area showing the inner channel development (contour interval is 10 feet; elevations in feet). Note topographic expression of longitudinal grooves.

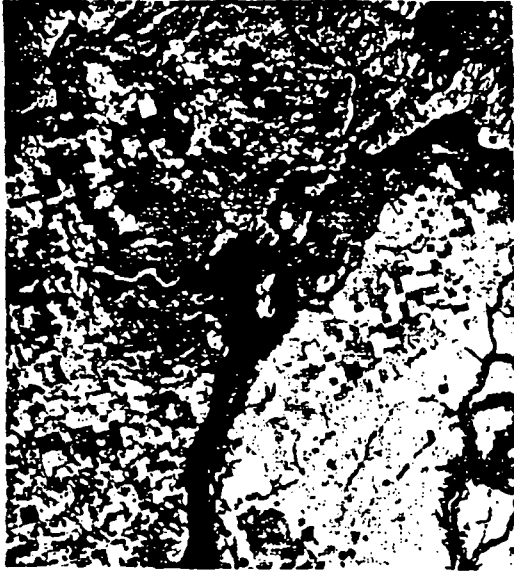


Figure 5.24. Orbital photograph of the upper Grand Coulee showing the immense inner channel left by headward retreat of the 250 m deep cataract. LANDSAT image E-2 936-17451-5.

11 km long and 30 m deep. Bretz (1932a, p. 26-28) described this combination of features as follows: "The channels run uphill and downhill, they unite and they divide, they deepen and they shallow, they cross the summit, they head on the back-slopes and cut through the summit; they could not be more erratically and impossibly designed."

Some typical scabland potholes are shown in Figure 5.30.

Longitudinal Grooves

Where broad expanses of a single basalt surface were eroded by catastrophic flood flows, the characteristic form of the erosion is a series of elongate grooves. These features generally have their long axes parallel to the prevailing flood flow streamlines. The grooves are common on basalt entablature surfaces in the Hartline Basin (Fig. 5.23). They average 5 m in depth and 50 m in width. Relatively large grooves are developed near Palm Lake in the Cheney-Palouse scabland tract (Fig. 5.31).

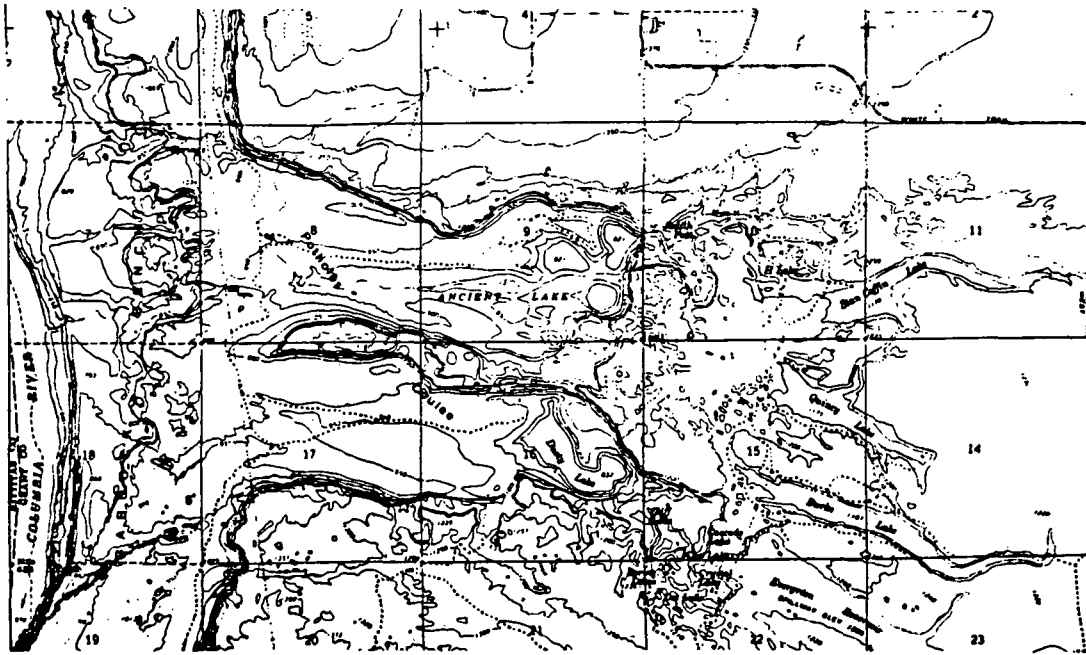


Figure 5.25. Topographic map of the Potholes Cataract. Topography is from the Babcock Ridge 7.5-minute quadrangle.

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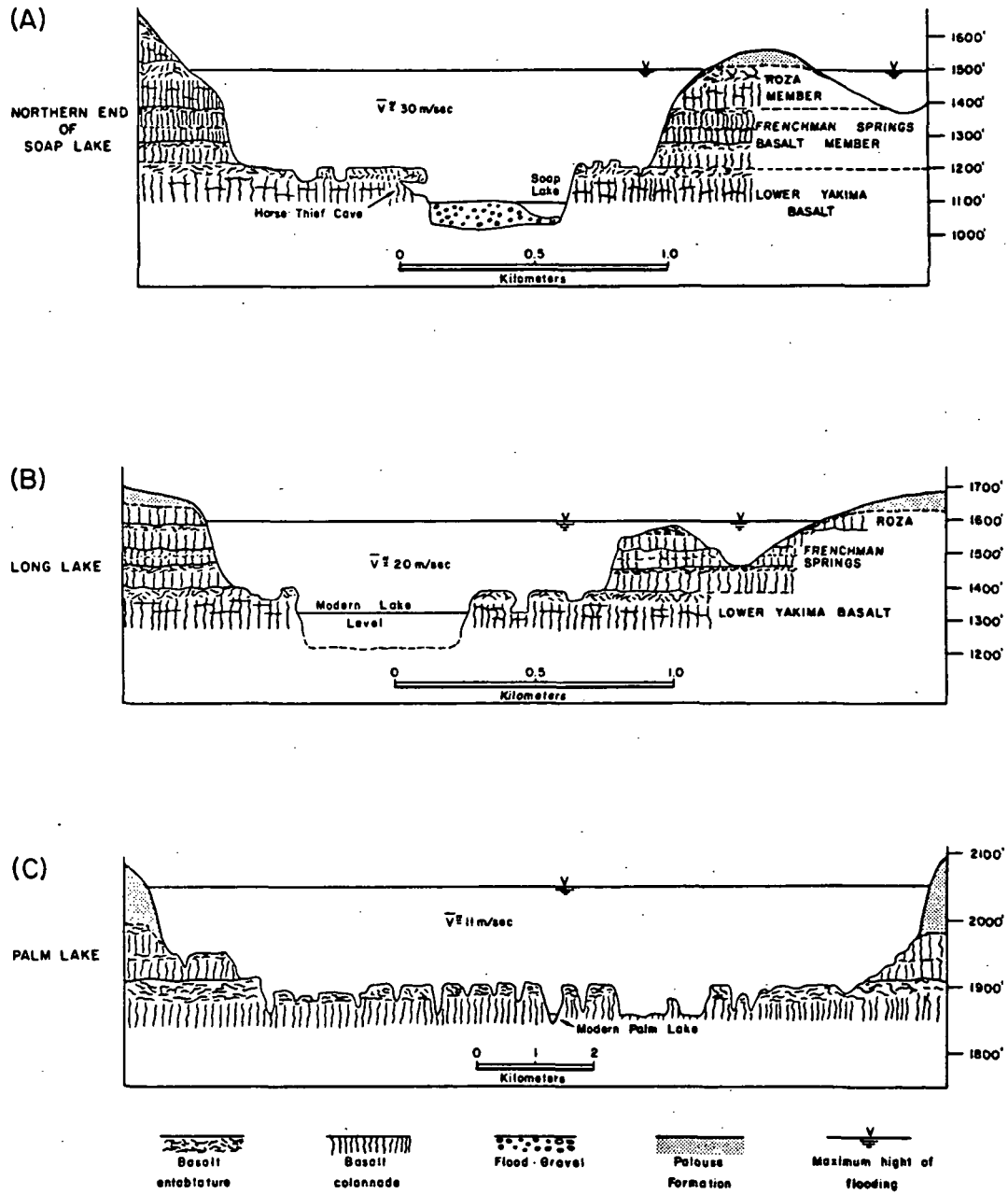


Figure 5.26. Representative cross sections of scabland channels showing structural characteristics of basalt, erosional features, and flood stage plus maximum mean flow velocity (\bar{V}). The sections at Soap Lake (A) and Long Lake (B) show inner channels with flanking butte-and-basin scabland topography. The section at Palm Lake (C) shows longitudinal groove topography.

Sequence of Erosional Forms

Recent experimental studies of fluvial erosion utilizing simulated bedrock (Shepherd, 1972; Shepherd and Schumm, 1974) indicate that a sequence of erosional bed forms may develop in bedrock as a function of time. First to appear in these experiments were the faint streaks of longitudinal lineations associated with potholes and transverse erosional ripples. The lineations then became enlarged into prominent longitudinal grooves. Eventually the grooves decreased in number, and finally one narrow, deep inner channel formed. In the experiments the inner channels were incised below base level, and nickpoints migrated headward upstream (see Schumm and Shepherd, 1973, p. 7, for a longitudinal profile). Deposition of bedload occurred downstream from the headcut, culmination in a low-gradient inner channel with a sand bed and simulated bedrock banks. The lower part of this channel was actually an elongate basin, incised below base level.

Shepherd (1972) observed that some of the scabland erosional topography described by Baker (1973a) was analogous to features produced in the flume experiments. The most important differences derive from the differential erosional resistance provided by the jointed basalt. As shown by Ippen and others (1962) this effect is a predictable consequence of the high shear stress gradient that occurs on the inside of curves in an experimental trapezoidal channel.

The progressive erosion of a typical scabland divide crossing (Fig. 5.32) is envisioned as follows. The first flood water to overtop a divide encountered soft Palouse loess and Ringold Formation (Phase I). The high velocity water quickly exposed the underlying basalt, leaving an occasional streamlined loess hill as a remnant of the former cover (Phase II). The entablature of the uppermost basalt flow was then encountered. This probably yielded to groove development, possibly associated with longitudinal roller vortices. The first exposure of well-developed columnar jointing, perhaps at the top of a flaring colonnade along the irregular cooling surface, introduced a very different style of erosion (Phase III). Large sections of columns could now be removed at this site with the simultaneous development of vertical

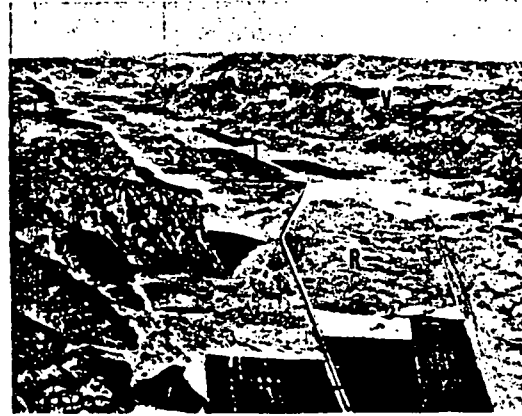


Figure 5.27. Inner channel development in Moses Coulee. This oblique aerial photograph shows a distinct inner channel (I) on the coulee floor with marginal butte-and-basin scabland (S). Giant current ripples (R) occur on the surface of a flood gravel bar at the mouth of the inner channel. Preflood tributaries to the drainage were truncated by flood erosion to form hanging valleys (V) separated by truncated spurs.

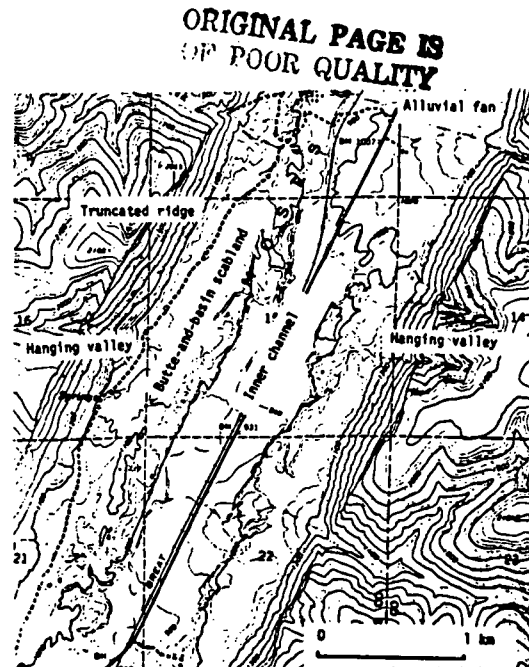


Figure 5.28. Topographic map of lower Moses Coulee showing the relationships discussed in Fig. 5.27. The contour interval is 10 feet (elevations in feet).



Figure 5.29. Butte-and-basin scabland developed 12 km south of Creston in the Telford-Crab Creek scabland tract. Topography (contour interval is 10 feet) is from Telford, Washington, 7.5-minute quadrangle.

vortices (kolks). With the enlargement and coalescence of the resultant potholes, the surface assumed the bizarre butte and basin topography that characterizes much of the Channeled Scabland (Phase IV). The eventual topographic form was the development of a prominent inner channel (Phase V). Such inner channels may have been initiated at downstream structural steps in the basalt, and then migrated headward by cataract recession. The lateral enlargement of inner channels probably proceeded by the undercutting of resistant entablature as columns were plucked out by kolks. Horsethief Cave, north of Soap Lake, is an excellent example of this type of erosion.

SCOUR MARKS

Large boulders on scabland bars and bedrock projections along scabland channels produced a

deformation of flood flow streamlines that resulted in distinctive erosional scour marks. Engineers have studied the hydrodynamics of such scour in order to protect bridge piers during floods (Laursen, 1960). The scour is generated by two basic systems of vortices (Shen, 1971): the horseshoe-vortex system and the wake-vortex (Fig. 5.33A).

The prominent crescentic scour hole on the northwest (up-stream) side of the boulder in Figure 5.33B was probably caused by the hydrodynamic stretching and accumulation of vortex filaments in the front of the boulder. Richardson (1968) described this process as a characteristic effect of a blunt-nosed obstacle on an approaching two-dimensional velocity field. The strong pressure field produced by the blunt obstacle causes a separation of the boundary layer which then rolls up ahead of the obstacle to form a horseshoe vortex. Karcz (1968) suggested that this mechanism is responsible for the current crescents that commonly occur upstream from obstacles.

The large elliptical scour hole that formed on the downstream side of the boulder in Figure 5.33B formed as a result of a wake-vortex system generated by flow separation in the rear of the boulder. Engineering experiments have shown that the wake vortex system is a function of Reynold's number (velocity of approach times obstacle diameter times fluid density, divided by dynamic viscosity of the fluid). With other variables in the Reynold's number held constant, increasing velocity results first in a pair of vertical vortices. When the flow changes from laminar to turbulent, vortices form and migrate downstream. Karcz (1968) notes that the average velocity in the wake region is quite low at this stage and deposition is likely in the lee of the obstacle. Indeed the scablands contain many examples of pendant bars that were deposited as elliptical forms in the lee of obstacles. Malde (1968) described similar features that occurred during the Bonneville Flood in the Snake River Plain of Idaho.

At very high flow velocities, vortex intensity grows and the sucking action of kolks dominates. Shen (1971, p. 23-25) observed that the wake-vortex system then acts like a vacuum cleaner in removing bed material at this stage. Thus, the scour hole in the lee of the boulder may indicate high flow velocities. Baker (1973a, p. 41-42)



Figure 5.30. Butte-and-basin topography near Long Lake (A) and Blue Lake (B). Potholes show undercutting

of their entablature rims particularly on the downstream side.

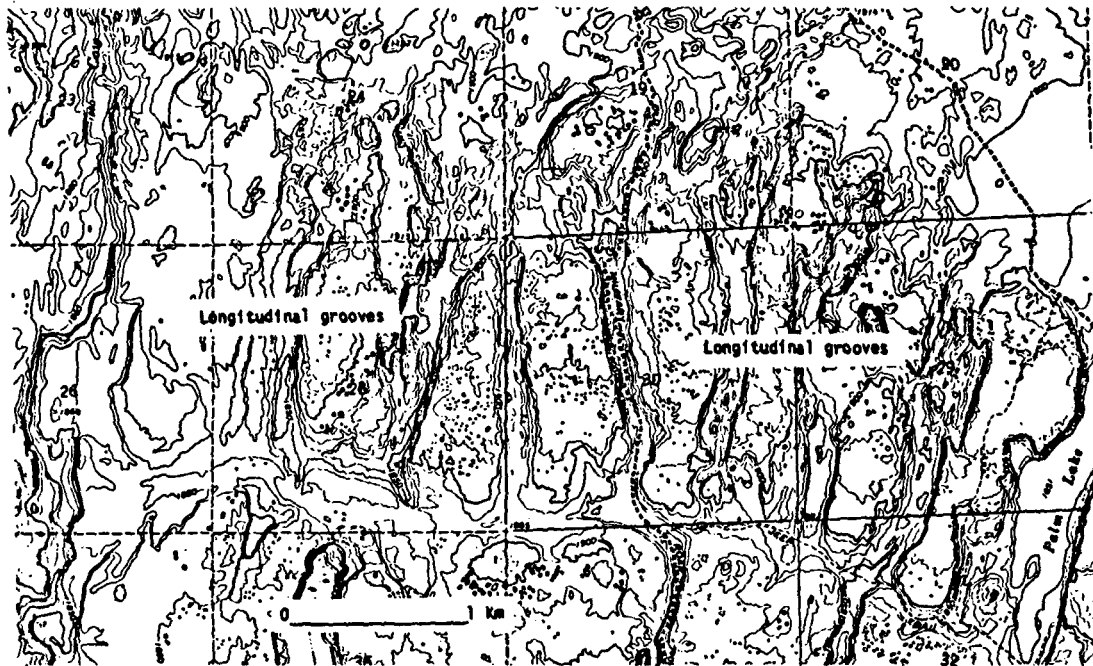
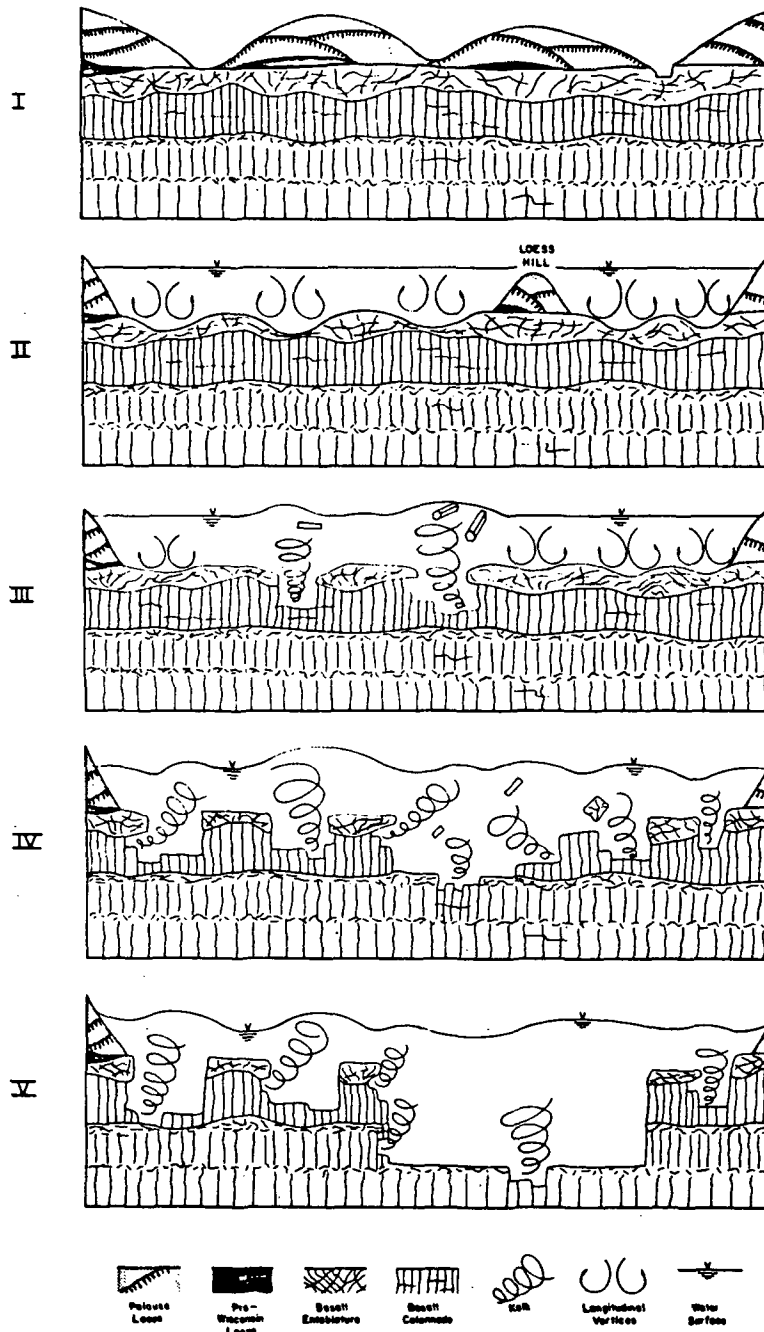


Figure 5.31. Longitudinal grooves developed on the scabland surface near Palm Lake. The small closed contours are mostly mounds of post-flood silt. The larger closed depressions are scabland basins. Topography is from the Palm Lake, Washington, 7.5-minute quadrangle.

suggested that the high flow velocities were generated during the draining of the Quincy Basin in the waning stages of the flood. This explanation is also consistent with the prominent armoring that characterizes the Quincy Basin fill.

GIANT CURRENT RIPPLES

Aerial photographs of some scabland gravel bars reveal patterns of parallel ridges and swales which Bretz and others (1956) identified as "giant current ripples." These constitute the most important bed forms used in the paleohydraulic reconstruction of the last major scabland flood (Baker, 1973a). Over 100 sets of these bed forms have been identified in various Missoula Flood channel ways. Figure 5.34 illustrates some of the variety that exists in ripple morphology.



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Figure 3.32. Hypothetical sequence of flood erosion for a typical scabland divide crossing. See text for discussion.

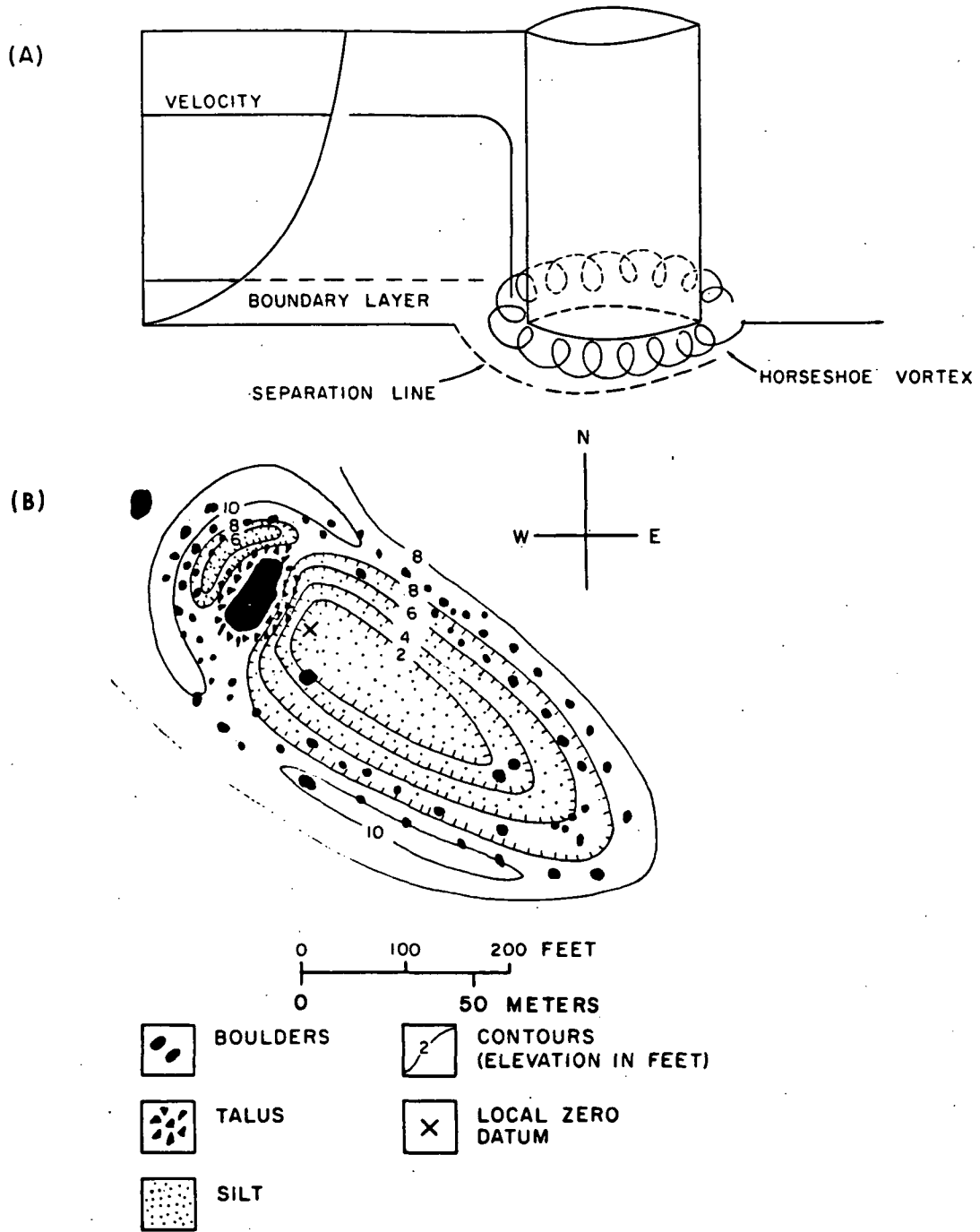


Figure 5.33. A. Formation of a horseshoe-vortex system at the front of a vertical cylinder mounted on an experimental flume bed (after Moore and Masch, 1963).

B. Scour hole development near an 18 x 11 x 8 m boulder (Fig. 5.15) 2.5 km west of Rocky Ford Fish Hatchery, Ephrata, Washington (Baker, 1973a).

Ripple Morphology

Measured heights and chords of the scabland ripple fields show a remarkable symmetry of form. Mean ripple heights for a ripple field are closely related to mean ripple chords (Fig. 5.35). The chords generally range from 20 to 200 m. Ripple heights, ranging from 1 to 15 m, have probably been somewhat reduced by waning flood stages and by the modification of post-flood processes.

The giant current ripples present an asymmetric appearance in profile (Fig. 5.36). The downstream-facing slopes of the ripples (lee sides) average about 18 to 20°. Upstream-facing slopes (stoss sides) average about 6 to 8°. Undoubtedly these slopes have been somewhat modified from their original depositional form.

In plan view the giant ripples show a form (Fig. 5.37) that is classified according to Allen (1968, p. 65) as transverse catenary and out of phase. The lee slopes of the ripples occur in a series of cusped troughs on the down-current side, morphologically analogous to the troughs that form on the leeward sides of some transverse wind dunes (Fig. 5.38). The ripple chords also tend to decrease away from deeper (higher velocity) portions of the channel.

The giant current ripples are very difficult to recognize on the ground unless the observer is paying close attention to minor ridges and swales on scabland depositional surfaces. This fact prevented the recognition of these bed forms until extensive aerial photography of the region was undertaken for the Columbia Basin irrigation project (Bretz and others, 1956). The ripple patterns on aerial photographs derive from a combination of localized eolian silt deposition and vegetation patterns (Fig. 5.39). Ripples near Spirit Lake, Idaho have a cover of pine (*Pinus*

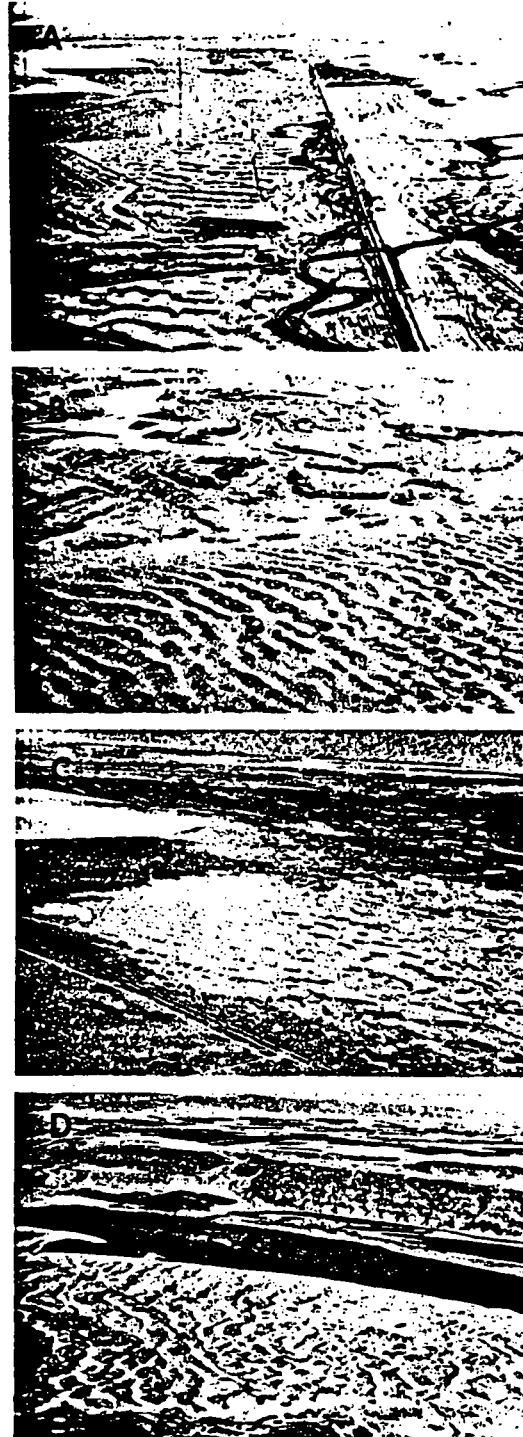


Figure 5.34. Typical sets of giant current ripples in the Channeled Scabland: A. Lind Coulee, B. Marlin, C. Artesian Lake, D. West Bar. Locations for these ripple trains are given in Baker (1973a).

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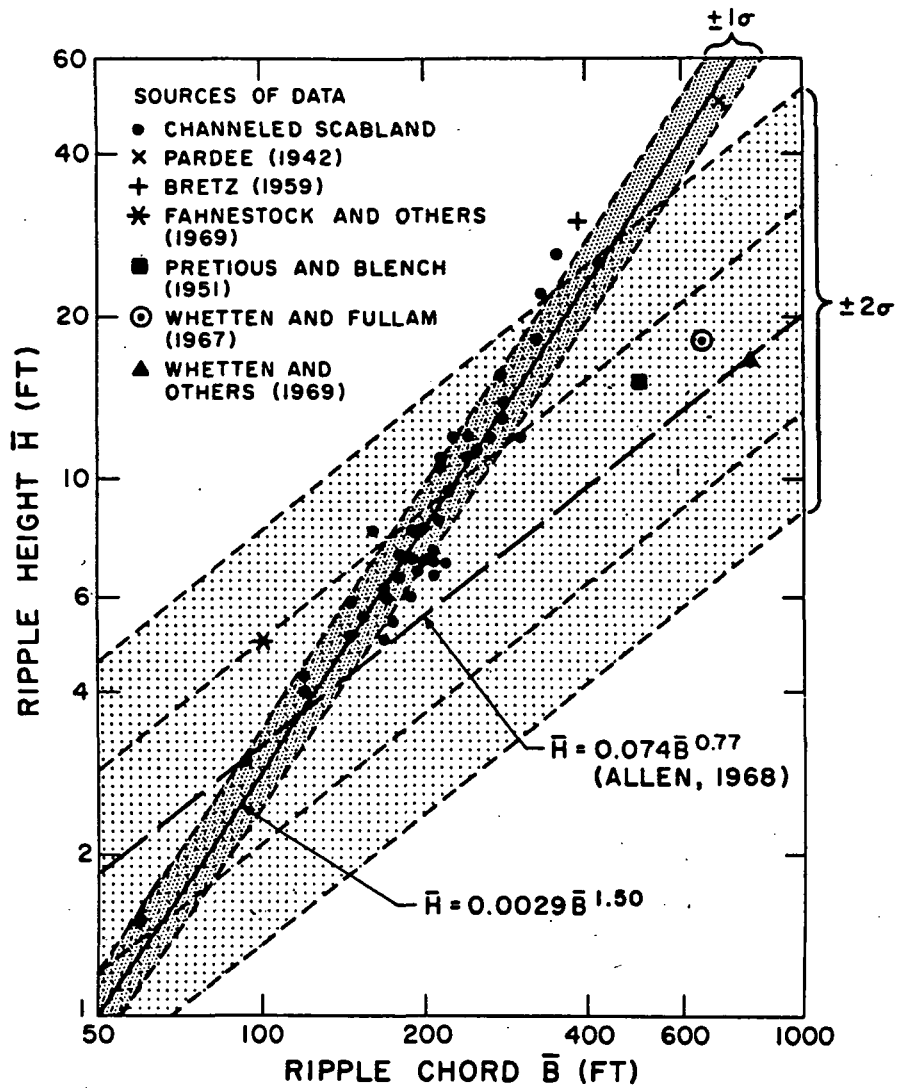


Figure 5.35. Logarithmic relationship of height as a function of chord for 40 sets of giant current ripples. The general relationship for large-scale current ripples determined by Allen (1968) is shown by a dashed line.

Five additional measurements by various authors are also plotted. Standard errors on the regressions are indicated by the letter σ .

ponderosa) that reflects the higher rainfall of that region (Fig. 5.39A). In drier parts of the Columbia basin eolian silt fills the swales between ripple crests. This silt supports grass while the adjacent ripple crest supports sagebrush (*Artemisia tridentata*) (Fig. 5.39B). The contrast between these two plants results in especially striking patterns.

In some local areas, the postflood loess deposition has completely buried giant current ripples. At TSCR ripples near the Tokio grain elevator the silt has built up to form ridges in the former swales between the ripple crests. These ridges mimic the ripple pattern that is buried beneath (Fig. 5.39C), but the image is reversed. The modern swales lie over the former ripple crests, and the modern ridges (of silt) occur along the former ripple troughs.

Over the last 10 years high-quality large-scale topographic maps have become available for most of the Channeled Scabland. These maps clearly show the superposition of giant current ripples on scabland bars (Fig. 5.40).

Internal Structure and Sediments

The sediment comprising the giant current ripples is some of the coarsest known to occur in large-scale depositional bed mesoforms (Fig. 5.41). The largest particles may be 1.5 m or greater in diameter, and the median size generally occurs in the pebble fraction. In all observed examples less than ten percent of the sediment is finer than granule gravel.

Internally the ripples consist of foreset-bedded gravel deposited at an angle of about 27°. Exposures (Fig. 5.42) show that individual foresets are remarkably well sorted. Layers of cobbles alternate with discrete layers of granules or pebbles. The sorting gives the gravel a distinctive open-work texture.

Trough-filling cross stratification (Fig. 5.43) is relatively rare in scabland bars of flood gravel. Where present, this stratification type probably represents a filling of scour holes or the migration of sinuous crested giant current ripples.

The coarsest fraction of the ripple sediments, boulders and cobbles, generally form an armor on the ripple stoss slopes. The armor is in the form of an imbricate pavement that probably acted to decrease flow resistance on the ripple surface during the waning stages of flood flow.

This smoothed surface may be partially responsible for the preservation of the ripples. In most rivers the depositional mesoforms (usually composed of sand) are washed out during waning flood flow stages (Jackson, 1975).

Hydraulic Significance

Bretz and others (1956, p. 980) suggested "An interesting sidelight on the hydraulics of these glacial rivers will appear when the giant current ripples are given careful detailed study." Baker (1973a) analyzed 43 sets of giant current ripples in Missoula Flood reaches. Statistical correlation and regression analyses were used, treating the values of mean ripple height (H) and chord (B) as the dependent variables. The independent variables were the various hydraulic parameters calculated for each reach. These results will now be summarized.

Depth

Depth is defined as the difference in elevation between the ripple field and the high-water surface above that field. Therefore, this is the maximum depth achieved during the passage of the flood through the reach.



Figure 5.36. Oblique aerial photograph of giant current ripples 3 km west of Odessa, Washington. The current flowed from right to left in the photo, creating gravel bed forms with a mean height of 2.6 m and a mean chord 66 m. The west-facing lee sides of these ripples have an average slope of 18.2 degrees. The east-facing stoss sides of the ripples have an average slope of 6.5 degrees.

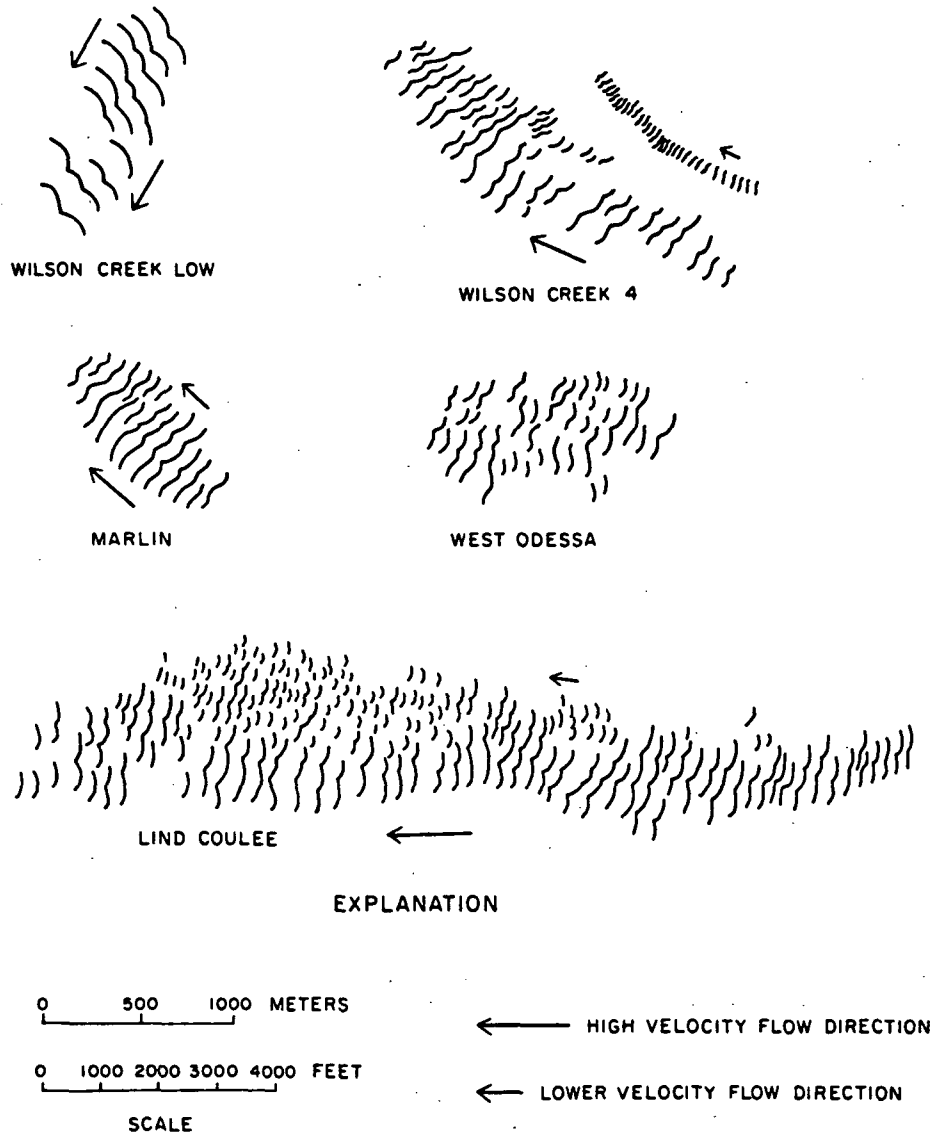


Figure 5.37. Giant current ripple trains viewed in plan. These patterns of ripple crests, traced off vertical aerial photographs, show a decrease in bed form size away from the thread of maximum flow velocity.



Figure 5.38. Oblique aerial photograph of the giant current ripples at Malaga, Washington (Sec. 28, T. 22N., R. 21E). Note the sinuous crests and cusped troughs that are accentuated in this low sunangle photograph. These ripples were created by catastrophic flooding down the Columbia approximately 13,000 years B.P. (photograph by David A. Rahm).

In Figure 5.44, mean ripple height for a given ripple train is plotted as a function of maximum depth. In Figure 5.45, mean ripple chord is plotted as a function of depth. Both relationships show relatively low correlation coefficients (0.648 and 0.758 respectively). The broad bands defined by one standard error to either side of the regression lines are a further measure of the scatter in the data. The somewhat greater correlation of ripple chord to depth, in contrast to height versus depth, probably arises from the fact that original depositional ripple chords are less modified by waning flows than are ripple heights.

Depth-Slope Product

The product of the slope of the high-water surface over a ripple field and the depth defines a second hydraulic parameter:

$$DS = \tau/\gamma,$$

where D is the depth, S is the slope, τ is the shear stress, and γ is the specific weight of the fluid ($9.8 \times 10^3 \text{ N/m}^3$ for clear water). "Depth-slope" represents a maximum shear stress achieved by the passage of the flood through the reach containing the giant current ripples.

In Figure 5.46 mean ripple height is plotted as function of depth-slope. Figure 5.47 shows mean ripple chord versus depth-slope. The correlation coefficient is greater for chord than for the height

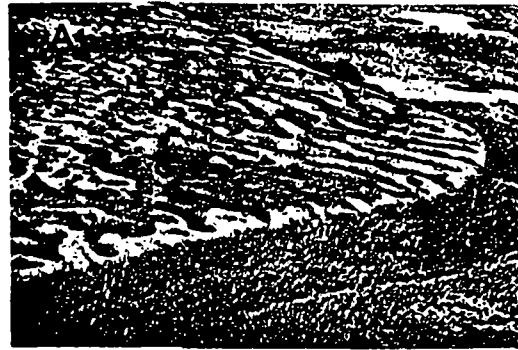


Figure 5.39. Vegetation patterns associated with giant current ripples: A. Spirit Lake, Idaho, B. West Bar, showing appearance of ripples to an observer standing on ripple crest, C. TSCR ripples.

(0.945 and 0.931 respectively). In contrast to the depth correlation, however, 89% of the variation in ripple chord is explained by the depth-slope product. Figure 5.46 shows that the equation:

$$\bar{H} = 24.6 (DS)^{1.17} \quad (5-10)$$

may be used within the standard error indicated

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Figure 5.40. Topographic map of giant current ripples shown in Fig. 5.36.

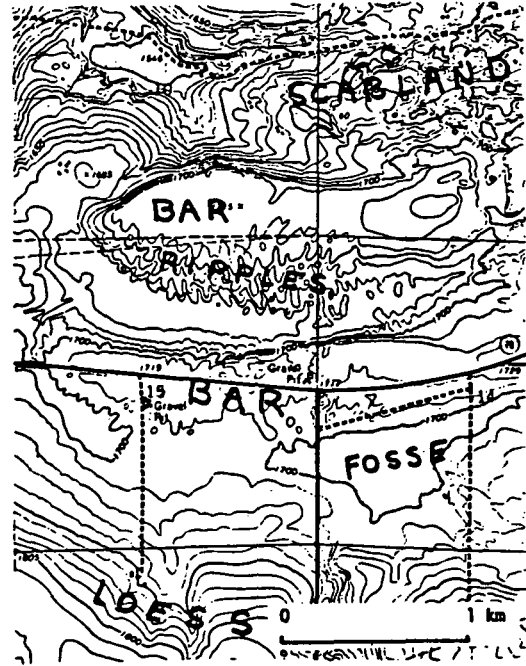
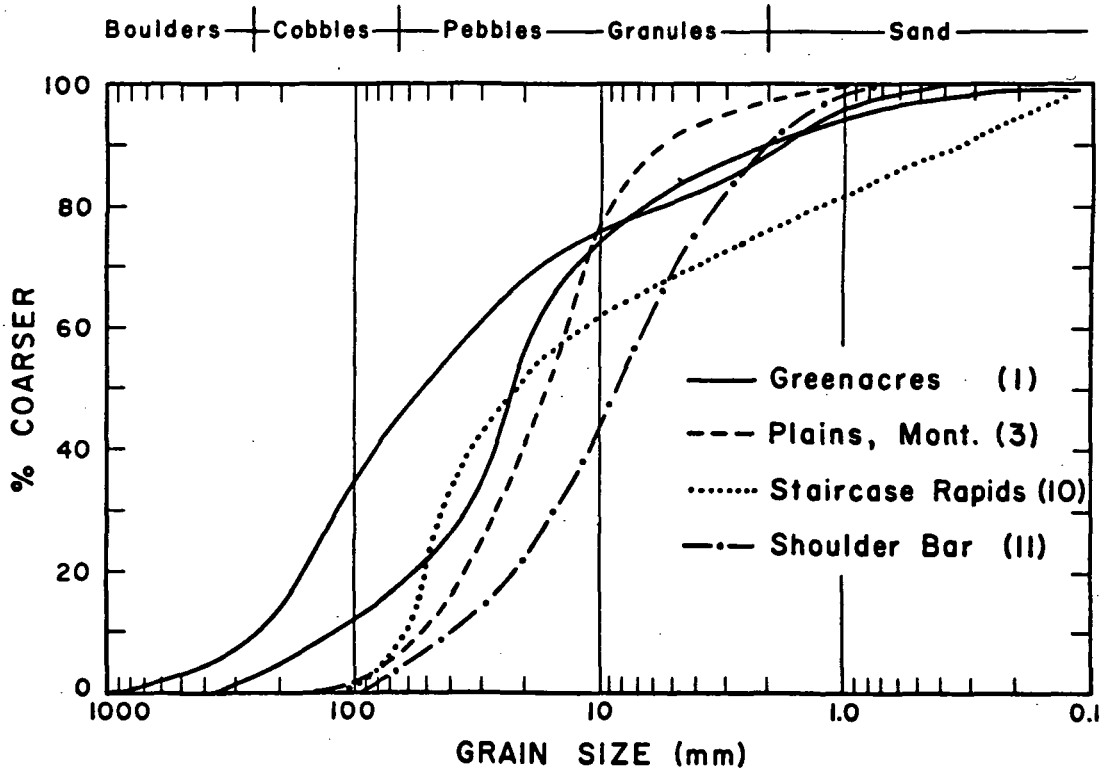


Figure 5.41. Grain-size distribution for sediment comprising giant current ripples in the Channeled Scabland. The numbers for the ripple sets are keyed to locations and descriptions in Baker (1973a, Appendix II).



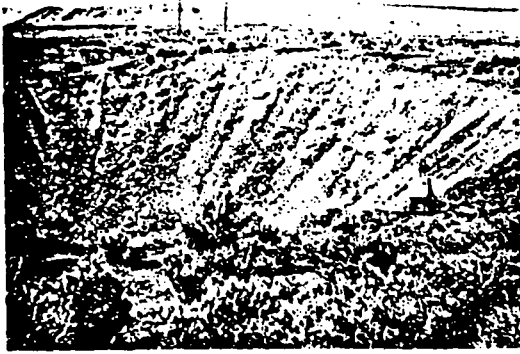


Figure 5.42. Foreset bedding exposed in a gravel pit cut through a Lind Coulee ripple oblique to the probable current direction. Apparent dips of the bedding vary from 10 degrees on the left to 25 degrees on the right portion of the photograph.

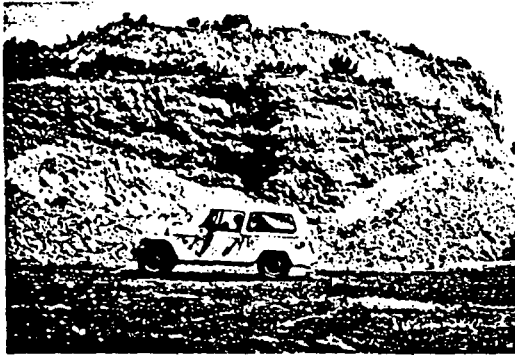


Figure 5.43. Trough-filling cross stratification in flood gravel exposed near Coulee City, Washington.

to predict mean ripple heights as a function of depth and water-surface slope. Similarly,

$$\bar{B} = 393.5 (DS)^{0.06} \quad (5-11)$$

may be used to predict mean ripple chords.

Because critical shear stress is related to the maximum grain size moved in traction through a given reach, ripple chord should also be related to the maximum grain size moved through the reach. Figure 5.48 shows that this is indeed the case. The correlation coefficient between ripple chord and maximum grain size, 0.993, is only slightly less than that between ripple chord and depth-slope.

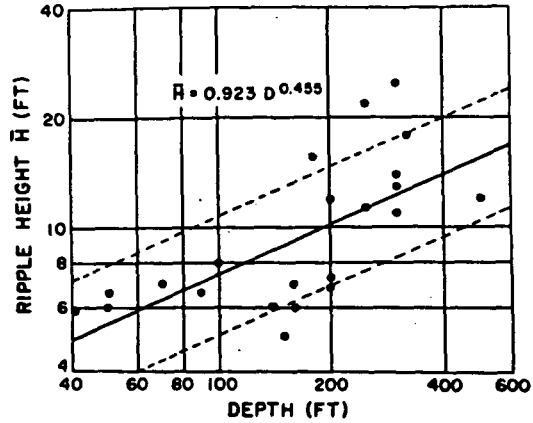


Figure 5.44. Logarithmic regression of ripple height as a function of depth. The dashed lines represent one standard error.

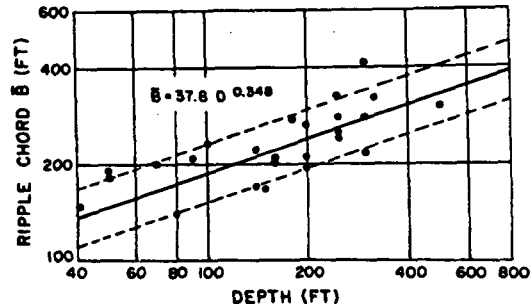


Figure 5.45. Logarithmic regression of ripple chord as a function of depth. The dashed lines represent one standard error.

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Mean Flow Velocity

Figure 5.49 shows the mean ripple chord plotted as a function of mean flow velocity. The correlation coefficient for this regression is 0.81, which may be interpreted as a 65% explanation of ripple chord variation in terms of mean velocity. This relatively low correlation may arise from the fact that the mean velocity in deep flows lies considerably above the mobile bed. The bed forms, like the boulder movement discussed earlier, actually respond to velocities close to the bed.

The mean velocity (\bar{V}) and the mean depth (\bar{D}) for a subsection containing current ripples may be used to estimate the Froude number (F)

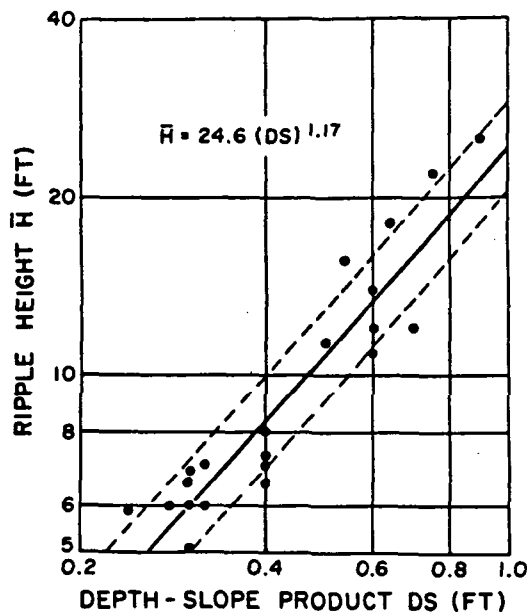


Figure 5.46. Logarithmic regression of mean ripple height as a function of the product of depth times slope. The dashed lines represent one standard error.

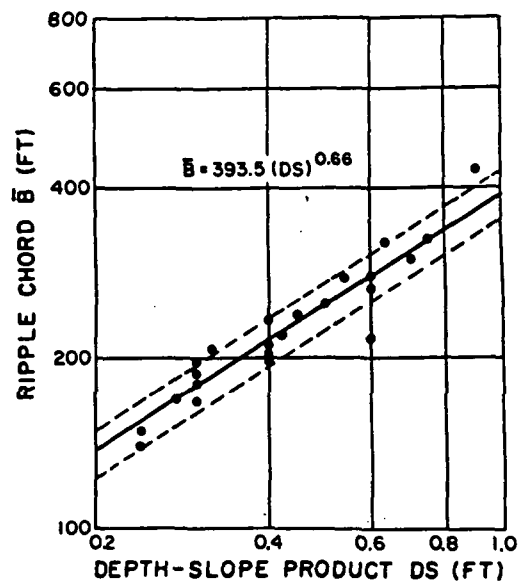


Figure 5.47. Logarithmic regression of mean ripple chord as a function of the product of depth times slope. The dashed lines represent one standard error.

through that subsection according to the formula

$$F = \frac{\bar{V}}{\sqrt{gD}}, \quad (5-12)$$

where g is the acceleration of gravity. Froude numbers for the scabland ripples generally vary from 0.5 to 0.9. The bed forms are formed within the tranquil-turbulent range that characterizes the lower flow regime of Simons and others (1965). The Froude number calculations reinforce the concept that scabland giant current ripples are the large-scale, coarse-grained analogs of bed forms known as "dunes" (Simons and others, 1965).

Stream Power. The product of mean flow velocity (\bar{V}) and bed shear (τ) gives a measure of stream power (Bagnold, 1966). Stream power is commonly used to predict certain bed forms

(Simons and others, 1965, their Fig. 21; Harms, 1969, his Fig. 9C). Stream power (ω) was calculated for the various scabland ripples from the expression:

$$\omega = \tau \bar{V} = \gamma (DS) \bar{V} \quad (5-13)$$

where \bar{V} is the mean flow velocity, DS is the depth-slope product, obtained from the high-water surface, and γ is the specific weight of the fluid. In Figure 5.50 ripple chord is plotted as a function of this measure of stream power. The correlation coefficient of 0.977 is the highest obtained in this analysis. The best predictive equation is:

$$\bar{B} = 8.65 \omega^{0.458} \quad (5-14)$$

in which the units are as shown in Figure 5.50.

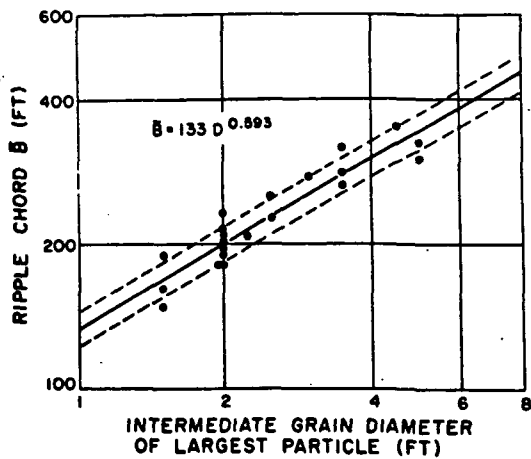


Figure 5.48. Mean ripple chord as a function of the maximum grain size found on the surface armor of the ripples. The dashed lines represent one standard error.

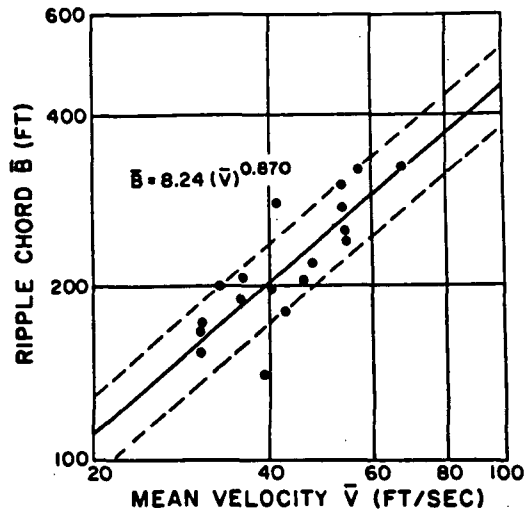


Figure 5.49. Ripple chord as a function of mean flow velocity (discharge velocity) as calculated by the slope-area method. The dashed lines represent one standard error.

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CONCLUSION

The bed forms of the Channeled Scabland show a remarkable consistency in scale and genesis to the hydraulics of the scabland flooding (see previous chapter). This consistency can be demonstrated in a quantitative fashion, at least as a first approximation. Nevertheless, one cannot help but look back on the remarkable insights of fifty years ago (Bretz, 1928c, p. 475-476):

"All scabland channels possess discontinuous mounds, hillocks or hills of stream gravel. . . . They are unlike any other detrital accumulations except the much smaller features of river channels commonly called bars. With these there is exact parallelism except for size. When considered in their setting in the scabland system, with all its other evidence for great volume and great erosion, they are seen to be an integral part. They should exist! And if they are bars, the great scoured channel ways should exist! Again this assemblage of unique land forms in the Pacific Northwest is seen to be a genetic group. A lively imagination is required for the acceptance of the hypothesis, but a scientific imagination withal."

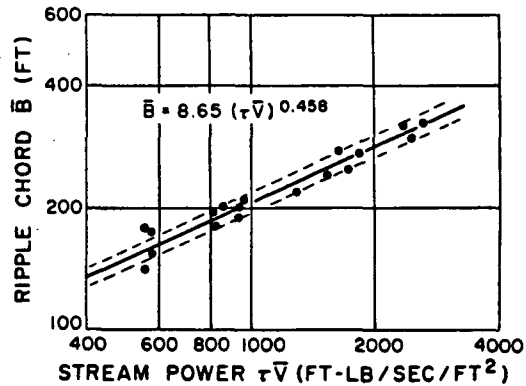


Figure 5.50. Mean ripple chord as a function of stream power. The dashed lines represent one standard error.