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Chapter 4

# Paleohydraulics and Hydrodynamics of Scabland Floods

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## ABSTRACT

The last major episode of scabland flooding (approx. 18,000-13,000 years B.P.) left considerable high-water mark evidence in the form of (1) eroded channel margins, (2) depositional features, (3) ice-rafted erratics, and (4) divide crossings. These can be used to reconstruct the maximum flood stages and water-surface gradients. Engineering hydraulic calculation procedures allow the estimation of flood discharges and mean velocities from these data. Despite a number of limitations on the accuracy of this reconstruction as discussed herein, the estimated paleohydraulic conditions are, nevertheless, consistent with a wide range of erosional and depositional phenomena in the Channeled Scabland.

Lithologic and structural irregularities in the Yakima Basalt were significant in localizing the plucking-type erosion of high velocity flood water. The anticlinal structures of the western Columbia Plateau resulted in numerous constricted channels. The phenomenal flow velocities achieved in these constrictions produced the most spectacular scabland features, including rock basins, potholes, and abandoned cataracts. In contrast, the eastern scabland region was characterized by more uniform flow conditions associated with more subdued erosional topography.

Secondary flow phenomena, including various forms of vortices and flow separations, are considered to have been the principal erosive processes. The intense pressure and velocity gradients of vortices along the irregular channel boundaries produced the plucking-type erosion. The great depth of the flood flows probably considerably reduced the effectiveness of cavitation as an erosional agent.

## INTRODUCTION

Geomorphic features result from forces acting on resistant materials at the interface between the lithosphere and the atmosphere or hydrosphere. Until the last decade the dynamics of the forces involved in making the Channeled Scabland were generally ignored in the controversy that surrounded the origin of those forces. Baker (1973a) was the first to use quantitative procedures in relating the pattern of scour and deposition to the regimen of scabland floods. Because the Missoula floods involved the largest discharges of fresh water that can be documented in the geologic record, continued study of scabland processes establish some upper limits to our knowledge of the short-term erosive and transport capabilities of running water.

Bretz recognized that eventually the Channeled Scabland problem needed to be investigated in physical terms. In questioning his own hypothesis (Bretz, 1932a, p. 82) he states, "somewhere must lurk an unrecognized weakness. Where is it? If it exists, it probably lies in the hydraulics of the concept." Bretz asserted that the turbulence of the glacial flood and the jointing of the basalt were both important, an idea gained from his earlier studies of the Columbia River (Bretz, 1924). He added (Bretz, 1932a, p. 83), "we do not know enough about great flood mechanics to make any conclusions valid . . . Hydraulic competency must be allowed the glacial streams, however much it may differ from that of stream floods under observation."

The present paper will review some of the hydraulic and hydrodynamic principles that apply to the Channeled Scabland. Moreover, these data will be related in the next chapter to the distinctive erosional and depositional features of the region. The results will support an earlier conclusion (Baker and Milton, 1974) that the distinctive bed forms in scabland channels are a hydrodynamic consequence of exceedingly swift, deep flood water acting on closely jointed bedrock.

## RECONSTRUCTING AN ANCIENT FLOOD

Several fortuitous circumstances have combined to allow an approximate reconstruction of maximum flood flows through the Channeled Scabland: (1) the last flooding episode was the last major event of the Pleistocene, occurring perhaps as recently as 13,000 years B.P. (but no older than about 18,000 years B.P.), (2) post-flood drainage on the Columbia Plateau was isolated by the paths of major rivers around the plateau, (3) the postglacial dry climate produced only intermittent streamflow, (4) flood deposits and erosive effects contrast sharply with the loess sediments and processes immediately adjacent to scabland channels, (5) exotic lithologies transported by the flooding are easily recognized, and (6) earlier flood events were followed by the massive loess deposition over the plateau.

### High-Water Marks

For the last major Missoula flood to cross the Channeled Scabland a variety of features have been preserved which serve as high-water marks. These features may be studied by geomorphic field work and by the interpretation of topographic maps and aerial photographs. The most abundant type of highwater mark is a divide crossing. Although these are best studied on vertical aerial photographs, they may also be recognized on detailed topographic maps (7.5-minute quadrangles).

*Eroded Channel Margins.* Eroded scarps on loess or the highest eroded rock surfaces (scabland) certainly mark a water level (Fig. 4.1). However, the various erosional features along

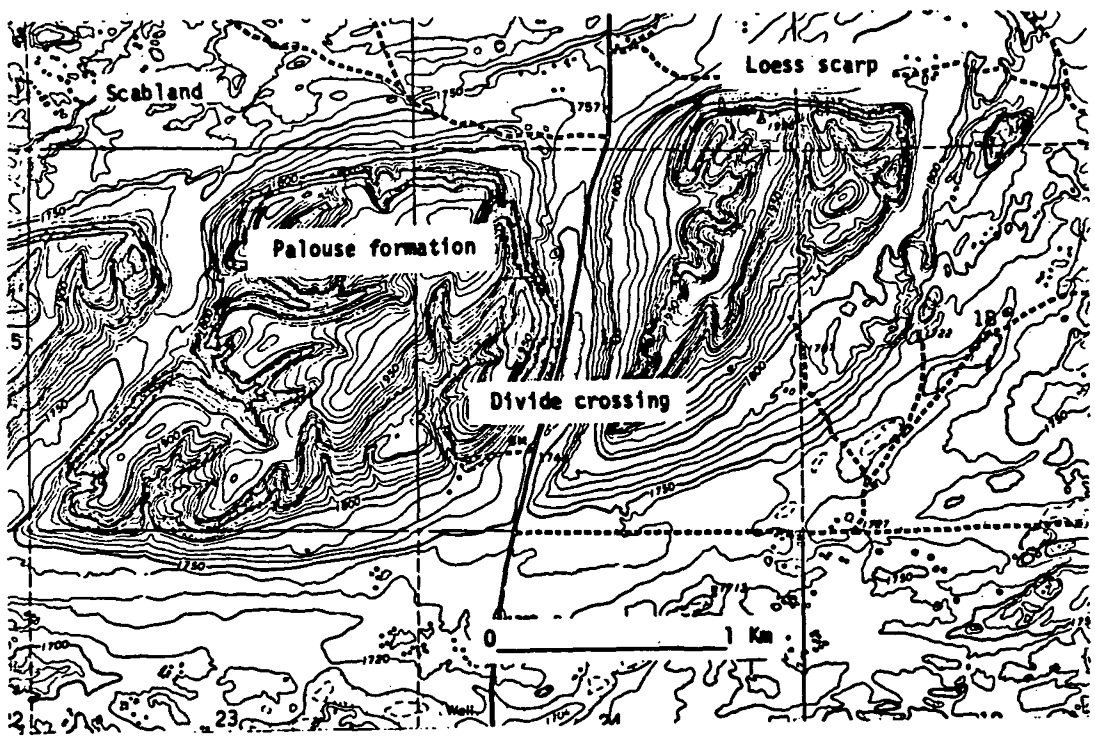


Figure 4.1. Loess scarps, scabland, and a major divide crossing in the Cheney-Palouse scabland tract. Topography is from the U.S. Geological Survey Texas Lake Quadrangle (7.5 minute, 10 feet contour interval).

scabland channel margins only provide minimum figures for high-water surface elevations.

*Depositional Features.* Bars of flood gravel also provide excellent evidence for flood flows, but, as with the large-scale erosive features, the flood gravel may have been under some unknown depth of water. These features must also be interpreted as minimum flow depth indicators. A more accurate procedure is to trace flood deposits laterally from scabland channels up the valleys of pre-flood tributaries. Such deposits rapidly fine away from the scabland channels and constitute "slackwater facies." Because tributaries to the Cheney-Palouse scabland drain the loess terrain of the Palouse hills, the slackwater basaltic sand can be easily recognized in these tributaries. The highest elevation of slackwater facies in the tributary valleys then provides a high-water indicator.

*Ice-rafted Erratics.* Exotic boulders of crystalline rocks from the Okanogan highlands or the Belt Series east of Spokane, are commonly found in mounds on the basalt or loess of the Columbia Plateau. The highest elevation of erratics that were transported by the floating ice in the Missoula Flood water is locally a useful high-water mark. The erratics are usually concentrated in areas where Missoula Flood waters were locally ponded.

*Divide Crossings.* Divide crossings, where flood water filled valleys and spilled over the loess-mantled interfluves, are the most common form of high-water mark evidence in the Channeled Scabland. In contrast to the dendritic drainage pattern on the loess divides, the flood-modified channels are linear or nearly linear (Fig. 4.2) and often expose bare basalt outcrops on their floors.

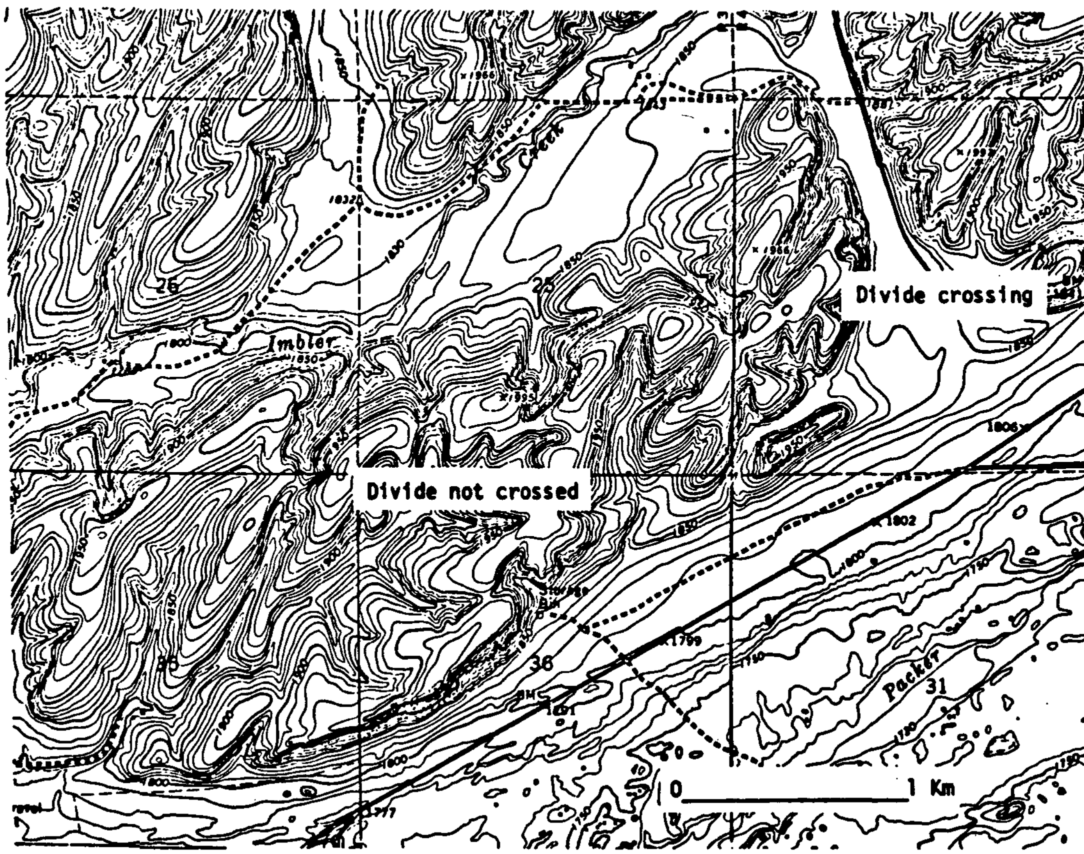


Figure 4.2. Examples of crossed and uncrossed loess divides in the Cheney-Palouse scabland tract. The divide crossing depicted is a definite example with a linear, flat-floored trough. Topography is from the U.S.G.S. Texas Lake Quadrangle.

Such divide crossings always establish lower limits for the high-water surface elevation and, in some cases, may fix upper limits as well. Obviously there is an expected range of error on these elevations. Elevation estimates are most accurate when divides were crossed by shallow water. These marginally-crossed divides are characterized by poorly-developed troughs. In areas where divides were crossed by deep flows, the approximation for the water-surface elevation is less precise. Thus, the most useful divide crossings are also the most difficult to recognize. Further refinement of the high-water surface elevation is possible through the study of aerial photographs and field observations of highest flood gravels, erosion of interfluves, and highest erratics.

The magnitude of the range of elevations obtained for the high-water surface depends on the contour interval of the topographic map and upon the nature of the crossing. Estimates are most accurate when uncrossed as well as crossed divides are present in the local study area (Fig. 4.3). A substantial portion of the topographic map coverage of the Channeled Scabland is at a 3 m (10 ft) contour interval. Thus the error range is rather small for maximum water depths of 100 m.

Uncrossed divides are generally relatively narrow and sharply defined. These divides are not dissected by flat-floored troughs and usually display no topographic irregularities with respect to surrounding terrain. The elevation of the lowest uncrossed divide may establish a maximum upper limit for the water surface. Divides that were apparently crossed are characterized by well-developed, flat-floored troughs through their crests. The incisions are commonly steep-walled and variable in length, but are usually elongated. Divides that were definitely crossed by deep flows commonly exhibit wide troughs and steep walls. These divides are found at lower elevations than those that are marginally crossed. They were probably crossed by at least 10 meters of water. Marginally crossed divides contain less well-developed troughs and are found at elevations higher than those crossed by deeper flows. These divides were in some places crossed by only a few meters of water (Fig. 4.4).

#### Water-surface Profiles

The plotting of the flood high-water surface begins with locating the obvious evidence, such as highest eroded scabland and major divide crossings. These data provide a rough approximation to the lower elevation limit. Refinement of the high-water surface elevation may follow through the location of the lowest divide not crossed (an upper limit) and the highest divide of marginal nature that was crossed (lower limit). It is important to select those divides just barely

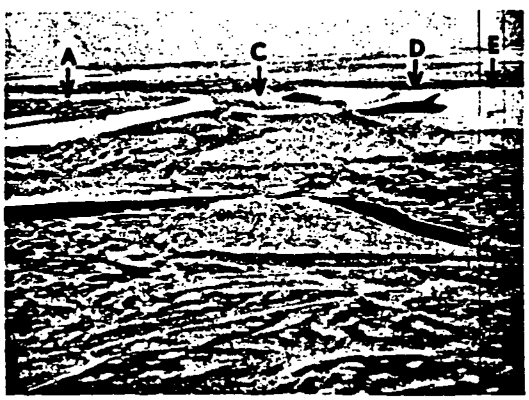


Figure 4.3. Oblique photograph of minor divide crossings (arrows) cut through a loess divide between Crab Creek and South Fork (Sections 26, 27, and 28, T.21N, R.36E.). The floors of the divide channels A, C, and D are at 600, 588, 594 meters (1970, 1930 and 1950 feet) elevation respectively. The uncrossed divide at E is at 606 meters (1990 feet).


Figure 4.4. Minor divide crossings located about 6.5 km north of Wilson Creek. The floors of the flood-eroded channels A and B are at 521 m (1710 ft) elevation. The nearby uncrossed divide at C is at 527 m (1730 ft) elevation. This type of relationship is of maximum utility in establishing the flood high-water surface.

crossed when establishing lower limits. There remains a nebulous zone between these two limits where it is uncertain whether divides have been crossed. The midpoint of this range can be taken to represent the high-water surface elevation for a local reach of the profile.

Figure 4.5 shows the high-water surface plot for the western part of the Channeled Scabland. The water surface slopes show a marked steepening at points where the flood waters encountered constricted channel sections. Water was ponded upstream from these constrictions and flowed at high velocity through the constrictions. Such constrictions were posed by the pre-flood drainage lines through the anticlinal structures of the western Columbia Plateau (High Hill, Frenchman Hills, and Saddle Mountains in Fig. 4.6). The steepening of the water surface gradient at the Saddle Mountains constriction (Othello Channels) is less pronounced than at the others. This is probably because the next constriction downstream, Wallula Gap, ponded water at 350 m (1150 ft) elevation.

## Flood Flow Calculations

The high-water surface slope can be used as an approximation of the energy slope in the slope-area method of indirect discharge calculation (see Benson and Dalrymple, 1968, and Baker, 1973a). Paleoflow depths are obtained by measuring the difference in elevation between the channel bottom and the high-water surface. Channel cross sections can be derived from the large-scale topographic maps. Baker (1973a) followed standard hydraulic engineering procedures

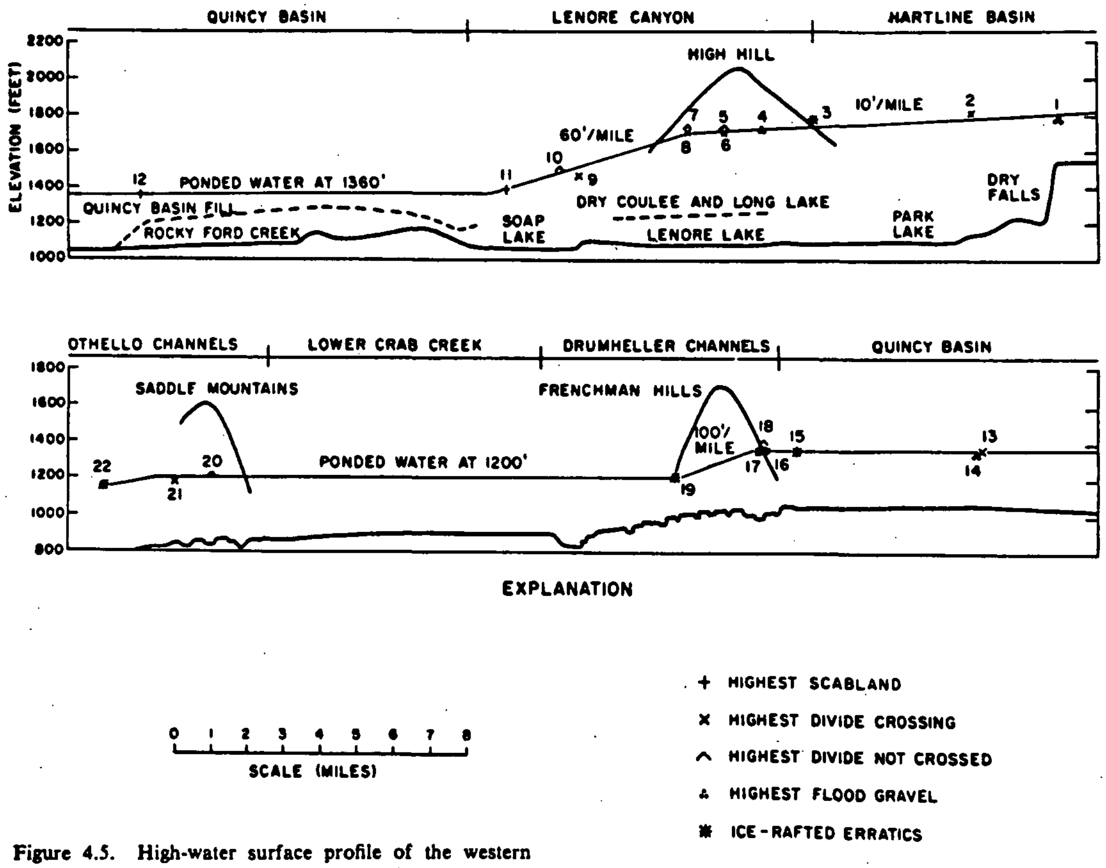


Figure 4.5. High-water surface profile of the western Channeled Scabland. The numbered high-water marks are individually described by Baker (1973a).

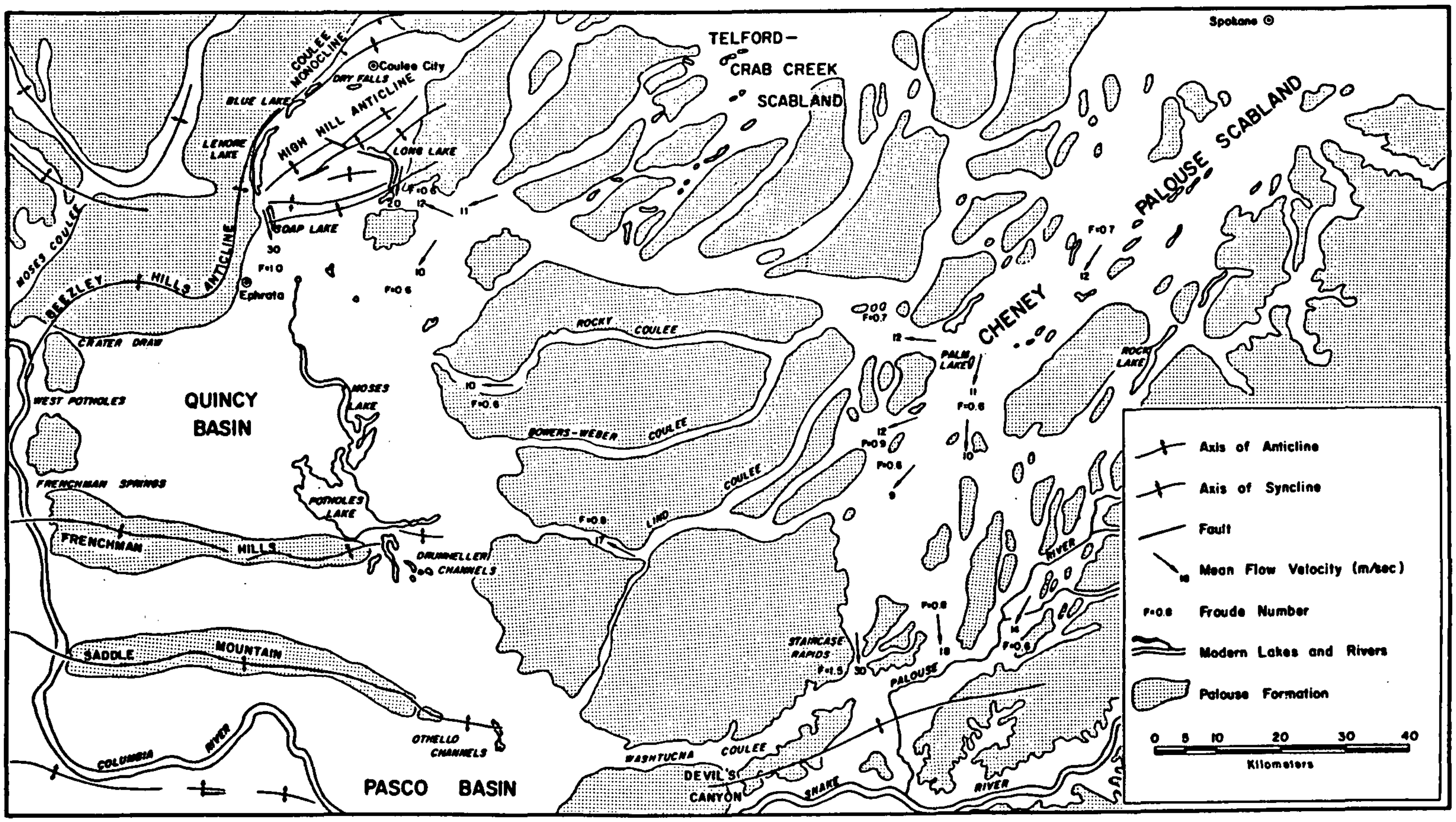


Figure 4.6. Regional paleohydraulic features of the Channeled Scabland. Mean flow velocities were determined from high-water mark evidence and channel geometry by Baker (1973a). Also shown is the regional structural pattern on the Columbia Plateau. The Palouse Formation is not shown on High Hill and Pinto Ridge (between Soap Lake and Long Lake) in order to emphasize the structural detail in that area.

to combine these data into estimates of probable flood discharges and mean flow velocities at the maximum flood stage. Discharges as great as 21.3 x $10^6$ $m^3$/sec were conveyed through the Channeled Scabland (Baker, 1971).

The distribution of mean flow velocities in the Channeled Scabland (Fig. 4.6) is important in understanding flood erosion processes. Constricted channels in the western scablands, such as Lenore Canyon, lower Grand Coulee, Othello and Drumheller Channels, achieved flood flow velocities as high as 30 m/sec. Such high velocities were reached only because of the unique combination of great flow depth (60-120 m) and very steep water surface gradients (2-12 m/km) that characterized the Missoula Flood. The constrictions contain the best developed erosional topography. Adjacent basins such as the Hartline, Quincy, and Pasco Basins, produced ponded water and the accumulation of sediment eroded from the constrictions (see Baker, 1973a, p. 15).

**Duration**

Baker (1973a) made some preliminary limiting calculations on the draining of Lake Missoula and the routing of flood flows through the Channeled Scabland. The peak discharge of flood flows in the Rathdrum Prairie, close to Lake Missoula's breakout point, was 2.13 x $10^6$ $m^3$/sec. Even at this phenomenal discharge, the lake volume of 2.0 x $10^{12}$ $m^3$ would have maintained the flow for about a day. Of course, the declining head in the lake would gradually reduce the discharge in time, spreading the water release over a broader time scale. It seems more likely that the flood peaks recorded by the high-water surface persisted no more than a few hours. Lower stages could have been maintained for a week or more.

The exact progress of the flood wave from the Rathdrum Prairie through the complex pathways of the Columbia River valley and the Channeled Scabland is not fully resolved. The initial flows filled the great valley of the Columbia upstream from the Okanogan Ice Lobe that blocked the river in the vicinity of modern Grand Coulee Dam. The water ponded to an elevation of about 770 m (2525 ft) and then spilled over the northern rim of the Columbia Plateau at the heads of the Cheney-Palouse, Telford-Crab Creek, and Grand Coulee scabland tracts. The subsequent flood events are less certain. The great cataract in the upper Grand Coulee, 250 m high, retreated northward at some time after this initial phase to the valley of the Columbia (Bretz, 1932a). The breakthrough of this cataract would then induce a great surge of water, 250 m deep, down the Grand Coulee and through the western scablands. Flood bars blocking the mouths of west-flowing scabland channels (Rocky, Bowers, Lind, and Washtucna Coulees, as well as Crab Creek) show that the Grand Coulee surge postdated flooding in the Telford-Crab Creek and Cheney-Palouse scabland tracts. The rapid drawdown of water by the upper Grand Coulee could have produced an abrupt cessation of flow in the eastern Channeled Scabland.

The stability of the Okanogan lobe during the flood event is also a problem. Waitt (1972a) has suggested that the last major scabland flood also destroyed the Okanogan lobe and released a great flood down the Columbia Valley. This flood produced a variety of catastrophic flood evidence that he has documented for the Wenatchee area, just northwest of the Channeled Scabland (Waitt, 1977b).

The ultimate control on outflow from the entire region of eastern Washington flooded by Lake Missoula was Wallula Gap, an antecedent canyon eroded by the Columbia River through the Horse Heaven Hills anticline (Bretz, 1925). Wallula Gap was inadequate to carry all the flood water supplied from upstream scabland channels. This constriction hydraulically dammed the flood, causing ponded water to reach 350 m elevation behind the "dam." Baker (1973a) calculated the maximum discharge achieved by this control point as 9.1 x $10^6$ $m^3$/sec. Thus, the total volume of Missoula Flood water entering from upstream would have required 2½ days to pass this point, even at the maximum discharge. More likely the high flows persisted for at least a week.

**The Question of Precision**

The discharge calculations of Baker (1973a) can only be considered a first approximation to flood hydraulics in the Channeled Scabland. The calculations were based on hydraulic formulae which were derived for use in streams whose discharges are 3 or 4 orders of magnitude less than those of the Missoula flooding through the Channeled Scabland. The data input into these formulae was not based on a dynamic record of the flooding, but rather on the time transgressive high-water surface. Time variant aspects of the flood surges cannot be quantitatively deduced from the existing field evidence. The complex geometry of Lake Missoula itself probably exerted an unknown dynamic influence on its draining. As discussed above, the flows were routed through an extremely complex set of anastomosing channel ways.

The exact significance of the high-water surface itself is open to some question. If the flooding burst on to the Columbia Plateau at near its peak discharge, then one might expect the high loess divides to have been eroded almost immediately. A subsequent long recession of the flood hydrograph might then have greatly deepened the channels, enlarging their capacity beyond what is implied by the high-water surface. Two lines of evidence argue against this view. First, the hydrographs of jökullhlaups (Icelandic: "glacier bursts") are precisely opposite to the above hypothesis. Flood flows rise slowly to a peak and then drop abruptly (Meier, 1964). The second line of evidence involves the field relationships. The extraordinary preservation of high-stage bed forms throughout the Channeled Scabland implies that the jökullhlaup hydrograph applies. Waning flood stages lasted so short a time that bed forms such as giant current ripples were not scoured away.

A more difficult problem arises in the estimation of roughness for the scabland channels. Baker (1973a, p. 19) followed Malde (1968) in assuming a value of Manning's "n" despite the many difficulties in scaling from the empirical Manning equation (derived for "normal" rivers) to the immense flow depths of the scabland flood. The value chosen from Chow's (1959) empirical tables was $n = 0.040$ for rock-bounded channels.

Komar (1978) has recently presented a lucid discussion of the problems of roughness estimation in very deep fluid flows, such as the Missoula floods and turbidity currents. Rather than employing the empirical Manning's "n" factor, he suggests use of a dimensionless drag coefficient $C_f$ that is related to other common resistance measures for flow in alluvial channels. By calculating the drag from the estimated flow depth and the particle diameter larger than 84% of the bed particles, one finds $C_f = 0.0026$ for the Missoula Flood. Baker (1973a) assumed a Manning's $n = 0.040$. Using Komar's equations this is equivalent to a $C_f = 0.0034$, in fair agreement with the estimate derived from the relative roughness measure. This correspondence lends added credence to the paleohydraulic reconstruction.

Despite the various limitations on the paleohydraulic reconstruction of flooding in the Channeled Scabland, a variety of sediment transport pheonmena have proven to be generally consistent with the quantitative reconstruction provided by Baker's (1973a) preliminary analysis. These include the bottom shear stresses for particle transport, the boulders carried by the flow, and especially the giant current ripples. These latter forms are almost certainly confined to the upper part of the lower flow regime of Simons and Richardson (1966). As expected, the calculated Froude numbers of the reaches containing those bed forms were in the range 0.5 to 0.9. There is no doubt that the future development of a better hydraulic theory for large floods may modify the absolute magnitudes of the events, but the relative pattern will probably remain unchanged.

## RESISTANCE

Except for occasional erratics of granite, loess, and Ringold Formation, 99% of the flood bedload sediments on the Columbia Plateau were derived from the Yakima Basalt. Thus, the resistance of this material partially dictated the activity of the phenomenal flow velocities described above. The Yakima Basalt consists of flow units which average about 30 m in thickness. The individual flows can be traced scores of kilometers. Bingham and Grolier (1966), Mackin (1961), Schmincke (1967a), Waters (1961), Wright and others (1973) have described the flow-by-flow stratigraphy in terms of such criteria as size and shape of columns, vesicle types, pillow zones, mineralogy, spiracles, and chemical composition. Local interbeds of tuffaceous sediments and buried soil zones are also distinctive. Swanson and Wright (Ch. 3, this volume) give a more complete discussion of this topic.

## Regional Structure of the Columbia Plateau

The Columbia Plateau is structurally a basin with the basalt surface dipping toward its center from surrounding uplifts. The Channeled Scabland begins on the northern rim of the basin, where the Columbia River has cut a deep gorge around the basalt plateau. From this northern rim the basalt surface has a general regional slope of 6-8 m/km southwestward from elevations of 850 m near Spokane to 120 m near Pasco. Bretz (1923a, 1928a) described the development of a consequent stream pattern along this dipslope prior to the late Pleistocene flooding. Deformation of the dipping plateau surface increases to the west (Fig. 4.6). A series of east-west ridges occurs in the western part of the plateau. These hils are structural anticlines that separate structural basins on the Plateau surface. The Frenchman Hills, High Hill-Pinto Ridge, Saddle Mountains, and Horse Heaven Hills all posed impediments to the passage of flood water across the Plateau. As shown by the high-water profile of the we··rn Channeled Scabland (Fig. 4.5), water was pond·d in the basins directly upstream from each of these anticlinal ridges.

The major structural basins of the western Channeled Scabland are the Hartline Basin, near Coulee City; the Quincy Basin, containing Moses Lake; lower Crab Creek, near Royal City and Othello; and the Pasco Basin, containing the Hanford nuclear works. Each of these basins filled with flood water until it spilled over the low points in the ridges that form the basin rims. The Hartline Basin contained water ponded upstream from High Hill and Pinto Ridge (Fig. 4.7). This water flowed through this southern divide at three points, Lenore Canyon, Dry Coulee, and Long Lake (Bretz, 1932a). Even greater ponding occurred in the Quincy Basin, covering over 1500 km$^2$. Water spilled from three divides (Fig. 4.8) in the upraised western rim of the basin, Babock Ridge and, through one divide (Drumheller Channels) on the southeastern margin of the basin. Bretz' pronouncement that all four outflow channels operated simultaneously shocked the geologic community of his day.


Figure 4.7. LANDSAT image of the Hartline Basin and the scabland channels leading from its southern margin: C-Lenore Canyon, D-Dry Coulee, and L-Long Lake. Other features are Dry Falls (F), High Hill (H), and Pinto Ridge (P). The scene measures 35 x 25 km. Compare to Fig. 4.6.

## Regional Structure and Stream Patterns

Studies that relate regional stream patterns to regional structure are a classical activity for geomorphologists. Since the field observations of J. W. Powell and the masterful deductions of W. M. Davis, there has evolved a concept of river valley adjustment to structure over long time scales—millions of years. A genetic terminology is applied to these relationships. *Consequent* streams follow the initial structural slope of the land. *Subsequent* streams become oriented along secondary features, such as faults and the regional strike. *Discordant* streams attract special interest because they cut across structural trends.

The Neogene outpourings of Yakima Basalt forced the Columbia River to assume a huge bight around the northern and western margin of the Columbia Plateau. In the west, the basalt was folded during the Pliocene to form the anticlinal ridges discussed above. The Columbia kept pace with this uplift, carving deep gorges through the Frenchman Hills, Saddle Mountains, and Horse Heaven Hills. Thus, its relationship to these anticlines is *antecedent*. Nevertheless, there was much local ponding of drainage by the active folds, as indicated by deposition of the Pliocene Ringold Formation. Ponding behind an anticline and subsequest overflow produces a discordant structure called *cross-axial consequent*.


Figure 4.8. Orbital (Skylab) photograph of the western margin of the Quincy Basin showing the divide crossings that drained water from the basin (Q) into the gorge of the Columbia River. These cataracts include Frenchman Springs (F) and Potholes (P). Note the giant current ripples on West Bar (W). This image measures 21 x 30 km.

The above relationships were well-known in the 1920's when Bretz began his studies. However, the discordant patterns he described in the Channeled Scabland completely defied the time scale that was generally applied to the structural control of rivers. The catastrophic flood outbursts filled basins and overtopped them in a matter of hours or days. Nevertheless, regional structure still exerted a dominant control on the channel (*not* valley) erosion by the flood flows.

The Lower Grand Coulee illustrates structural control of scabland flood erosion to a remarkable degree. Flood water spilling over a pre-flood divide to the north encountered the Coulee monoclinal flexure near present-day Dry Falls (Bretz, 1932a). This created a huge cataract, 250 m high that receded northward to form the Upper Grand Coulee. Below Dry Falls the flood water preferentially eroded the fractured rock on the steep eastward dipping limb of the monocline. This process excavated Lenore Canyon (Fig. 4.9) from the bent and broken basalt units.

Scabland divide crossings across anticlinal ridges also show the preferential erosion of fractured zones. Plucking erosion was concentrated at the tension-jointed anticlinal crests (Fig. 4.10). Joint control of rock basins and scabland channels is perhaps most spectacular on the Palouse-

Figure 4.9. Oblique aerial photograph of the head of Lenore Canyon near Park Lake. The dipping beds of the Coulee Monocline are exposed at the top, center. Downstream from this point, the dipping portion of the flexure has been completely eroded away to form Lenore Canyon. Longitudinal grooves mark the bare basalt surface in the foreground.

Figure 4.10. Erosion by scabland flooding at the crest of High Hill anticline in the southern Hartline Basin. Note the dip of Yakima Basalt units to the right and left of the eroded anticlinal crest (center).

Figure 4.11. Joint-controlled rock basins and channels on the Palouse-Snake divide crossing southeast of Washtucna, Washington. The Palouse River occupies the inner channel of the main canyon.

Snake divide crossing (Fig. 4.11). There it appears that the flood erosion simply etched out the regional joint pattern from the exposed basalt surface (Trimble, 1950).

### Small-scale Structural Features of the Yakima Basalt

Planes of weakness within the basalt bedrock were an important influence on fluvial erosional forms produced by Missoula flooding. The individual basalt flows average 25-60 m thick and are characterized by a variety of depositional and cooling features which allowed variable resistance to flood erosion (Fig. 4.12). Most flows have a scoriaceous upper portion grading into the rubbly top typical of aa lava. Some flows exhibit the classic Tomkeieff (1940) sequence of 2-tier columnar jointing. Columns in the lower colonnade may be 1-2 m in diameter (Fig. 4.13). An entablature of long slender columns and hackly fragments is present in all flows. In some flows an upper colonnade of much smaller columns occurs above the entablature. The cooling history of the flows controls the nature of columnar jointing (Spry, 1962). Irregular cooling surfaces developed because of erosion during the intervals between basalt emplacement. This often resulted in flaring columns in the next higher Yakima basalt flow. Rapid cooling produced hackly fragmented lava (Fig. 4.14A) and the "brickbat" zones described by Mackin (1961, p. 10).

Between the outpourings of basalt, enough time elapsed to permit weathering, growth of forest cover, and the formation of lakes. Local sedimentary intercalations in the basalts include conglomerate beds, clay layers, and freshwater diatomite. The lavas which overrode these lake beds formed pillow-palagonite complexes (Fig. 4.14B and C) and spiracles (gas chimneys) as described by Waters (1960).

Observations of basalt boulders transported by Pleistocene flooding (Baker, 1973a) revealed that the largest boulders were always portions of hackly jointed entablature (Fig. 4.15). Fragments of individual columns were common in scabland bars, but boulders greater than 1-2 m in diameter were not produced from lower colon-

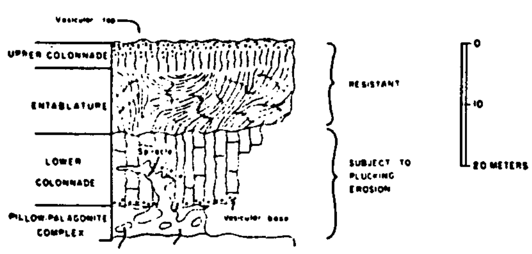


Figure 4.12. Cross section of an idealized Yakima basalt flow showing structural features important to flood erosion processes. The upper colonnade and the pillow-palagonite complex are only present in some basalt flows. Diagram is modified from Swanson (1967) and Schmincke (1967a) and uses the terminology of Tomkeieff (1940), Waters (1960) and Spry (1962).

Figure 4.13. Yakima Basalt (Rosa Member) showing massive columnar joints that were eroded by flood water flowing over Frenchman Springs cataract. The local basalt stratigraphy is discussed by Mackin (1961).

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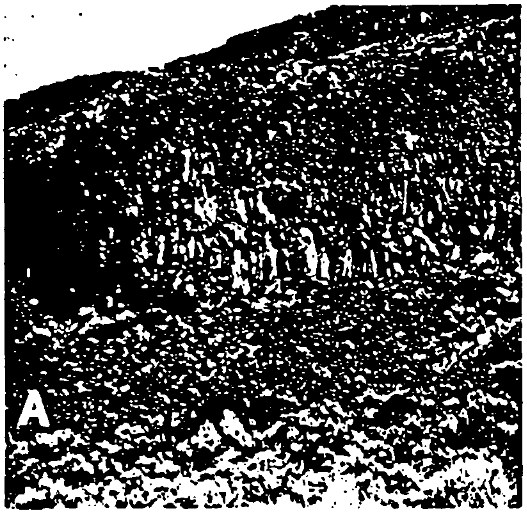

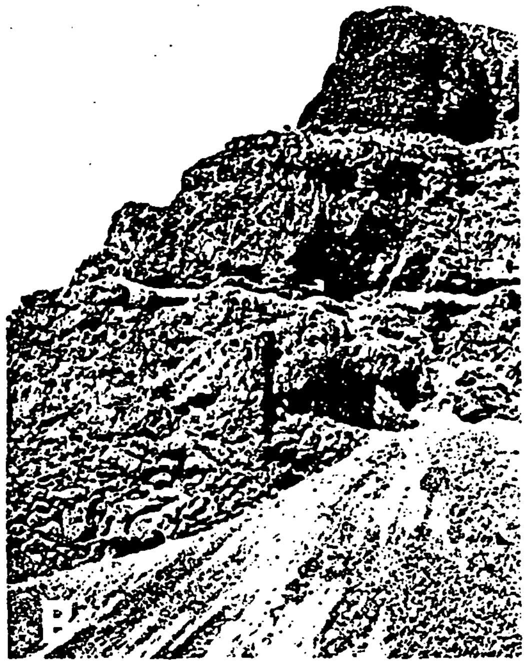

Figure 4.14. Structural characteristics of Yakima Basalt flow units. A. Entablature forming a resistant bench in flood-eroded scabland near Lenore Caves. B. Pillow lavas with overlying columns exposed in the wall of Moses Coulee near Appledale, Washington. C. Pillow-palagonite (Rosa Member) exposed in the Cheney-Palouse scabland tract. D. Entablature and colonnade exposure by flood erosion in Lenore Canyon.

nades because of the well-developed cooling-contraction joints in these zones.

## Ring Structures

Basalt flow surfaces in the northern Columbia Plateau sometimes display unusual circular structures outlined by dikes that surround a crater-like depression or central mound (Fig. 4.16). The structures range from 75-500 m in diameter and occur either as single rings (Fig. 4.17) or as multiple concentric dike segments. The relatively resistant dikes were preferentially preserved by flood erosion of the surrounding basalt.

McKee and Stradling (1970) attribute the structures to lava escape through tension joints that surrounded sags on a cooling lava crust (Fig. 4.18). However, new studies by Hodges (1976) revealed palagonite in the central mounds of many structures. The structures may result from ground water rising into the molten interior of a very thick basalt flow. Doming results from rapidly accumulating volatiles, with associated auto-intrusion of ring dikes and possible venting. The fragile palagonite cores were then easily eroded by the scabland flooding.

Figure 4.15. Boulder of basalt entablature measuring 18 x 11 x 8 m. This boulder is located on the Ephrata Fan 2.5 km west of the Rocky Ford Creek Fish Hatchery.

## HYDRODYNAMIC EROSION

### Cavitation

The enormous flow velocities achieved in many scabland channels caused considerable reduction

Figure 4.16. Oblique aerial photograph of concentric ring structures on a flood-eroded basalt surface north of the Karakul Hills, near Keystone Siding, Washington.

of absolute pressure, perhaps to the fluid vapor pressure. Shock waves produced by collapse of the air bubbles which form in such situations are recognized as intense erosional agents by civil engineers (Symposium, 1947). Embleton and King (1968, p. 271) suggested that this process

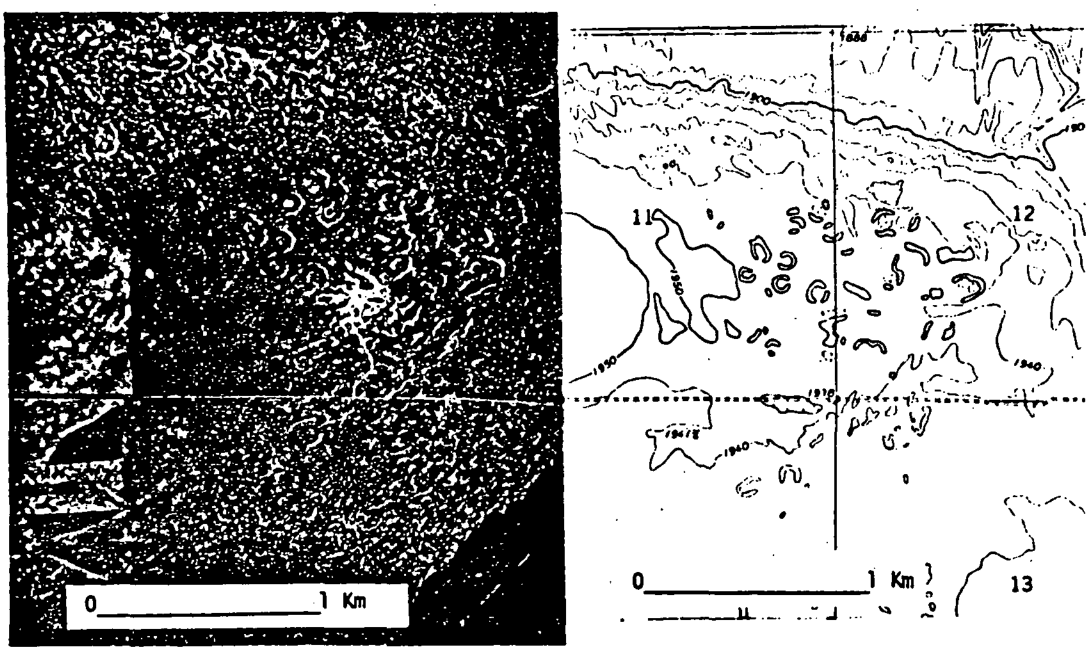


Figure 4.17. Vertical aerial photograph and topographic map (10 feet contour interval) of the ring structures shown in Fig. 4.16. These rings occur at 40 km east of the prominent rings described for the Odessa area by McKee and Stradling (1970) and by Hodges (1976).

of cavitation may have been responsible for the excavation of scabland rock basins. The bizarre "p-forms", potholes, and other erosional features which occur on rock surfaces in Scandinavia are interpreted as the result of cavitation erosion by subglacial melt-water streams flowing in concentrated routes or tunnels (Dahl, 1965). However, Schumm and Shepherd (1973) have suggested that such topography need not be subglacial.

Engineering experiments have shown that bubble collapse near materials such as steel or concrete produce hammer-like blows with local pressures as high as 30,000 atmospheres (Barnes, 1956, p. 494). These very intense local pressures are capable of shattering the surface of nearly any solid material. The fractured surface layers are quickly carried away by turbulent water and fracturing continues. Barnes (1956, p. 503) proposed that such processes in high velocity glacial streams may initiate the formation of potholes. He added, however, that cavitation forms are very short-lived. Once the potholes form, they would then trap abrasive particles of bedload and continue to enlarge by the rotary action of the water (Alexander, 1932).

I

II

III

IV

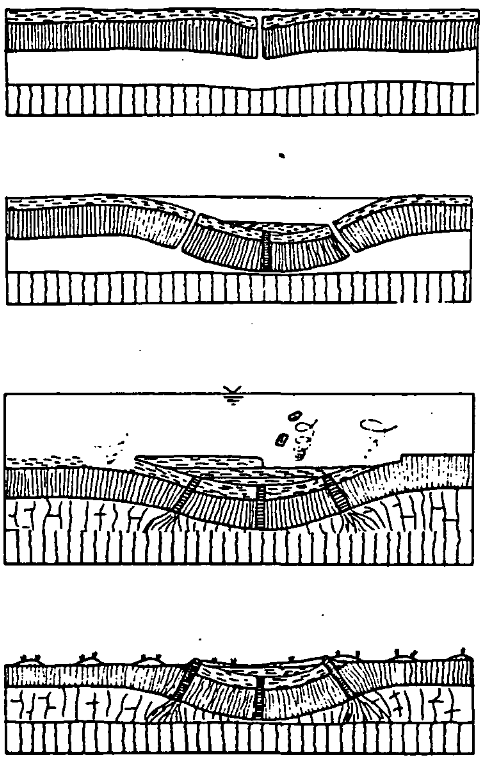

Figure 4.18. Hypothetic sequence of development for basaltic ring structures (modified from McKee and Stradling, 1970). I. Flowout of lava from the molten interior of a crusted lava flow. II. Sagging of the lava flow crust induced by loading from a point source induces circular fractures that allow additional lava escape. III. Plucking erosion by macroturbulent flood water. IV. Present relationships showing grass-covered silt mounds.

Cavitation can only occur for certain critical conditions. Hjulström (1935, p. 311) noted that the minimum velocity necessary for cavitation to occur in a river is about 12 m/sec. This figure is somewhat misleading, however, because it applies only to relatively shallow, swift streams.

The critical conditions necessary to initiate cavitation can be estimated from Bernoulli's equation (conservation of energy in fluid flow) written as follows:

$$\frac{V^2}{2g} + \frac{P_a}{\gamma} + d = \frac{V_c^2}{2g} + \frac{P_v}{\gamma} + Z, \quad (4\text{-}1)$$

where V is the mean stream velocity, $V_c$ is the velocity at the cavitation point, $P_a$ is the absolute pressure (atmospheric), $P_v$ is the fluid vapor pressure, d is the stream depth, Z is the assumed datum for position head ($Z = 0$ by this convention), $\gamma$ is the specific weight of water, and g is the acceleration of gravity.

Barnes (1956) assumed that obstructions on the streambed would generally increase the local fluid velocity to about twice the mean stream velocity. If $V_c = 2V$, then

$$V = \sqrt{\frac{2g}{3}} \sqrt{\frac{P_a - P_v}{\gamma} + d}. \quad (4\text{-}2)$$

For terrestrial conditions near sea level, this equation can be further reduced by assuming $g = 9.81 m/sec^2$, $P_a = 1$ atm., $P_v = 0.024$ atm. (21°C), and $\gamma = 9.8 \times 10^3 N/m^3$. The equation for the critical cavitation velocity $V_c$ (m/sec) is then a simple function of flow depth (m):

$$V_c = 2.6\sqrt{10 + d}. \quad (4\text{-}3)$$

Equation 4-3 indicates that at very high flow depths, the velocity must rise to very high values in order to achieve cavitation (Fig. 4.19). Only in a few narrow constrictions were the Missoula flood flows of sufficient velocity to achieve cavitation. All these areas, which are named in Fig. 4.19, show intense bedrock erosion. Nevertheless, it is likely that the erosion was localized by flow separations and other macroturbulent phenomena as discussed below.

## Macroturbulence and Plucking

A fundamental hydraulic characteristic of very deep, high gradient flood flows is the development of secondary circulation, flow separation, and the birth and decay of vorticity around obstacles and along irregular boundaries. Such three-dimensional flow phenomena in rivers are poorly understood even by hydraulic engineers. Indeed, Simons and Gessler (1971) have suggested that fluvial morphologic studies can make a valuable contribution to engineering river mechanics by describing examples of such phenomena in nature. Matthes (1947) has termed these phenomena collectively as "macroturbulence". The most important erosive form of macroturbulence is the "kolk", a Dutch term which Matthes used to designate intense energy disipation by upward vortex action. The intense pressure and velocity gradients of the vortex produce a phenomenal hydraulic lift force along the filament of the vortex (Fig. 4.20). The precise magnitude of this suction effect is unknown, but it certainly is greatly in excess of normal hydraulic lift forces.

The conditions necessary for the generation of kolks according to Matthes (1947) include the following:

(1) A steep energy gradient,

(2) A low ratio of actual sediment transported to potential sediment transport,

(3) An irregular, rough boundary capable of generating flow separation.

Jackson (1976) has recently presented an alternative model for kolks in which he attributes them to the oscillatory growth and breakup stages of the turbulent bursting phenomenon. Bursting, as described by Offen and Kline (1975), characterizes the turbulent structure of the outer part of the turbulent boundary layer. It is not yet clear, however, that this mechanism as observed

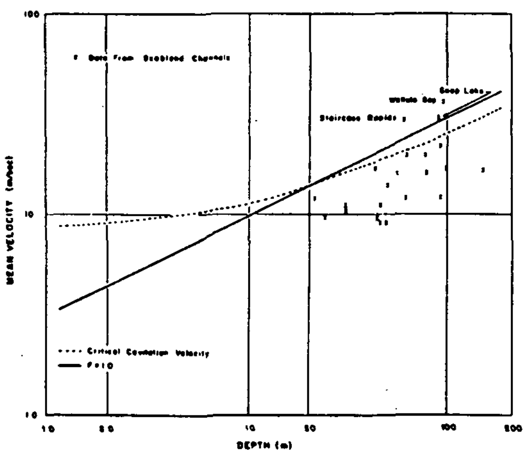

Figure 4.19. Influence of flow depth on the critical mean flow velocity necessary for cavitation to occur according to Barnes (1956, p. 499). Hydraulic data from scabland channels are plotted for comparison. Most of the flood flows were sub-critical and below the critical depth-velocity combinations necessary to produce cavitation erosion.

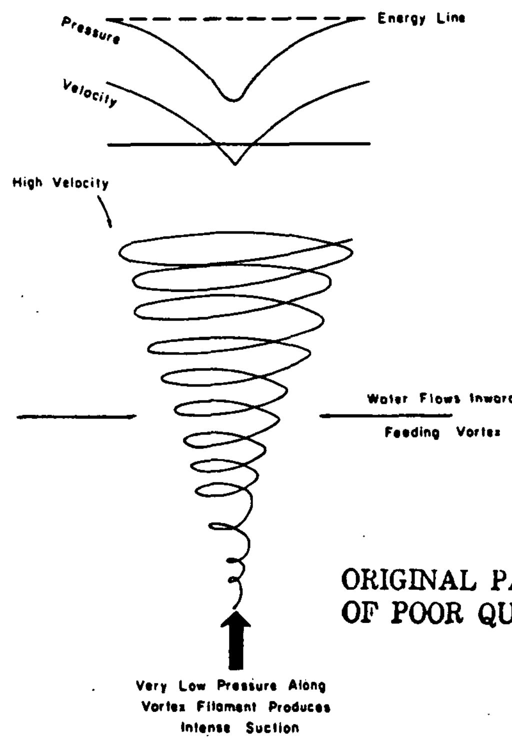


Figure 4.20. Characteristics of a hydraulic vortex ("kolk"). Such macroturbulent effects are set up by flow obstructions. The intense pressure and velocity gradients of the vortex produce a hydraulic lift. Alternatively, Jackson (1976) ascribes "kolks" to the turbulent bursting phenomenon.

in smooth-walled flumes will apply to the extremely irregular flow boundaries of scabland channels.

Turbulence as a general phenomenon is the end result of chaotic motion of various eddies that are superimposed on an average motion. When the size of the eddies corresponds to the absolute size of the flow, the scale of turbulence dictates the term "macroturbulence". The smallest eddies, "microturbulence", are almost independent of the size of the flow and tend to vary with the flow Reynolds number. The dimensionless period of macroturbulent flows, on the other hand, is generally independent of the Reynolds number. Whereas microturbulence is treated in stochastic terms, macroturbulence constitutes a more-or-less deterministic secondary flow that is superimposed on the prevailing mean flow components that result in the overall unidirectional movement of water.

The concept of macroturbulence is intuitively obvious to anyone observing fluid motions. Jackson (1976) cites Mark Twain's (Clemens, 1896, p. 44-48) description of distinctive patterns of turbulent fluid motions on the water surface of the Mississippi. Such concepts contributed to the original eddy concept of turbulence, a concept perhaps best summarized in the following rhyme, attributed to Richardson (1920):

> Big whorls have little whorls,
> Which feed on their velocity;
> Little whorls have smaller whorls
> And so on unto viscosity.

Thus eddies of a given size, or order, develop from larger eddies by borrowing energy from their "parents." This division process continues to such a small scale (microturbulence) that the eddies can no longer borrow sufficient energy to further divide.

Bretz (1924) first suggested that scabland-type erosion was a result of plucking rather than abrasion. He observed that eddies generated by the irregular channel walls at The Dalles area of the Columbia River are responsible for the plucking of basalt columns. He stated that inner channels of "The Dalles" type, potholes, and other scabland features could only be produced by a river of high discharge and steep gradient eroding close and vertically jointed rock. It was later suggested by Bretz and others (1956, p. 1028) that the hydraulic lift necessary for plucking action was provided by the kolk action of Matthes (1947). Bretz (1969) added, "plucking action thus was greatly augmented wherever submerged basalt ledges with the proper form appeared during flooded regimens."

The morphology of pothole erosion at Lenore Caves (Fig. 4.21) provides some insight into the details of the hydraulic plucking process. Like many scabland potholes, the one at Lenore Caves is asymmetrical in cross section parallel to the flood flow direction. Columns were preferentially plucked on the downstream side from beneath the resistant pothole rim of entablature. The result was a cave extending 20 m under the basalt (Fig. 4.21A). The symmetry of erosion is consistent with experimental flume studies of vortex scour. These studies indicate that at a flood peak, vertical vortices will scour at the downstream end of a hole on a stream bed and that the vortex axis will slant upstream (Fig. 4.21B).

The irregular rock steps of the Channeled Scabland created numerous points of flow separation. The lips of potholes (Fig. 4.21), canyon walls, and cataracts are but a few examples. Of course, separated flow is one of the most basic concerns of fluid mechanics. Leonardo da Vinci observed and sketched recirculating eddies. Surprisingly, however, the problem has not been fully explored in analytical terms. Nevertheless, an excellent empirical body of knowledge has evolved for the interaction of various geometries with fluid flow fields.

The dynamics of a turbulent separated flow at a downward step has been especially well studied. The temporal mean fluid pressure measured on the bed reaches a maximum at the point of reattachment and gradually decays downstream. Throughout the separated region and downstream there are large fluctuations of instantaneous pressure at any given observation point. The bed experiences rapidly varying normal forces that alternately pull and push at it. Turbulent shear stresses developed in and downstream from separated regions are much greater than in the boundary layer. Experiments by Allen (1971b) have confirmed that the greatest erosion by turbulent separated flows occurs at the points of flow reattachment. This conclusion applies to any of several erosional mechanisms, including cavitation, corrosion, fluid stressing, and solution.

Thus, secondary flow is easily induced around various kinds of obstacles or obstructions on a river bed. However, another important type of macroturbulence can be generated in straight channels free of obstruction. Prandtl (1926) showed empirically that secondary currents exist at the corners of rectangular closed conduits. Einstein and Li (1958) showed theoretically that transverse instability is often spontaneous in straight stream channels. The result of superposing transverse movements on the main flow in a stream is a helical array of vortices aligned parallel to the flow direction and showing alternating senses of rotation. These longitudinal vortices are distinctive in having their axes in the streamwise direction.

The theory initially yields two helical flows near the two banks of a relatively wide shallow channel. These two helical flows would then induce similar flows throughout the flow section until the entire fluid mass is split into secondary rotating cells (Karcz, 1967).

Karcz (1967) observed current-aligned ridges and troughs of sediment created by floods along ephemeral streams in southern Israel. These longitudinal structures have heights of 1-10 cm and spacings of 5-50 cm. They result from flows no greater than 3 m deep with velocities of .3 to

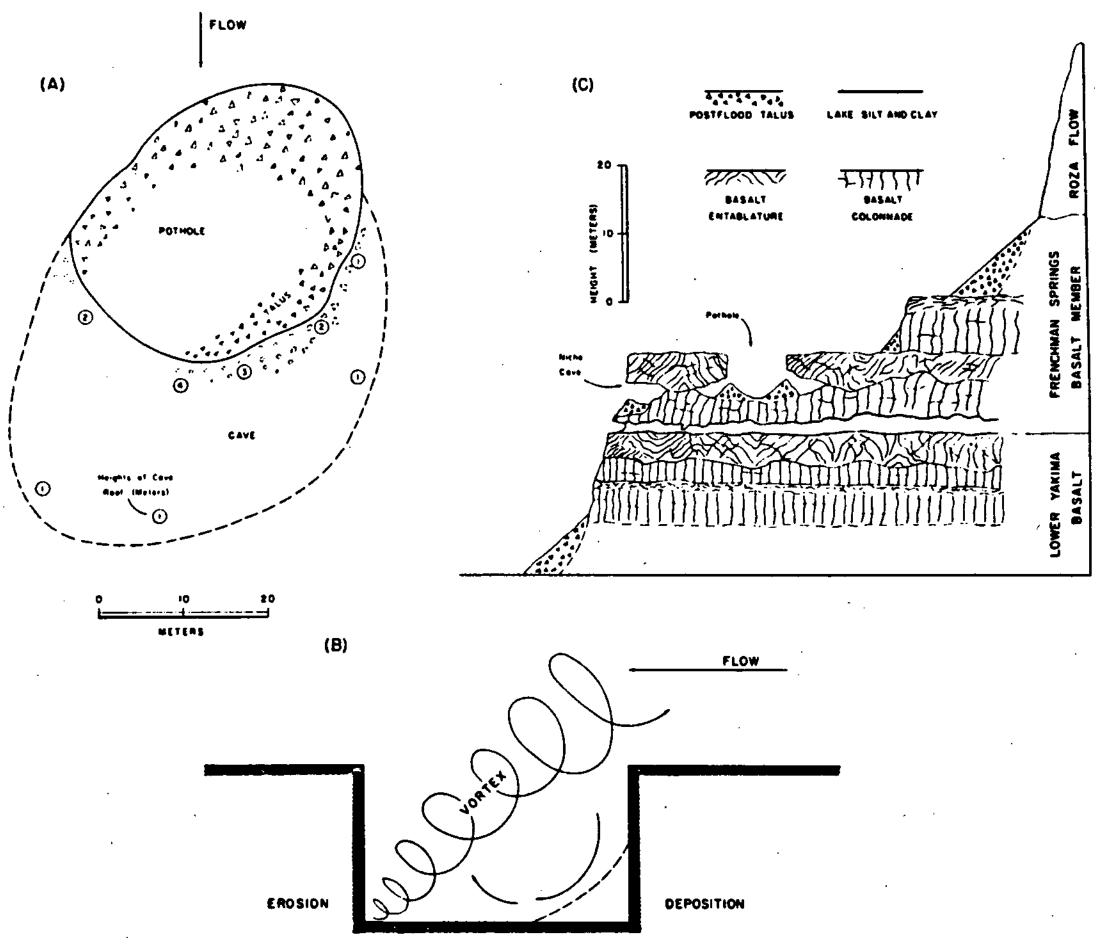


**Figure 4.21.** Pothole erosion at Lenore Caves, 15 km north of Soap Lake in Lenore Canyon. The sketch map (A) shows undercutting of the pothole rim on the downstream side. This is consistent with experimental observations of vortex scour (B). The relationship of pothole to the local basalt section is shown in (C).

3 m/sec. Karcz (1967) attributed these fluvial structures to secondary currents aligned in a series of longitudinal vortex tubes. Unfortunately, an adequate theory has not emerged for relating such large-scale vortex filament spacings to flow properties. For the small-scale longitudinal forms known as parting lineations in aqueous sedimentation, Allen (1970, p. 69) has shown that the responsible vortices increase in their spacing with both depth and mean flow velocity. The immense scale differences preclude the quantitative application of these results to the much larger groove phenomena in the Channeled Scabland.

The Einstein and Li (1958) analysis applied to fully turbulent flow in mathematically straight channels. Of course, any additional irregularity will create additional vorticity to interact with the longitudinal flow field. If the longitudinal vortices produce longitudinal bedforms, the bedforms will further enhance the flow field leading to stronger longitudinal erosion. This will eventually produce morphological forms that mimic certain hydraulic attributes of the responsible fluid. If defects occur in the bed, these may produce transverse flow separations that break up the longitudinal pattern of vorticity. Self-enhancement would then create transverse bed forms such as potholes and cataracts. Scabland erosion appears to derive from pronounced positive feedback.

In Chapter 5 it will be shown that a sequence of erosional forms appears to be developed on rock surfaces in the Channeled Scabland. The initial channels had relatively smooth floors. These are marked by longitudinal grooves, which mimic the longitudinal vorticity of the streaming fluid. Deeper erosion created irregular surfaces that generated flow separation and/or kolks. As erosive activity concentrated at these sites, the result was greater accentuation of the surface irregularities. A critical threshold had to be crossed to achieve this change from longitudinal forms to the production of irregular scabland.

The macroturbulent erosion mechanism has yet to be adequately evaluated in physical terms. Future research needs to focus on problems that are similar to those faced by Williams (1959) in his analysis of meteorite pitting: (1) the vortices should be able to form at the observed sites of intense erosion according to the principles of fluid physics, and (2) the presence of a vortex should set up a velocity distribution that will locally increase the erosion rate. Problem (1) seems to be satisfied by the development of flow separation at rock steps and by the development of longitudinal vortices as discussed above.

Problem (2) appears to have several ramifications. Williams (1959, p. 62) states the physical rationale for the velocity being greater in vortex tubes than in the general downstream flow field. In the idealized case of a vortex with constant circulation, the tangential velocity V at each point in the fluid is inversely proportional to the distance from the vortex center r:

$$V = \frac{A}{2\pi r}, \qquad (4\text{-}4)$$

where A is a constant of circulation. The gradient of V is inversely proportional to $r^2$:

$$\frac{dV}{dr} = -\frac{A}{2\pi r^2}. \qquad (4\text{-}5)$$

If a fluid is brought into a vortex with rotation A, it gradually moves down the axis of flow. While in the vortex the fluid will behave such that it develops less vorticity ($2\pi Vr$) when it leaves the flow than when it enters. By this picture, velocity from the general flow field tends to accumulate at the periphery of vortices until it is as great or greater than in the fluid entering the vortex.

## SEDIMENT TRANSPORT MECHANICS

### Incipient Boulder Motion

An early step in any sediment transport problem is to predict the conditions necessary to initiate the movement of bedload. One criterion for incipient motion expresses these critical conditions by the DuBoys equation for boundary shear, a function of flow depth and energy slope. Both depth and slope can be directly determined from the high-water surface evidence in the scablands, assuming that energy slope is approximated by the water surface slope. These parameters are then combined in the equation:

$$\tau = \gamma RS, \qquad (4\text{-}6)$$

where $\tau$ is the mean shear stress for initiating particle transport, $\gamma$ is the specific weight of the transporting fluid ($9.8 \times 10^3$ N/m$^3$ for clear water), R is the hydraulic radius (cross-sectional area/wetted perimeter), and S is the energy slope. In relatively wide channels (as typical for the Channeled Scabland) the depth of flow D is a close approximation to the hydraulic radius.

Actual particle entrainment is considered to be a stochastic process, dependent on instantaneous shear values exerted on the bed rather than on the mean values. Nevertheless, equation 4-6 gives a crude guide to the physical conditions associated with incipient particle movement.

Several geological studies of boulder movement have developed correlations between the grain sizes of the largest transported boulders and tractive force. Figure 4.22 compares the results of these studies to data from the Channeled Scabland. The analysis is described in detail by Baker and Ritter (1975). The data are approximated by the trend line:

$$D = 65\tau^{0.54}, \qquad (4\text{-}7)$$

where D is the intermediate particle diameter (mm) and $\tau$ is the shear stress (kg/m$^2$). Only the smaller scabland boulders from giant current ripples were used in this analysis to avoid macroturbulent effects on the boulder initiation and movement processes.

## Size-Distance Relationships

In rivers which have established an equilibrium between form and process over a long geologic history there are very regular downstream changes in hydraulic geometry (Leopold and Maddock, 1953), meander wavelength, and gra-

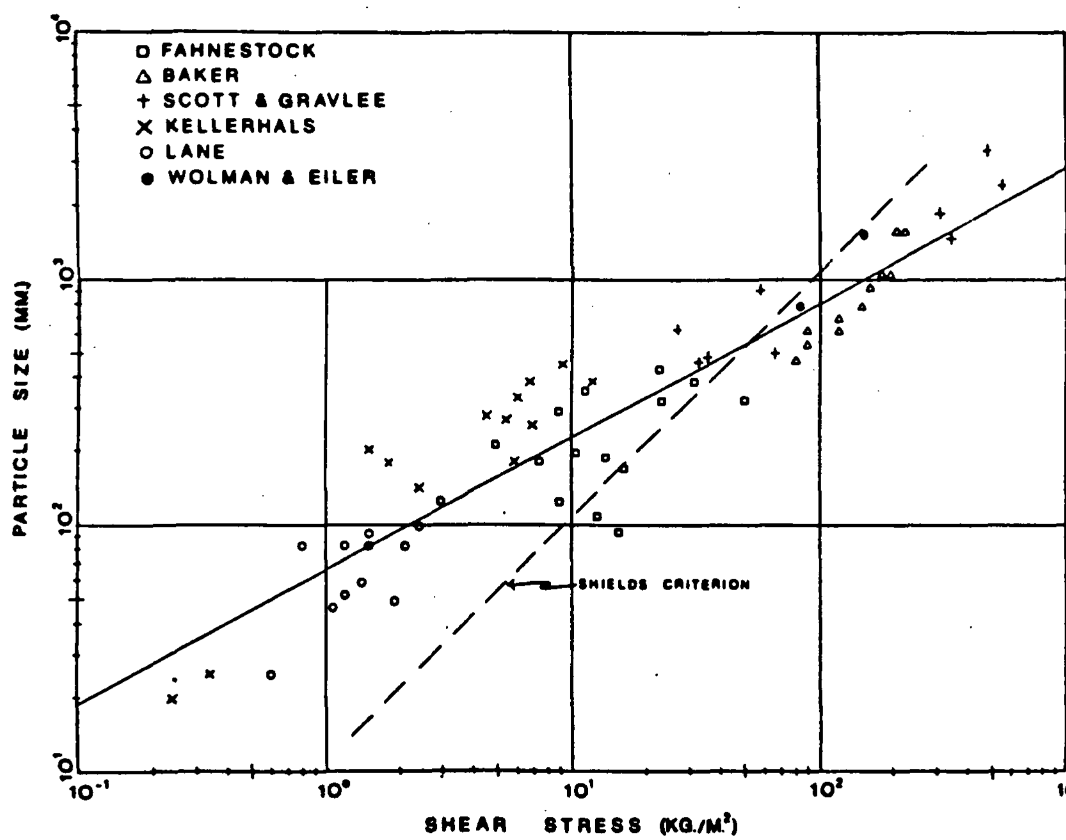


Figure 4.22. Intermediate particle diameter versus shear stress for extremely coarse bedload. The triangles indicate data for boulder sizes and shear in relatively uniform scabland reaches.

dient. Changes in the maximum sizes of transported sediment generally conform to Sternberg's Law:

$$D = ae^{-bx}, \qquad (4\text{-}8)$$

where D is the maximum particle diameter (mm) in a given reach, x is distance (km) downstream, and a and b are empirical constants. In the downstream reaches of rivers that are comparable in length to the Missoula Flood flow path (100's of kilometers), particle sizes are generally less than 50 mm. Coarser sizes may occur in steep, headwater tributaries, but these short stream segments are unsatisfactory as analogues to the large-scale Martian examples.

Alluvial fans typically have a-values (Equation 4-8) of 4 meters (Blissenbach, 1954), but size declines extremely rapidly down-fan. The b constant is often as great as 0.8. Similarly, outwash channels have coarse material in their proximal reaches, a being .3 to .4 meters (Bradley and others, 1972; Boothroyd and Ashley, 1975). The b constant indicates slower decline in size, typically .3 to .15. Even at these rates, 25 km of transport and selective sorting may reduce sizes to 10 mm in the distal portions of outwash trains. Most outwash plains have sandy (<2mm) distal facies (Boothroyd and Nummedal, 1978).

Baker (1973a) showed that debris eroded from catastrophic flood constrictions is distributed downstream according to the following equation:

$$D = ax^{-b}, \qquad (4\text{-}9)$$

in which variables are defined as in Equation 4-8. Measurements from a variety of catastrophic flood channels (Fig. 4.23) show that boulder size falls off very rapidly in the expanding reaches below constrictions. Studies in the Cheney-Palouse Scabland (Baker and Patton, 1976) indicate that very coarse debris (boulders) is localized in a narrow zone immediately below the constriction. Sizes fall off to granules and sand laterally toward channel margins. Material deposited in the lee of scabland streamlined forms (loess "islands") is 99% granules and finer sediment.

### Hydraulic Jumps

Figure 4.19 shows that the flow through several scabland constrictions (Soap Lake, Wallula Gap, and Staircase Rapids; see Fig. 4.6 for locations) was supercritical. When this flow passed through the subsequent downstream expansion, if the change in flow regime was abrupt enough, a hydraulic jump may have occurred at the mouth of the constriction. Komar (1971) has analyzed this problem in general terms for the passage of a turbidity current from a submarine canyon on to the submarine plain at the mouth of the canyon. The jump is a stationary surge or shock wave in which the speed of advance of the upstream wave is balanced by the velocity of flow downstream. Kinetic energy in the constricted, high-velocity flow is transformed to potential energy in the deep, low-velocity flow downstream from the jump. This transition involves an abrupt loss of mechanical energy through the generation of intense turbulence within the jump. Whereas the transition point may be a site of erosion or nondeposition, the immediate downstream area is a site of intense deposition as revealed in the laboratory experiments of Jopling and Richardson (1966). Komar (1971) suggests that only the

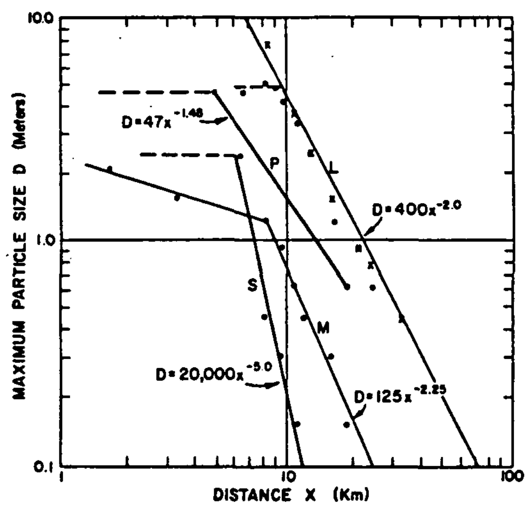


Figure 4.23. Downstream decrease in intermediate diameter of the largest boulders deposited below constrictions by catastrophic terrestrial floods. Data were obtained from the following sources: (L) Lenore Canyon, Channeled Scabland (Baker, 1973a), (P) Portland area below Columbia Gorge (Trimble, 1963), (M) Michaud Fan of the Bonneville Flood (Trimble and Carr, 1961), (S) Sentinel Gap, channeled scabland (Baker, 1973a). The constricted channels vary in width from 5 km to 16 km and water flowed between 120 m and 220 m deep at peak flood stage.

coarsest fraction will be deposited because the downstream flow will still be highly competent.

The Soap Lake constriction (Fig. 4.24) appears to show the morphologic and sedimentologic features that one would expect from a hydraulic jump. Soap Lake itself occupies a deep rock basin cut into the basalt bedrock (Bretz and others, 1956). Its location right at the mouth of the constriction is precisely at the point where one would expect the maximum flow velocities and the abrupt transition with the generation of intense turbulence. Downstream from Soap Lake is the immense gravel fan of the northern Quincy Basin. Baker (1973a) has described the unusual fall-off in maximum particle sizes that occurs on this fan. Although these factors point to the presence of a hydraulic jump at Soap Lake during the Missoula flooding, the precise physical description of that jump is hampered by the difficulty in reconstructing the flow geometry that prevailed at the precise time of the jump.

Figure 4.24. View north across the Ephrata Fan toward the Soap Lake constriction.

## CONCLUSION

Recent studies of catastrophic flood phenomena have provided the beginnings of a quantitative understanding for the hydrodynamics of exceedingly swift, deep flood water acting on bedrock. Although these processes may be of limited academic interest on the earth, they may have major significance for Mars (Baker and Milton, 1974). In future applications of these ideas, however, we must remember that our new understanding of catastrophic flood hydraulics and dynamics would not have been possible were it not for the thorough qualitative studies of earlier investigators. This relationship has been explicitly stated by Mackin (1963, p. 139), as follows:

> "In general, the larger the problem, the more many-sided it is, the more complicated by secondary and tertiary feedback couples, and the more difficult it is to obtain the evidence, the more essential it is to the efficient prosecution of the study that the system first be understood in *qualitative* terms; only this can make it possible to design the most significant experiments, or otherwise to direct the search for the critical data, on which to base an eventual understanding in quantitative terms."

The "critical data" for the Channeled Scabland were discovered only because of nearly one half century (1923-1969) of qualitative research by J Harlen Bretz.

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