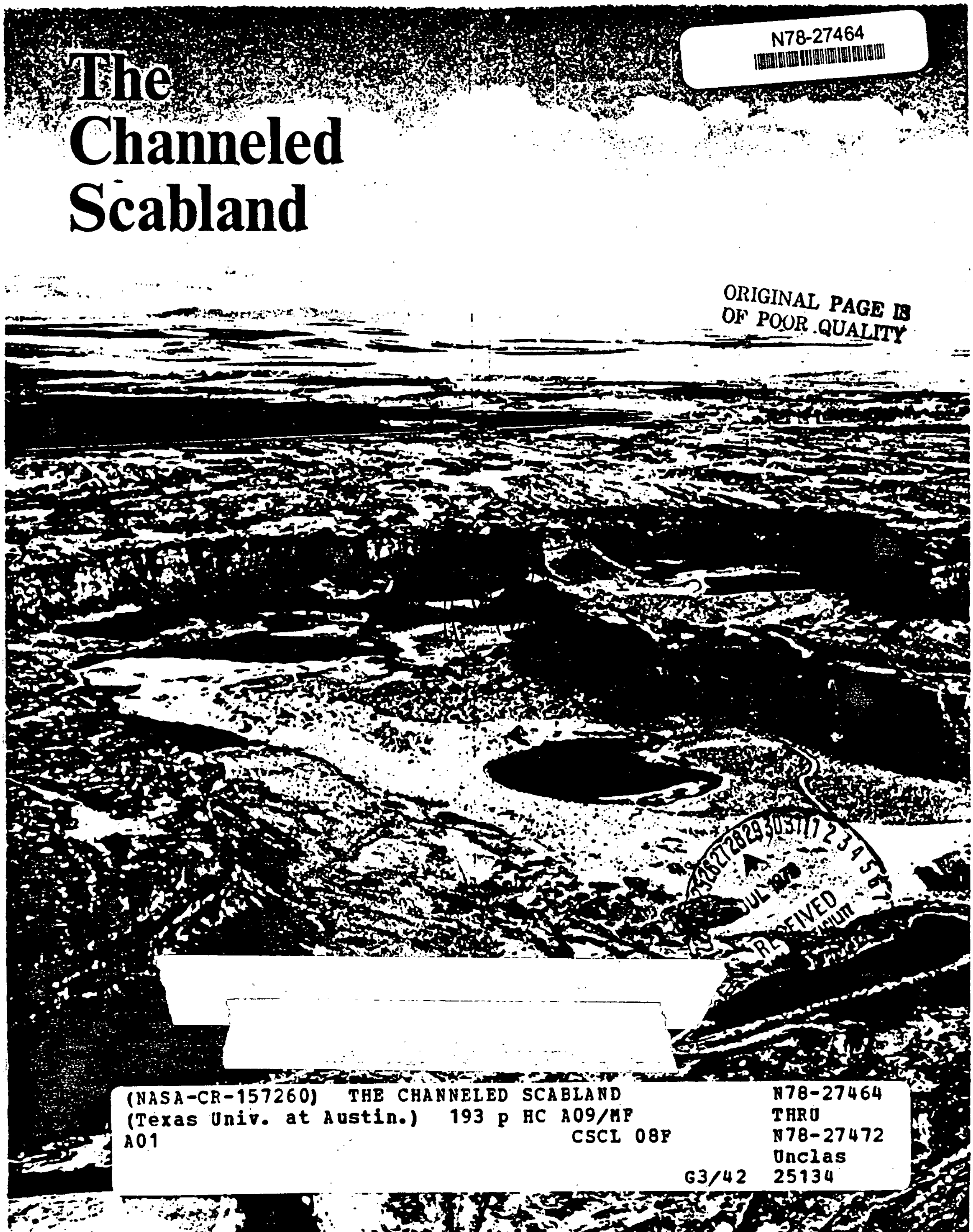
The Channeled Scabland



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The Channeled Scabland

A Guide to the Geomorphology of the Columbia Basin, Washington
Prepared for the Comparative Planetary Geology Field Conference held in the Columbia Basin
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edited by

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Preface

As the Mariner and Viking spacecrafts photographed the large channels on Mars, they generated a renewed interest in the erosional effects of large quantities of running water. As the debate continues whether these channels were caused by catastrophic floods, wind, lava or mudflows, it is imperative to examine, in the field, the morphologic criteria which appears to be diagnostic of catastrophic flooding. Hence this field conference to the Channeled Scabland.

Although we now propose to use the Channeled Scabland as a testing area for the Martian flood hypotheses, the flood origin of the scabland itself has a history of intense controversy. Ironically, many of the arguments against catastrophic floods on Mars are the same as those advanced against Bretz' "outrageous hypothesis" for the scabland origin. These include the source of such large quantities of water, the mechanism of its sudden release and the magnitude and nature of subfluvial erosion. Also, some of the alternative fluids proposed for channel generation on Mars were advanced in the debate about the origin of the Channeled Scabland.

THE CHANNELED SCABLAND intends to provide a thorough description of the morphology of the region and a discussion of its inferred cause, including the hydrodynamics of high-velocity flood erosion. It is designed as a guide, written specifically for the Comparative Planetology Field Conference in the Columbia Basin in June, 1978. The editors hope, however, that the book will provide a useful guide also for future field trips and an introduction to the area for geologists who intend to launch their own studies on specific attributes of this fascinating landscape.

The book is dedicated to J Harlen Bretz in recognition of his pioneering advocacy of the flood origin of the scablands. It is also an expression of our appreciation of the many other scientists who directly or indirectly have contributed to our understanding of the scabland origin. The ideas on the Channeled Scabland set forth in this book have matured through discussions with numerous colleagues, including Jon C. Boothroyd, William C. Bradley, J Harlen Bretz, Robert K. Fahnestock, Roald Fryxell, Harold E. Malde, George E. Neff, Peter C. Patton, Russell G. Shepherd, Richard B. Waitt, Jr. and Paul L. Weis.

The field studies and analytical work on the problem involved the assistance of Pauline M. Baker, Frances A. Heaton, R. Craig Kochel, and John M. MacGregor. Field work by V. Baker was initially supported by National Science Foundation Grant GA-21478. Subsequent studies were supported by NASA Grants NGR 44-012194 and NSG-7326, Planetology Program Office, and by The Geology Foundation, The University of Texas at Austin. Field studies by D. Nummedal have been supported by NASA Grant NSG-7076, Planetology Program Office, and the University of South Carolina.

P. Jeffrey Brown and Duane T. Eppler, the University of South Carolina, provided great help in guidebook editing, logistics planning and coordination of the field trip.

We extend our sincere appreciation to John S. Shelton for providing numerous illustrations for the book and to Henry D. Neumann for assistance in obtaining the photos for the color plates.

Neither the book nor the field trip would have been possible without the enthusiastic support by Stephen E. Dwornik, translated into financial assistance through grants to the participating principal investigators in NASA's planetary geology program.

We also thank Priscilla Ridgell and Carleen Sexton for the typing, Burk Scheper for the photo-technical work and Nanette Muzzy for the graphics.

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J HARLEN BRETZ

This book is dedicated to Dr. J Harlen Bretz, who, by painstaking field work, scientific insight and firm conviction, proved the catastrophic flood origin of the Channeled Scabland.

Biographical Sketch of J Harlen Bretz

J Harlen Bretz was born on September 2, 1882, in Saranac, Michigan. He attended Albion College, where he studied biology and met classmate Fanny Challis, whom he married in 1906. As a high school teacher in Flint, Michigan and later Seattle, Washington, he developed an avid interest in geology. His spare time was consumed with studies of glacial geology in the Puget Sound region. His hobby became his profession when the Puget Sound study became the basis for a Ph.D. summa cum laude at the University of Chicago. The entire dissertation was published (Bretz, 1913) and is distinguished for his revisions of the area's glacial history which had been described earlier by Bailey Willis and George Otis Smith. Bretz was also the first to recognize the significance of the great Osceola mudflow from Mount Rainier.

Bretz next spent a year on the faculty of the University of Washington. Unfortunately, his colleagues there did not share his zeal for field work, and Bretz accepted an invitation from R. D. Salisbury to return to the University of Chicago. He began teaching a field course at Devil's Lake, Baraboo, Wisconsin. His love for field work soon brought him back to Washington for summer field courses. Starting in the Columbia Gorge between Oregon and Washington, his field course soon moved into the Channeled Scabland as Bretz formulated his flood hypothesis (Baker, Ch. 1, this volume).

While a faculty member at the University of Chicago for over 30 years, 300 students completed their field geology training with "Doc" Bretz. The graduates include E. Dorf, M. K. Hubbert, W. C. Krumbein, J. L. Hough, A. N. Strahler, and H. S. Yoder, to name just a few.

He wrote 20 major articles and monographs on the Channeled Scabland, served as an associate editor of the Journal of Geology and was associated with the Washington, Illinois and Missouri Geological Surveys. In the 1930's, his research turned to physiographic studies in Greenland and the geology of the Chicago region. His two monographs on Chicago contain an ingenious analysis of the draining of Glacial Lake Chicago (a predecessor of modern Lake Michigan). From 1938 to 1961, Bretz established one of the most important American schools of thought on the origin of limestone caverns. His studies of caves in 17 states, Mexico, and Bermuda placed speleology on a firm scientific base. His insights and energy are a major reason for the modern resurgence of karst geomorphic and hydrologic studies in the United States.

Dr. Bretz officially retired from the University of Chicago in 1947. However, his years as Professor Emeritus were nearly as productive scientifically as those on the active faculty. With Leland Horberg, he published the first modern geological analysis of petrocalcic soil formation (caliche). Perhaps the most extensive survey of caves in one state was his book "Caves of Missouri" (1956). In the 1960's, Bretz' monumental analysis of geomorphic history in the Ozarks of Missouri had him questioning the new geomorphic paradigm of "dynamic equilibrium." He is a member of the American Association for the Advancement of Science and a fellow of the Geological Society of America.

Today, at age 95, Dr. Bretz still maintains a vigorous correspondence with former students and colleagues from his home at 2114 Cedar Road, Homewood, Illinois 60430.

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The publication in 1923 of a geomorphic study of "The Channeled Scabland of the Columbia Plateau" in southeastern Washington state launched a controversy that lasted for decades. A map accompanying the paper depicted a pattern of abandoned erosional waterways, many of them streamless canyons (coulees) with former cataract cliffs and plunge basins, potholes and deep rock basins, all eroded in the underlying basalt of the gently southwestward dipping slope of that part of the Columbia Plateau. The pattern of dry stream ways was described as at network, a plexus, an anastomosis; totally unlike any other drainage pattern on earth. A debacle was asked for, the volume of which filled existing normal stream valleys to overflowing. This great flood spilled over former divides, eroding their summits to complete the network. Associated with the enormously enlarged drainage ways in favorable places were similarly huge mounds of stream gravel which the writer called great river bars. Huge stream-rolled boulders occurred in these bars. The boulders were obviously plucked from the columnar basalt bedrock by the postulated high-velocity currents.

The term valley would no longer suffice. The abandoned rock-bound former waterways were called channels, and the entire composite was named "Channeled Scabland." The total area involved was 18 townships wide by 22 townships long (approximately 40,000 km^2).

The Cordilleran ice sheet had advanced to make contact with the heads of the dominant channel ways. Obviously the plexus of channels was in some way genetically related to the northern ice sheet. But where on our planet had glacial wastage produced a record of flooding and the total remaking of existing topography? The writer called for a catastrophic event.

Catastrophism had virtually vanished from geologic thinking when Hutton's concept of "the Present is the key to the Past" was accepted and Uniformitarianism was born. Was not this debacle that had been deduced from the Channeled Scabland simply a return, a retreat to catastrophism, to the dark ages of geology? It could not, it must not be tolerated.

This, the writer of the 1923 article learned when, in 1927, he was invited to lecture on his finding and thinkings before the Geological Society of Washington, D.C. an organization heavily manned by the staff of the United States Geological Survey. A discussion followed the lecture, and six elders spoke their prepared rebuttals. They demanded, in effect, a return to sanity and Uniformitarianism.

The upstart theorist was not upset nor silenced. Despite his knowledge that the country was full of other dissenters to his flood theory, he proceeded to publish more papers on his favorite topic, now named the Spokane Flood. He described other features of the afflicted plateau which he claimed were inexplicable without his flood of glacially derived meltwater. His apostacy would not be corrected as advised by the elders. The one-man rebellion was still alive.

1924 saw two new papers, 1925 saw one more and 1927, three more. One of the three contained the lecture and discussion already noted. Another paper in 1928 was the writer's reply to all alternative hypotheses thus far suggested. Also in 1928, he traced his flood down the Columbia as far as Portland, Oregon, adding a 200 square mile delta in the Willamette Valley.

By 1930, he had found a source for that immense discharge across the plateau. Clark Fork of the Columbia River, draining a large mountainous region of western Montana, had been dammed by the Cordilleran ice sheet at its traverse of the Idaho panhandle. This formed an immense glacial lake with an estimated volume of 500 cubic miles. The lake had been named some years before as Glacial Lake Missoula. The first geologist to describe the lake ironically was one of the six challenging elders in Washington in 1927 and the author of a short paper on problematical features, perhaps glacial in origin, in what came to be known as the Channeled Scabland.

If Lake Missoula had a properly located place for its ice dam and a clear route thence to the Channeled Scabland, then presto, we would have the big problem solved. Missoula's depth at the dam was known from its shorelines to have been 2000 feet, and there was a clear route to Spokane and the Scabland. A catastrophic failure of the dam would release 500 cubic miles of glacially derived water with adequate gradient to Spokane.

Late in the field study, a criterion of undeniable validity for the occurrence of a flood, or several such, came to light. Hidden largely by sagebrush were numerous occurrences of current ripple marks. They were discovered because the U.S. Bureau of Reclamation had taken aerial photographs of the area to be irrigated with Grand Coulee water. Then it became clear that some gravel surfaces, curiously humpy, were covered with giant current ripples. An investigator, standing between two humps, could not see over either one. Indeed, the size of these ripple ridges made them really small hills. Finally came the discovery of giant current ripples in parts of Lake Missoula where, in a catastrophic emptying, strong currents were formed. When investigated by a specialist in fluvial hydraulics and hydrodynamics there could be no question whatever of their origin, no other explanation for their rhythmic patterns than that of bedform development by amazingly deep, swift flood water.

Measurements of records for depths of water and gradients of water surfaces in channels with proper cross sections also yielded amazing figures for velocities. Plucking of channel bedrock to yield and transport huge boulders became understandable when this 1973 paleohydraulic and hydrodynamic study was made.

Back in 1927 geologic opinion still wavered. One man accepted flooding but did it with local iceberg dams in different channels. Another was a die-hard Uniformitarian who in 1938 still argued that no flood had occurred; that the erosional complex on the plateau was made by "leisurely streams no larger than the Snake in flood today." (But that was before the giant ripple marks were discovered.)

"Meticulously detailed" criticisms of the gross errors and assumptions in the 1938 non-flood interpretation appeared when three geologists spent much of the summer of 1952 (published in 1956) in studying chiefly the excavations made by the U.S. Bureau of Reclamation in diverting Columbia River water behind Grand Coulee dam. This water irrigates from a million acres of the Quincy Basin, an extensive fill made by scabland floods. There were repeated burstings of Lake Missoula dams and refillings when later advances of the ice front made other episodes in the lake's history. How many is not yet definitely determined, despite recent and continuing research on this problem.

When the Geological Society of America held its annual meeting in Seattle in November, 1977, and field trips were made to the flood-ravaged Columbia Plateau and valley, guidebooks made clear that the *catastrophic* history of plateau and master river valley is being pursued with enthusiasm.

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Chapter 1

The Spokane Flood Controversy

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ABSTRACT

The Spokane Flood controversy is both a story of ironies and a marvelous exposition of the scientific method. In a brilliant series of papers between 1923 and 1932, J Harlen Bretz shocked the geological community with his studies of an enormous plexus of proglacial channels eroded into the loess and basalt of the Columbia Plateau, eastern Washington. This region, which he named the "Channeled Scabland," contained erosional and depositional features that were unique among fluvial phenomena. With painstaking field work, before the advent of aerial photographs and modern topograhic maps, Bretz documented the field relationships of the region. He argued that the landforms could only be explained as the product of a relatively brief, but enormous flood, which he called the "Spokane Flood." Considering the nature and vehemence of the opposition to this outrageous hypothesis, the eventual triumph of that idea constitutes one of the most fascinating episodes in the history of modern geomorphology.

INTRODUCTION

The inimitable words of J Harlen Bretz (1928c, p. 446) describes the scene in eastern Washington:

> "No one with an eye for landforms can cross eastern Washington in daylight without encountering and being impressed by the "scabland". Like great scars marring the otherwise fair face of the plateau are these elongated tracts of bare, or nearly bare, black rock carved into mazes of buttes and canyons. Everybody on the plateau knows scabland. It interrupts the wheat lands, parceling them out into hill tracts less than 40 acres to more than 40 square miles in extent. One can neither reach them nor depart from them without crossing some part of the ramifying scabland. Aside from affording a scanty pasturage, scabland is almost without value. The popular name is an expressive metaphor. The scablands are wounds only partially healed—great wounds in the epidermis of soil with which Nature protects the underlying rock.
>
> With eyes only a few feet above the ground the observer today must travel back and forth repeatedly and must record his observations mentally, photographically, by sketch and by map before he can form anything approaching a complete picture. Yet long before the paper bearing these words has yellowed, the average observer, looking down from the air as he crosses the region, will see almost at a glance the picture here drawn by piecing together the ground-level observations of months of work. The region is unique: let the observer take the wings of the morning to the uttermost parts of the earth: he will nowhere find its likeness.
>
> Conceive of a roughly rectangular area of about 12,000 square miles, which has been tilted up along its northern side and eastern end to produce a regional slope approximately 20 feet to the mile. Consider this slope as the warped surface of a thick, resistant formation, over which lies a cover of unconsolidated materials a few feet to 250 feet thick. A slightly irregular dendritic drainage pattern in maturity has been developed in the weaker materials, but only the major stream ways have been eroded into the resistant underlying bed rock. Deep canyons bound the rectangle on the north, west, and south, the two master streams which occupy them converging and joining near the southwestern corner where the downwarping of the region is greatest.

Conceive now that this drainage system of the gently tilted region is entered by glacial waters along more than a hundred miles of its northern high border. The volume of the invading water much exceeds the capacity of the existing stream ways. The valleys entered become river channels, they brim over into neighboring ones, and minor divides within the system are crossed in hundreds of places. Many of these divides are trenched to the level of the preexisting valley floors, others have the weaker superjacent formations entirely swept off for many miles. All told, 2800 square miles of the region are scoured clean onto the basalt bedrock, and 900 square miles are buried in the debris deposited by these great rivers. The topographic features produced during this episode are wholly river-bottom forms or are compounded of river-bottom modifications of the invaded and over-swept drainage network of hills and valleys. Hundreds of cataract ledges, of basins and canyons eroded into bed rock, of isolated buttes of the bed rock, of gravel bars piled high above valley floors, and of island hills of the weaker overlying formations are left at the cessation of this episode. No fluviatile plains are formed, no lacustrine flats are deposited, almost no debris is brought into the region with the invading waters. Everywhere the record is of extraordinarily vigorous subfluvial action. The physiographic expression of the region is without parallel; it is unique, this channeled scabland of the Columbia Plateau."

A mere glance at a modern LANDSAT photograph of the Channeled Scabland (Fig. 1.1)

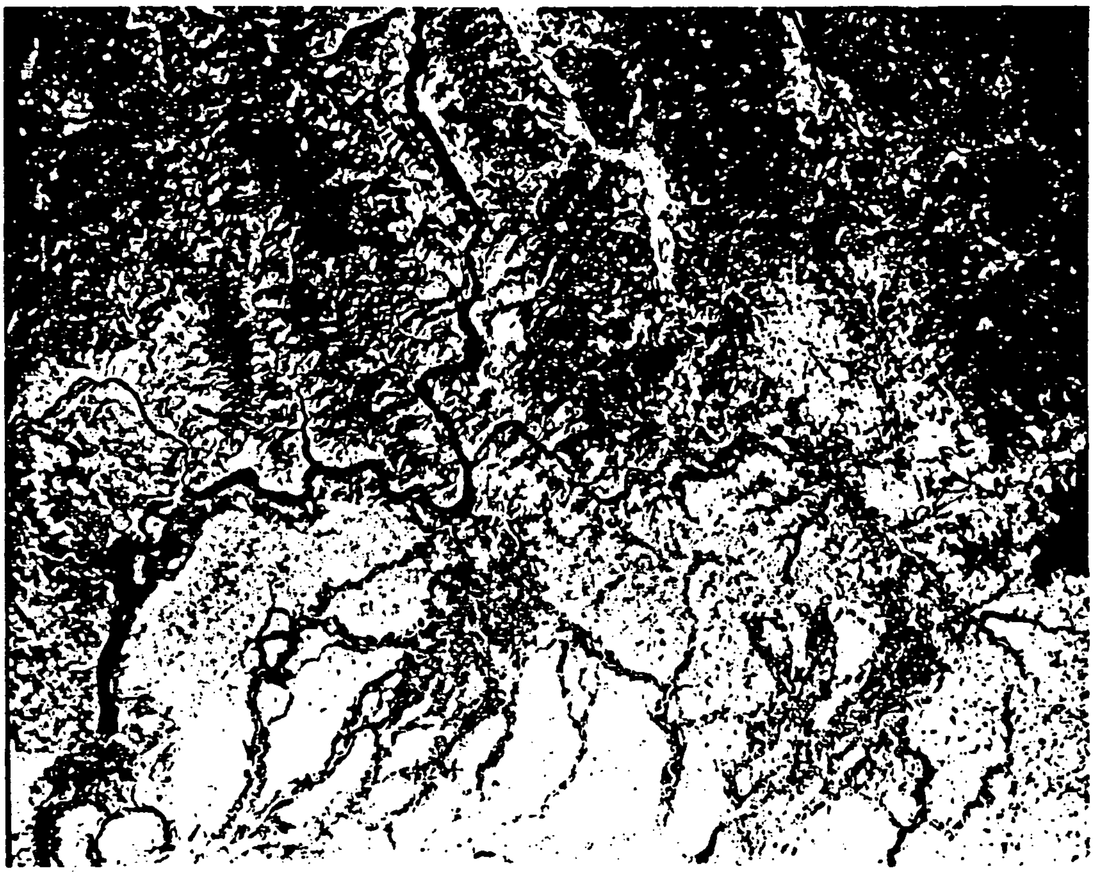

Figure 1.1. LANDSAT photograph of the northern part of the Channeled Scabland. Scabland channels form the dark-toned anastomosis that contrasts with the wheat farms on the light-toned Palouse loess. The Columbia and Spokane Rivers occur at the top (north) of the photograph. The far left scabland complex is the Grand Coulee-Hartline Basin-Lenore Canyon tract. At the center is the Telford-Crab Creek Scabland complex. At the right (east) is the Cheney-Palouse scabland tract (LANDSAT E-1003-18150 composite, 26 July 1972).

will show the features that Bretz, studying from the ground, developed as the basis of his flood hypothesis. The extensive wheat cultivation on the loess presents a vivid contrast to the flood-scared basalt exposed in the channel ways.

The unique character of the dry river courses ("coulees") of the Channeled Scabland was appreciated by the first scientific observers of the region. Rev. Samuel Parker (1838) provided the first published statement on the Grand Coulee: "[it] was indubitably the former channel of the river [Columbia]." Lieutenant T. W. Symons (1882) of the U.S. Army traversed the Grand Coulee, stating that he, "went north through the coulee, its perpendicular walls forming a vista like some grand old ruined roofless hall, down which we traveled hour after hour." Symons (1882) initiated the widely held notion that during glacial episodes of the Pleistocene the Columbia had simply been diverted across the Columbia Plateau. Variations on this general theme were standard in the early literature (Russell, 1893; Dawson, 1898; Salisbury, 1901; Calkins, 1905).

The Grand Coulee gained international fame in 1912 when it was traversed by the American Geographical Society's Transcontinental Excursion. Karl Oestreich (1915) of the University of Utrecht described the coulee as "eines mächtigen Flusses Bett . . . ohne jede Spur von Zerfall der frischen Form." He provided an excellent description of significant features that required a special origin: exhumed granite hills, perpendicular walls, and the hanging valleys marginal to the upper Coulee. He ascribed these hanging valleys to glacial erosion and to deepening of the coulee by the glacial Columbia River. Moreover, he recognized that the upper Grand Coulee was carved through a preglacial divide, which he correctly located just north of Coulee City.

Another foreign observer on the American Geographical Society excursion was H. Baulig, University of Rennes. Baulig (1913) described the loess, coulees, dry falls ("cataracte desséchée de la Columbia"), rock basins, and plunge pools. The origin of these features was ascribed to a glacial diversion of the Columbia. Nevertheless, he marveled at the scale of erosion (Baulig, 1913, p. 159): "peutêtre unique du relief terrestre,—unique par ses dimensions, sinon par son origine."

Dr. O. E. Meinzer, the eminent hydrologist of the U.S. Geological Survey, took an early interest in the western part of the Channeled Scabland. He observed (Meinzer, 1918) that the glacially diverted Columbia at Grand Coulee "cut precipitous gorges several hundred feet deep, developed three cataracts, at least one of which was higher than Niagara, . . . and performed an almost incredible amount of work in carrying boulders many miles and gouging out holes as much as two hundred feet deep." He implies that the great erosion occurred because the Columbia River was diverted across the steeply dipping basalt surface of the northern Columbia Plateau.

It was not until the studies of J Harlen Bretz (1923-1932) that the scientific study of this region began in earnest. Bretz interpreted the erosional and depositional features of the region as the product of a brief but enormous flood, which he called the Spokane Flood. For geology in the 1920's this was clearly an outrageous hypothesis. Olson (1969) has described the reception of the idea. "During its not always calm history, the story of the development of the Channeled Scabland was thought by some to have brushed beyond the dividing line in flaunting catastrophe too vividly in the face of the uniformity that had lent scientific dignity to interpretation of the history of the earth." The reaction of the scientific community was predictable, "this heresy must be gently but firmly stamped out" (Bretz and others, 1956, p. 961).

AN OUTRAGEOUS HYPOTHESIS

Because the Spokane Flood controversy is so tied to Bretz as its central figure, this review will consider part of his professional career during the years of his formulation of the flood hypothesis. The ensuing debates were not always marked by scientific objectivity, but their recounting is a fascinating example of the triumph of an outrageous hypothesis. Only in the last two decades has the flood hypothesis gained general acceptance. It is a measure of scientific maturity that in current studies of the Channeled Scabland, "the idea, but not the man has become central" (Olson, 1969).

While teaching at the University of Chicago, Bretz began conducting a summer field course in the wilds of the Columbia Gorge between Washington and Oregon. The idea for a study of the Channeled Scabland came during the summer of 1922. As he relates the story, "One summer I was out in Spokane. I saw a section of a topographic map of what is now called the Channeled Scabland, and from that I got the idea" (Quotation from Seattle Times, Sunday Magazine, July 11, 1971, p. 13).

Without the benefit of modern aerial photographs or even adequate topographic map coverage, Bretz began to take parties of advanced students into the region for month-long field studies. The work continued over the next 7 years. He soon revised an earlier notion that a marine submergence had occurred just downstream from the Channeled Scabland (Bretz, 1919). Nevertheless, the erratic granite boulders, which he had used as evidence for the submergence, were scattered about the basalt plateau far beyond the limits reached by Pleistocene glaciation. Bretz (1923a) named the glaciation responsible for these erratics the "Spokane Glaciation."

Although his first paper on the Channeled Scabland (actually the text of an oral presentation to the Geological Society of America) took care not to call upon cataclysmic origins, Bretz (1923a) provided a detailed description of physiographic relationships in the region. An example in his description (Bretz, 1923a, p. 601) of the pre-flood drainage line that was later enlarged to form the lower part of Moses Coulee: "The cliffs here are deeply notched by wide-open V-shaped tributary valleys. . . . These notches give the cliffs a striking resemblance to a series of great rounded gables in alignment. . . . Both widening and deepening in the basalt occurred and the tributaries were left hanging. They have since attained topograhpic adjustment by building large alluvial fans out on the canyon floor." He further noted that prodigious quantities of water were involved in the erosion. Referring to three outlets at the south end of the Hartline Basin (Dry Coulee, Lenore Canyon, and Long Lake Canyon), Bretz (1923a, p. 593-594) states, ". . . these are truly distributary canyons. They mark a distributive or braided course of the Spokane glacial flood over a basalt surface which possessed no adequate pre-Spokane valleys."

Bretz (1923a, p. 603) originally thought that the scabland gravels were organized into terrace remnants. However, after noting that they lacked a "sharp terrace form," this interpretation was quickly modified (Bretz, 1923b, p. 643): ". . . the evidence seems conclusive that all gravel deposits of the scablands are bars, built in favorable situations in the great streams which eroded the channels." With this conclusion he was forced to call upon catastrophic quantities of water. If the bars were over 100 feet in height, even greater water depths were required to form them. The second paper (Bretz, 1923b) also included the first detailed geomorphic map of the entire Channeled Scabland, showing the overall anastomosing pattern assumed by a great flood of water.

Bretz (1923b, p. 624-626) was the first to recognize the streamlined loess hills of the Cheney-Palouse scabland. He described them as follows: "A very striking and significant feature of the steepened slopes is their convergence at the northern ends of the groups to form great prows, pointing up the scabland's gradient. . . . The nose of a prow may extend as a sharp ridge from the scabland to the very summit of the hill. It is impossible to study these prow-pointed loessial hills, surrounded by the scarred and channeled basalt scablands, without seeing in them the result of a powerful eroding agent which attacked them about their bases and most effectively from the scabland's up-gradient direction."

Bretz knew that his interpretation would be controversial. He argued (Bretz, 1923b, p. 621), "All other hypotheses meet fatal objections. Yet the reader of the following more detailed descriptions, if now accepting the writer's interpretation, is likely to pause repeatedly and question that interpretation. The magnitude of the erosive changes wrought by these glacial streams is nothing short of amazing."

Bretz subsequently argued that the rugged scabland of anastomosing channels and rock basins cut into the basalt was the product of subfluvial quarrying. He described this process for the modern Columbia River near The Dalles, Oregon (Bretz, 1924). Moreover, he asserted that only large vigorous streams could produce such forms. The eventual conclusion from these varying lines of evidence was that so much glacial meltwater occupied the pre-existing valleys on the Columbia Plateau that it must have constituted a vast but short-lived flood, the "Spokane flood" (Bretz, 1925, p. 98). The flood spilled across pre-glacial stream divides, eroding the maturely dissected loess topography to form linear channels, and leaving a legacy of scoured loess scarps, hanging distributary valleys, and high-level fluvial deposits. It also built the huge constructional bars of gravel and then subsided so quickly that these bedforms were left almost unmodified by the subsiding water (Bretz, 1925, p. 105).

Bretz (1925) was able to trace the path of the great flood downstream through the Columbia Gorge to its debouchure into the Willamette lowland, where it built the "Portland delta." On this great subfluvial fan he recognized the significance of macroturbulence in accounting for certain flood features: "The Rocky Butte fosse is but the unfilled locus of an eddy caused by downward deflection where the current impinged on the east face of the butte. . . . The dependent terrace to the west was deposited in the slack water below the obstruction" (Bretz, 1925, p. 256).

Bretz (1925) even made the first estimate of the flood discharge. He chose Wallula Gap for this calculation because of the ponding effect of the constriction. His calculated maximum flow rate was 1.9 x 10^6 m^3/s (66.1 x 10^6 cfs), but he noted that this erred toward the low side. Nevertheless, he stated, "it represents the melting of about 42 cubic miles of ice daily" (Bretz, 1925, p. 258). He then notes that the insolation properties of ice and the total available ice mass north of the Channeled Scabland brings the whole concept into doubt. "The writer," he says (Bretz, 1925, p. 259), "has repeatedly been driven to this position of doubt, only to be forced by reconsideration of the field evidence to use again the conception of enormous volume. . . . These remarkable records of running water on the Columbia Plateau and in the valleys of the Snake and Columbia Rivers cannot be interpreted in terms of ordinary river action and ordinary valley development. . . . Enormous volume, existing for a very short time, alone will account for their existence."

Bretz (1925) then speculated on the somewhat obscure conditions that produced the Spokane Flood. He could only think of two possible explanations: (1) a very rapid and short-lived climatic amelioration, and (2) a gigantic glacier burst produced by volcanic activity beneath an ice cap. He noted severe objections to either hypothesis, but held that the great flood had occurred in spite of the problems in accounting for its source.

THE SPOKANE FLOOD DEBATE

In 1927 the Geological Society of Washington, D.C., invited Bretz to give a lecture "Channeled Scabland and the Spokane Flood." It was a purposeful invitation: a veritable phalanx of doubters had been assembled to debate the flood hypothesis. Bretz (1927a) presented the basic outline of his theory to date, citing the detailed field evidence which he could not explain by any hypothesis other than a great flood of water. The first discussant was W. C. Alden, who cautiously warned of the difficulties with the hypothesis. Lacking personal field experience in the region he suggested that the rock basins might be collapsed lava caves, but he realized that the major features indicated stream erosion. "It seems to me impossible that such part of the great ice fields as would have drained across the Columbia Plateau could, under any probable conditions, have yielded so much water as is called for in so short a time. . . . It appears that ice sheets of three distinct stages of glaciation invaded the borders of this region and may have afforded conditions of repeated floodings of much smaller volume" (Alden, 1927, p. 203).

O. E. Meinzer voiced a commonly held view of the Channeled Scabland, ". . . the Columbia River is a very large stream, especially in its flood stages, and it was doubtless still larger in the Pleistocene epoch. Its erosive work in the Grand Coulee . . . appears to me about what would be expected from a stream of such size when diverted from its valley and poured for a long time over a surface of considerable relief that was wholly unadjusted to it" (Meinzer, 1927, p. 207). He argued that the glacially swollen Columbia could have easily cut the Dry Falls and deposited the great gravel fan of the northern Quincy Basin. He described the Quincy Basin as containing an extensive series of terraces. Moreover, the high-level channels were explained by progressive abandonment as the glacial Columbia progressively cut down to lower levels.

One difficulty that Meinzer appreciated from his field work in the Quincy Basin (Schwennessen and Meinzer, 1918) was the fact that four great spillways led out of the region where water had obviously been ponded. Bretz (1923a) had shown that the upper limits of the torrents that poured through these spillways occurred at the same altitudes. Rather than ascribing this coincidence to contemporaneous operation, Meinzer actually published the idea that the spillways had been cut one at a time, and subsequent minor earth movements had later brought them to an equivalent altitude. "This recent deformation may account to some extent for channels cut through ridges that can not otherwise be well explained except by assuming excessive depths of flood water" (Meinzer, 1927, p. 208).

E. T. McKnight was also a participant in the Washington discussions. He subsequently suggested (McKnight, 1927) that a glacially diverted Columbia River was a viable alternative to Bretz' hypothesis. In response Bretz (1927b) argued that the great flood channels and bars near Gable Mountain (in the Pasco Basin) were far too large to be ascribed to the Columbia River. He made his position quite clear (Bretz, 1927b, p. 468): "I think I am as eager as anyone to find an explanation for the Channeled Scabland of the Columbia Plateau which will fit all the facts and will satisfy geologists. I have put forth the flood hypothesis only after much hesitation and only when accumulating data seemed to offer no alternative."

Bretz continued to answer various criticisms of his flood hypothesis (Bretz, 1928a, 1928b), and he established some new lines of inquiry into the problem. He (Bretz, 1929) showed that each of the valleys entering the eastern margin of the scabland spillways contained flood deposits emplaced by phenomenally deep water flowing *up* the tributaries away from the scabland channels. Along the Snake River he traced these deposits to beyond Lewiston, Idaho, more than 85 miles upstream from the nearest scabland channel. The conclusion again defied conventional wisdom (Bretz, 1929, p. 509): "Upvalley currents of great depth and great vigor are essential. . . . No descending gradient of the valley floor can be held responsible. The gradient must have existed in the *surface* of that flood. The writer, forced by the field evidence to this hypothesis, though warned times without number that he will not be believed, must call for an unparalleled rapidity in the rise of the scabland rivers." Each subsequent study produced yet another affirmation of the flood theory. Bretz (1930b) writes: "The writer, at least normally sensitive to adverse criticism, has no desire to invite attention simply by advocating extremely novel views. Back of the repeated assertion of the verity of the Spokane Flood lies a unique assemblage of erosional forms and glacial water deposits; an assemblage which can be resolved into a genetic scheme only if time be very short, volume very large, velocity very high, and erosion chiefly by plucking of the jointed basalt."

Among the spectators at the Washington lecture was J. T. Pardee. Pardee (1922) also had written on the origin of the Channeled Scabland. W. C. Alden, who was Chief of Pleistocene Geology, U.S. Geological Survey, had sent Pardee to study the scablands. He published a brief article (Pardee, 1922) proposing that the Cheney-Palouse scablands tract had been created by glaciation of rather unusual character. Bretz later visited Pardee's field locations and found that his "glacial" deposits were flood bars (Bretz, 1974). Correspondence between Alden, Bretz, and Pardee suggests that Pardee was really considering a hypothesis that the scablands might be related to drainage from a large Pleistocene lake that he had studied in the western part of Montana (Fig. 1.2) (Pardee, 1910). It appears that Alden dissuaded him from that idea (Bretz, 1974). In his memorandum of September 25, 1922, to

Figure 1.2. Late Pleistocene strandlines of Lake Missoula at Missoula, Montana. The highest strandlines reach 1280 m (4200 feet).

David White, Chief Geologist of the U.S.G.S., Alden notes of Pardee's work: ". . . very significant phenomena were discovered in the region southwest of Spokane. . . . The results so far . . . require caution in their interpretation. The conditions warn against premature publication." David White later asked Bretz if he knew what Alden's middle name was. When Bretz replied in the negative, White said, "It's Cautious, Bretz, Cautious."

It seems clear that the source of the great scabland floods was known even as Bretz was struggling to defend his hypothesis to doubters at the Washington meeting. One story has it that during the discussion Pardee leaned over to Kirk Bryan and said, "I know where Bretz' flood came from."

Bretz finally solved the source problem for the Spokane Flood in 1928. Although Harding (1929) without consultation or acknowledgement made the first announcement of Bretz' idea, Bretz (1930a) later published the discovery that scabland flooding resulted from an abrupt failure of the ice dam that retained Glacial Lake Missoula. Bretz (1932a) clearly illustrated the relationship of Lake Missoula to the Channeled Scabland.

James Gilluly was another of those at the Washington meeting who was upset with Bretz' hypothesis. Although he had not studied the Channeled Scabland in the field, he presented an imaginative and persuasive argument for the creation of the unusual landforms by the long-continued erosion of present-sized streams (Gilluly, 1927, p. 203-205). He took exception to a minor point concerning the use of talus heights as time indicators and then attacked the major weak point in the flood hypothesis. At that time the only two explanations offered for achieving the great volumes of flood water were (1) a very sudden climatic amelioration, and (2) subglacial volcanism and a resulting glacier burst. Some simple calculations demonstrate the inadequacy of either explanation in producing the required volumes of water in so short a time. He concluded, in essence, that Occam's razor did not apply to the Channeled Scabland and called for a more complex sequence of adjustments by rivers or floods not much larger than the Columbia. In reply Bretz (1927a) asked whether the lack of a documented source for the flood was proof that the flood had not occurred. He argued that the scabland phenomena themselves required the existence of a great flood.

Aaron Waters (*in* Bretz, 1972) relates that Gilluly was later to change his mind in this matter. Many years after the incident at the Washington Academy of Science Gilluly visited the Channeled Scabland on a field excursion. As he observed the Palouse-Snake divide crossing, a major scabland stream channel, his astonishment changed to a smiling comment, "How could *anyone* have been so wrong?" Nevertheless, the emotion of those days is evinced by the geologists who continued to deny the flood hypothesis and apparently never changed their minds on the matter: W. C. Alden, K. Bryan, W. H. Hobbs, F. Leverett, C. R. Mansfield, J. C. Merrian, O. E. Meinzer, and G. O. Smith.

The published record of the Spokane Flood debate is clear on one major point. Bretz repeatedly asked only that his flood hypothesis be considered not by emotion or intuition, but by the established principles of the scientific method. His detailed paper on the scabland bars contains the most eloquent expression of this plea (Bretz, 1928b, p. 701):

> "Ideas without precedent are generally looked on with disfavor and men are shocked if their conceptions of an orderly world are challenged. A hypothesis earnestly defended begets emotional reaction which may cloud the protagonist's view, but if such hypotheses outrage prevailing modes of thought the view of antagonists may also become fogged.
>
> On the other hand, geology is plagued with extravagant ideas which spring from faulty observation and misinterpretation. They are worse than "outrageous hypotheses," for they lead nowhere. The writer's Spokane Flood hypothesis may belong to the latter class, but it can not be placed there unless errors of observation and direct inference are demonstrated. The writer insists that until then it should not be judged by the principles applicable to valley formation, for the scabland phenomena are the product of river channel mechanics. If this is in error, inherent disharmonies should establish the fact, and without adequate acquaintance with the region, this is the logical field for critics."

THE REVISIONISTS

By the early 1930's the Channeled Scabland problem had become something of a sensation for American geology. Bretz (1932a, 1932b) had published the last of his field results, and he had embarked on new problems in Greenland and Alberta and ground-water studies in the U.S. His monumental but controversial field study was now open to the kind of attack that he himself had so strongly urged—new field studies.

Ira S. Allison (1933) was the first to enter the new foray. His view was not a denial of the Spokane Flood, but a modification. He argued that it was ice, rather than mere volume, that was the critical factor in the flood. He presented detailed evidence for the ponding of flood water all the way from the Columbia River gorge to the Wallula Gap. This ponding was produced ("in spite of the obvious difficulties of such an explanation") by a blockade of ice in the Columbia gorge. The blockade grew gradually headward until it extended into eastern Washington. As water was dammed to higher levels it spilled across secondary drainage divides creating the enigmatic hanging valleys, high-level gravels and widely distributed erratics. One of the key insights of Allison's motivation was in his last sentence, "perhap this revision will make the idea of such a flood more generally acceptable" (Allison, 1933, p. 722).

Hodge (1934) published a brief interpretation of the Channeled Scabland involving mainly glacial processes. He hypothesized a complicated alternation of ice advances and drainage changes. The basalt was quarried by glacial erosion, and channel complexes in the basalt were produced by the diversion of meltwater streams around blocks of stagnant glacier ice and jams of berg ice. The theory was never adequately supported by published field evidence.

Perhaps the most serious alternative to the Spokane Flood hypothesis was posed by Richard Foster Flint (1938b). In many ways Flint's study is one of the most ironic in the annals of geology. He presented a carefully worded argument that cited a considerable amount of field data. He stated that the scabland gravel was relatively fine: "Gravel coarser than pebble size is common only in the northern part of the tract" (Flint, 1938b, p. 472). This description was combined with the observation of relatively good size sorting and fair to good rounding to suggest, "a picture of leisurely streams with normal discharge" (Flint, 1938b, p. 472). It is obvious from Flint's sedimentological descriptions that he was giving most of his attention to the slackwater facies of the Missoula flood deposits in the various scabland channels.

One of Flint's most important arguments was that the surface form of the scabland deposits was that of "non-paired, stream-cut terraces in various states of dissection" (Flint, 1938b, p. 475). It was an idea that Bretz had introduced (Bretz, 1923a) and subsequently rejected after closer field study. Flint thought that Bretz' revised interpretation of the deposits as constructional bar forms could explain some, but not all of the field relationships. He suggested that a sequence of channel aggradation by normal proglacial outwash was followed by dissection to leave remnants of fill that occasionally resembled bar forms.

Flint (1938b) accepted Bretz's (1928b) arguments that the flood gravel often (1) occurred in the lee of island-like areas, (2) had rounded upper surfaces, and (3) exhibited a parallelism of surface slopes with the dip of underlying foresets. He argued that "terraces" had been extensively dissected by a downstream base level reduction. The "terraces" were prefentially preserved in the lee of island-like areas. In addition, the low precipitation plus the high permeability of the gravel prevented gullying, so the gravel deposits developed rounded slopes by dry creep. Finally, he showed that many of the gravel slopes did indeed truncate the underlying bedding. As specific cases, he argued that Bretz' Willow Creek bar, Staircase Rapids bar, Palouse Canyon bar, Midcanyon bar, and Shoulder bar were all simply terrace remnants. Subsequent studies have shown that three of these bars have prominent giant current ripples on their upper surfaces (Fig. 1.3).

Flint also described multiple scarps and benches on the Palouse loess. Instead of recording the high-water mark of the Spokane Flood (Bretz, 1928b, p. 701), he interpreted these scarps as evidence of lateral planation by proglacial streams. Subsequent studies in the Cheney-Palouse scabland by Patton and Baker (Chap. 6, this volume) reveal that these scarps resulted from differential erosion of Palouse Formation paleosols and from the exposure of calichified gravel underlying local areas of Palouse loess.

Flint traced the coarse scabland deposits downstream into the Pasco Basin. There he found that the deposits changed from sand and gravel to silt and fine sand containing erratic stones. He named the fine-grained facies the "Touchet beds." The deposits had already been described by Bretz (1928a, p. 325-328; 1929, p. 516-536; 1930b, p. 414), who ascribed them to ponded flood water; and by Allison (1933), who ascribed them to water ponded by ice jams. The silts are recognized only to a uniform elevation of about 350 m. The stratification ranges from rhythmic parallel bedding to cut-and-fill. The included erratic stones are granite, basalt, and other crystalline lithologies. Intense folding, fracturing, and clastic dikes imply slumping and sliding of the water-saturated silt on gentle subaqueous slopes. Flint thought that these relationships were most consistent with a large lake, which he proposed was ponded by a landslide dam or glacier ice in the Columbia gorge. Following Symons (1882) he named this water body Lake Lewis.

At this point Flint had the necessary tools to erect his hypothesis. The proglacial meltwater streams of normal discharge overran the northern margin of the Cheney-Palouse tract. This flow was derived from lobes of ice at the heads of the Cheney-Palouse and Telford-Crab Creek scabland tracts. Flint thought water from Lake Missoula (Bretz, 1930a) need not be involved. Instead, he observed that the discharge "was less than that of the Snake River today" (Flint, 1938b, p. 515).

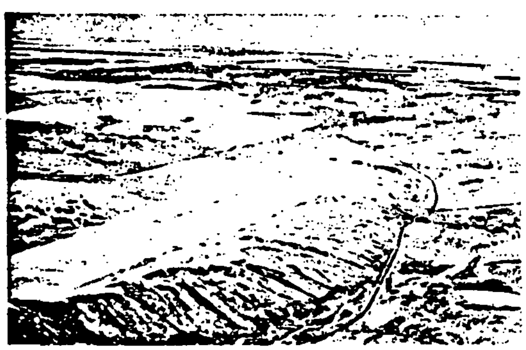

Figure 1.3. Oblique aerial photograph of Staircase Rapids bar. The bar is approximately 50 m high and composed of coarse flood gravel. The giant current ripples on the upper bar surface (left foreground) were actually first described by Flint (1938b) who did not recognize their origin. Bretz and others (1956, p. 1000-1002) later used these and other giant current ripple sets to demonstrate Flint's "faulty reasoning."

As Lake Lewis rose, the "leisurely" streams that Flint envisioned aggraded, forming a thick fill. This fill blocked preglacial tributaries to the Channeled Scabland, such as the Snake River, and formed marginal lakes which accumulated fine-grained sediments. The steep scarps on the Palouse loess were then cut by lateral planation of the streams flowing on this fill. When Lake Lewis finally drained, the streams gradually incised the fill to form terraces. Moreover, Flint was able to explain the enigmatic notched spurs and slotlike hanging canyons as the result of superposition of streams from the widespread fill rather than a consequence of divide crossing by catastrophic flood water.

Flint argued that the complex of anastomosing channel ways cut into basalt was a consequence of erosion by relatively small streams operating on various profiles. He stated that scabland-type erosion should occur wherever rock material with vertical planes of weakness is subjected to stream flow. As examples of such erosion he cited Red Rock Pass, Idaho, an outlet of pluvial Lake Bonneville (Gilbert, 1890). He also noted the scabland erosion at Twin Falls, Idaho, where the Snake River flows in a canyon nearly as spectacular as the scabland channels. He noted, "the . . . [basalt] flows yielded to the hydraulic force of the Snake River as similar flows on the Columbia Plateau yielded to the hydraulic force of proglacial streams, yet I am not aware that unusual floods have been held to have affected the upper Snake River" (Flint, 1938b, p. 492). These words were written 30 years too soon! Malde (1968) described the catastrophic outburst of Lake Bonneville that eroded the scabland forms at Red Rock Pass and Twin Falls.

In yet another ironic passage, Flint (1938b, p. 504-505) calculated the probable rate of filling for Lake Lewis at the modern discharge of the Columbia River. He stated, "the calculated time, 13 years 1 month, seems grossly inadequate for the deposition of the fill in the scabland tracts." He rationalized his interpretation, however, by referring back to the interpreted filling episode. Bretz' flood theory was so despicable that even circular reasoning could be employed to erect an alternative hypothesis.

A careful examination of Flint's (1938b) paper reveals that he observed and described the morphological feature which, more than any other, was absolutely incompatible with his elegant theory. On the surfaces of the scabland "terraces" he described an intricate microtopography of anastomosing channels, small depressions, and crescentic channels (Flint, 1938b, p. 475). In other areas he observed "mamillary undulatory topography." As an example he gives the precise location of the train of giant current ripples on the upper surface of Staircase Rapids Bar, 3 km north of Washtucna (Flint, 1938b, p. 486). Although the ripples that he describes are somewhat masked by overlying slackwater sediments, Flint (1938b, p. 499-500) even states the characteristic ripple magnitude: "The undulations are 20 to 100 feet long, and have amplitudes up to 10 feet. Their axes are generally transverse to the Snake River." How ironic that Flint was the first to accurately describe (without knowing what they were) the very feature that Bretz and others (1956) later presented as incontrovertible evidence for catastrophic flood flows (Fig. 1.3)!

It was Allison (1941) who published the first criticism of Flint's fill hypothesis for the origin of the Channeled Scabland. The first shortcoming noted was that the anastomosing channel patterns and deep rock basins could not have been eroded by "normal" streams. Second, Allison disputed Flint's correlation of the scabland gravels to the Touchet beds, suggesting that the Touchet sequence was younger than the gravels. Third, he agreed with Bretz that the peculiar shapes of the scabland deposits required extraordinary processes. The conclusion was that the complex jamming of various channels with ice was the only reasonable explanation for the unusual drainage patterns and depositional features.

Another example of the strong emotions evoked by the Spokane Flood controversy involves W. H. Hobbs, an eminent glacial geologist from the University of Michigan. He spent several weeks studying the terrain in southeastern Washington and prepared a paper explaining the landforms as the product of a "Scabland Glacial Lobe." Both Bretz and Flint reviewed the paper for the Geological Society of America, and both recommended rejection. The paper was then submitted to the American Philosophical Society, which had supplied part of the funds for the study. Bretz again reviewed the paper, and again it was rejected. Although a brief statement of the hypothesis was published (Hobbs, 1943), the main manuscript had to be published privately (Hobbs, 1947). The author expressed his feelings in the "Foreword" to his paper:

> "In the winter of 1942-43 I was listening with much interest to a lecture on the late geological history of the so-called Scabland area which is southwest of Spokane and close to the supposed southern front of the Pleistocene Cordilleran continental glacier. A map projected on the screen dozens of lakes, none of which transgressed its border, an almost sure indication that this lobate area had once been actually covered by a Pleistocene glacier lobe.
>
> Surrounding this lobe on the lecturer's map could be seen a broad apron of gravels, and enveloping the gravels were heavy deposits of silt. These relationships of glacier lobe to outwash and loess duplicated what I had observed in west Greenland. The lecturer explained, however, that the deposits represented upon his map had been laid down by a great flood of water of unknown origin, the "Spokane Flood."
>
> In the belief that my Greenland observations had given me an advantage in interpreting the evidence within the Scabland region, I then and there decided to make a personal study of it on the ground. Although two other very extended studies had already been made of it by Fellows of the Geological Society of America, and their conclusions had been published *in extenso* in its *Bulletin*, the Society provided me with a grant of money which made possible a new study of the area. This field investigation was carried out during two seasons, and the results and conclusions met with unusually enthusiastic general approval when they were presented to the Society in 1945 at its Pittsburgh meeting. Following tumultuous applause in the crowded section the discussion was throughout approving."

The Hobbs paper contains so many fundamental errors that one marvels at the absurd limits that were being stretched to find an alternative to catastrophic flooding as the cause of the Channeled Scabland. Hobbs (1947) argued that the scabland was a product of glacial scour and that the Palouse loess was deposited contemporaneous to this glaciation by anticyclonic winds off the ice that lay in the various "channels." He interpreted many scabland gravel deposits as moraine remnants modified by glacier-border drainage.

VINDICATION

At long last Pardee (1942) shared his observations of Glacial Lake Missoula that firmly indicated its role as the source of catastrophic floods through the Channeled Scabland. He noted that about 500 cubic miles of water were impounded behind a glacial lobe which occupied the basin of modern Lake Pend Oreille in northern Idaho. Pardee believed that this glacial dam had failed suddenly with a resultant rapid draining of the lake. Evidence for this failure included severely scoured constrictions in the lake basin, huge bars of current-transported debris (Fig. 1.4), and giant current ripple marks with heights of 50 feet and spacings of 500 feet (Fig. 1.5). Lake Missoula was the obvious source for the catastrophic flood flows required by Bretz' hypothetical origin of the Channeled Scabland (Fig. 1.6). Pardee did not state the connection, perhaps leaving that point generously to Bretz. Even Alden remained cautious to the end. His last published report on Lake Missoula observed (Alden, 1953, p. 155): "Abrupt release of water from lowering of the ice dam . . . might result in floods of great magnitude. . . . Each may, *perhaps,* have been the origin of many violent floods that are *supposed* to have swept over the scablands."

In the summer of 1952, Bretz, then nearly 70 years old, returned for his last summer of field work in the Channeled Scabland. The purpose was to investigate new data that had been obtained in surveys for the Bureau of Reclamation's Columbia Basin project. Professor H. T. U. Smith accompanied him, acting in the field as "skeptic for all identifications and interpretations" (Bretz and others, 1956). With the aid of Mr. George E. Neff of the Bureau of Reclamation that study (Bretz and others, 1956) answered with meticulous detail all previous criticisms of the flood hypothesis.

Central to the 1956 investigation was the study of the scabland depositional features. Extensive excavations for the irrigation project and new topographic maps proved that the gravel hills called bars by Bretz (1928b) were indeed that, subfluvial depositional bedforms. Most convincing of all was the presence of giant current ripples on the upper bar surfaces. These showed clearly that bars 30 m high were completely inundated by phenomenal flows of water. Numerous examples of giant current ripples were found on the same bars which Flint had interpreted as terraces. Such features could only have been produced by the flow velocities associated with truly catastrophic discharges. Bretz and others (1956) and Bretz (1959) modified Bretz' earlier interpretations to allow for several episodes of flooding. The central theme of their study, however, was that only a hypothesis involving flooding could account for all the features of the Channeled Scabland. More recent studies of the

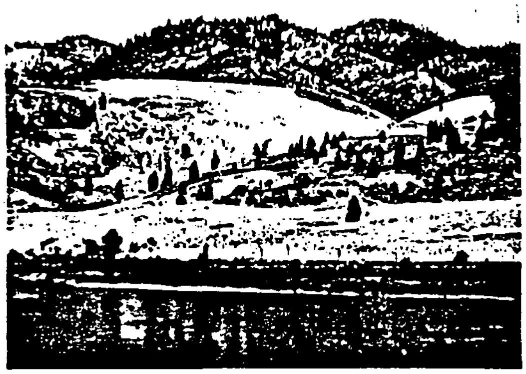

Figure 1.4. Large "gulch fill" formed at the mouth of a tributary canyon along the Flathead River, Perma, Montana. The deposit is an eddy bar (Baker, 1973a) formed during the rapid draining of glacial Lake Missoula. First recognized by Pardee (1942) this gravel deposit was later breached by a small stream to form the V-shaped notch visible at right. The low terrace in the foreground is composed of lacustrine silt.

Figure 1.5. Giant current ripples at Camas Prairie, north of Plains, Montana. The ripples are composed of gravel and consist of ridges up to 15 m high and spaced as much as 200 m apart. The ripples cover approximately 10 km² of the northern Camas Prairie. Faint strandlines of Lake Missoula are visible in the background.

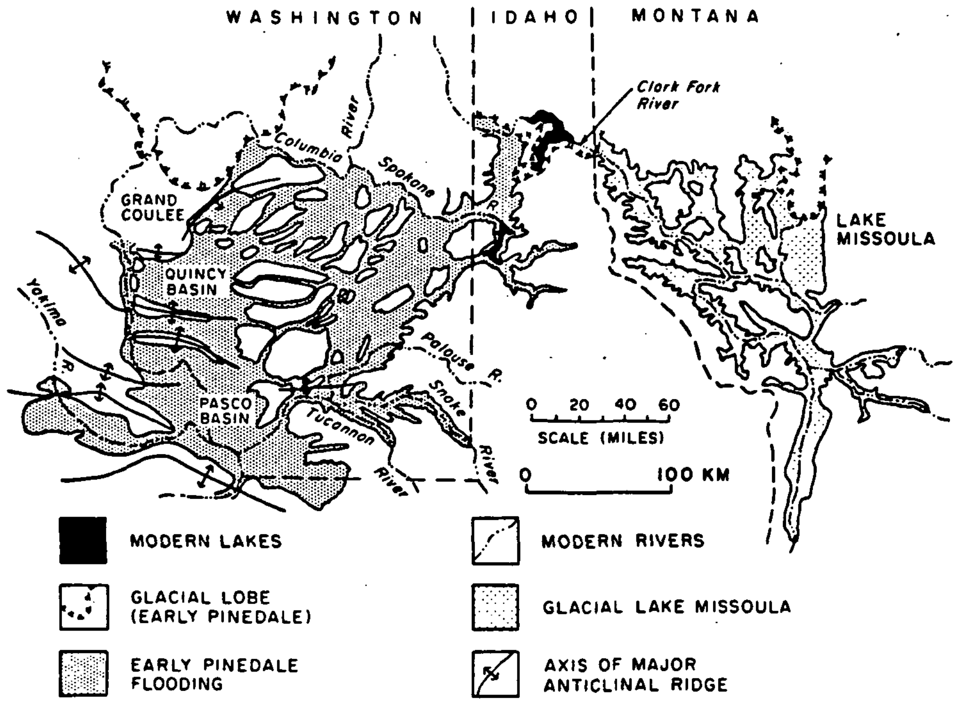

Figure 1.6. Relationship of glacial Lake Missoula to the Channeled Scabland of eastern Washington (Baker, 1973a).

Quaternary geology of eastern Washington have accepted this reasoning (Trimble, 1963; Fryxell and Cook, 1964; Richmond and others, 1965; Baker, 1973a).

Perhaps the final words on the Channeled Scabland controversy were delivered following a field trip, Field Conference E of the 7th Congress, International Association for Quaternary Research. During August, 1965, an international party of geologists observed the evidence in Montana for Lake Missoula's catastrophic outbursts. They then traveled through the Channeled Scabland studying the giant current ripples, flood gravel bars, and scabland erosion forms. Dr. Bretz was unable to attend the trip because of health. When the field party reached Pullman, they sent a long telegram to him at Homewood, Illinois. The telegram opened with "greetings and salutations" and closed with the sentence, "We are now all catastrophists" (Bretz, 1969, 1973).

DISCUSSION

When Bretz published his work on the Channeled Scabland, the paradigm of Geology was uniformity. The Spokane Flood hypothesis appeared to contradict the uniformitarian tradition that made geology a science in the nineteenth century. Indeed it was not until after 1840 that the flood theory fell into serious decline. The catastrophist idea of the Noachian debacle was finally laid to rest when Louis Agassiz showed that his glacial theory could explain erratics, striations, till, fluvioglacial activity, etc. Old ideas die hard, however, and catastrophist absurdities still appeared in the literature of the early 1900's (as they do even today). Little wonder then that Bretz' Spokane flood hypothesis appeared as an anathema to many of his contemporaries.

Simultaneously the Spokane Flood hypothesis established a conflict between two important cornerstones of geological philosophy: (1) the triumph of the glacial theory over diluvian myth, and (2) the scientific tolerance of outrageous hypotheses. It is a classic dilemma for the scientist to distinguish absurdity from outrage. A foolish idea is always self-evident, but not so with the rare, creative insight that happens to pass all reasonable bounds in the consensus of knowledge. The remarks of a former president of our society: "How narrowly limited is the special field, either in subject or locality, upon which a member of the Geological Society of America now ventures to address his colleagues. . . . I wonder sometimes if younger men do not find our meeting rather demure, not to say a trifle dull; and whether they would not enjoy a return to the livelier manners of earlier times . . . (Their) feeling of discouragement must often be shared by the chairman of a meeting when, after his encouraging invitation, 'This interesting paper is now open for discussion,' only silence follows. . . . We shall be indeed fortunate if geology is so marvelously enlarged in the next thirty years as physics has been in the last thirty. But to make such progress violence must be done to many of our accepted principles."

After speaking these words in 1926, William Morris Davis made a case for the value of outrageous geological hypotheses, even suggesting that geologists seriously consider "the Wegener outrage of wandering continents." He concluded by saying that the valuable outrage was that which encouraged the contemplation of other possible behaviors. Such outrages deserve contemplation followed not, he states, "by an off-hand verdict of 'impossible' or 'absurd', but a contemplation delibrate enough to seek out just what conditions would make the outrage seem permissible and reasonable."

Needless to say, W. M. Davis was one of the first to accept Bretz' interpretation in the 1920's. It is a commentary on those years that others were not so tolerant. "During all those years, I was fighting for my professional career." (Quotation of Dr. Bretz by the Seattle Times, July 11, 1971.) Bretz himself explored the consequences of his "outrage." His 1956 paper resoundingly confirmed the catastrophic flood theory by answering in meticulous detail all the previous objections to his grand hypothesis. It took over 30 years and the coming of a new generation of geologists for his theory to gain general acceptance.

The Spokane Flood controversy is both a story of ironies and a marvelous exposition of the scientific method. One cannot but be amazed at the spectacle of otherwise objective scientists twisting hypotheses to give a uniformitarian explanation to the Channeled Scabland. Undoubtedly these men thought they were upholding the very framework of geology as it had been established in the writings of Hutton, Lyell, and Agassiz. The final irony may be that Bretz' critics never really appreciated the scientific implications of Agassiz' famous dictum, "study nature, not books." Perhaps no geologist has understood and lived the spirit of those words more enthusiastically than J Harlen Bretz.

As the Viking spacecrafts were orbiting Mars in the summer of 1976, the cameras were trained on the great Martian channel systems. They revealed uplands streamlined by fluid flow, eroded scabland on the channel floor, and many other features that we now know to be diagnostic of bedrock erosion by catastrophic flooding. Fifty years after J Harlen Bretz' theory of scabland erosion on the Columbia Plateau was being denounced at an infamous meeting of the Washington Academy of Science, Viking scientists were using Bretz' well-documented studies of the Channeled Scabland as the major earth-analog to Martian channel erosion. Few geological concepts, born amid bitter controversy over a half century ago, have continued to have such relevance to our science.

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Chapter 2

Quaternary Geology of the Channeled Scabland and Adjacent Areas

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ABSTRACT

The Quaternary history of the Channeled Scabland is characterized by discrete episodes of catastrophic flooding and prolonged periods of loess accumulation and soil formation. The loess sequence is correlated with Richmond's Rocky Mountain glacial chronology. Two pre-Bull Lake (pre-Illinoian) loess units are characterized by siliceous petrocalcic horizons. The Bull Lake loess (Palouse Formation) apparently accumulated episodically be downwind accretion, followed by periods of relative stability and soil formation. The Palouse paleosols have platy calcic horizons but do not show petrocalcic horizons. Pinedale (= Fraser Glaciation) loess is paler in color than the older units; its paleosol calcic horizons lack the platy structure of the older loess paleosols.

At least five major catastrophic flood events occurred in the general vicinity of the Channeled Scabland. The earliest episode occurred prior to the extensive deposition of the Palouse Formation. Its surviving records are but fragmentary. Probable Missoula flood deposits in the Cheney-Palouse scabland tract are overlain by the younger pre-Palouse loess and a petrocalcic soil profile. Flood deposits in the Quincy Basin came from an unknown western source across Babcock Ridge. The Quincy Basin flood deposits are overlain by other flood deposits of probable Bull Lake age (Illinoian) also derived from a western source. Catastrophic flooding from Lake Bonneville affected the southern margin of the Channeled Scabland about 30,000 years B.P.

The last major episode of flooding occurred between about 18,000 and 13,000 years ago. It probably consisted of two outbursts from Glacial Lake Missoula. The earlier outburst predates the Vashon maximum (= Withrow Moraine of the Okanogan ice lobe). This flood affected Moses Coulee, the Grand Coulee (prior to its present configuration), and the eastern Channeled Scabland (Telford-Crab Creek and Cheney-Palouse scabland tracts). A second flood, probably involving less volume than the first, coincides with the deglacial phase of the Okanogan ice lobe. It mainly affected the Columbia River northwest of the Channeled Scabland downstream from a hydraulic constriction of the canyon at the site of Coulee Dam. The last phase of that flood probably also involved catastrophic flow down the Grand Coulee. Slackwater facies of this flood contain the Mount St. Helens set "S" ash erupted about 13,000 years B.P. according to D. R. Mullineaux and co-workers.

INTRODUCTION

The Channeled Scabland of eastern Washington (Fig. 2.1) consists of a spectacular complex of anastomosing channels, cataracts, loess "islands," and immense gravel bars created by the catastrophic fluvial erosion of the loess and basalt of the Columbia Plateau (Bretz and others, 1956). The erosion and deposition that produced the scabland topography resulted from the failure of the ice dam impounding glacial Lake Missoula. At its maximum outflow, near the end of the last major Pleistocene glaciation, the lake discharged as much as 21.3 x 10^6 m^3/sec into the vicinity of Spokane, Washington (Baker, 1973a). Recently the origin and history of the Channeled Scabland has assumed new significance because of morphologic similarities to outflow channels of probable flood origin on Mars (Baker and Milton, 1974; Sharp and Malin, 1975; Masursky and others, 1977). Despite detailed study since the 1920's (Bretz, 1923a, 1928c, 1932b, 1959, 1969; Richmond and others 1965), the exact number and timing of major floods which resulted from the outpourings of Lake Missoula remains a major unresolved problem in the Quaternary history of eastern Washington. This paper will summarize the general Quaternary geology of the Channeled Scabland and present some new data on the number of catastrophic floods.

PHYSIOGRAPHY, CLIMATE, AND SOILS

Freeman and others (1945) have formalized the physiographic divisions of the regions covered in this report. The entire study area lies in the Columbia Basin Subprovince of the Columbia Intermontane Physiographic Province. The Columbia Basin is a regional lowland surrounded by the Blue Mountains to the south, Cascade Mountains to the west, the Okanogan Highlands to the north, and the mountains of northern Idaho in the east. The region is also informally called

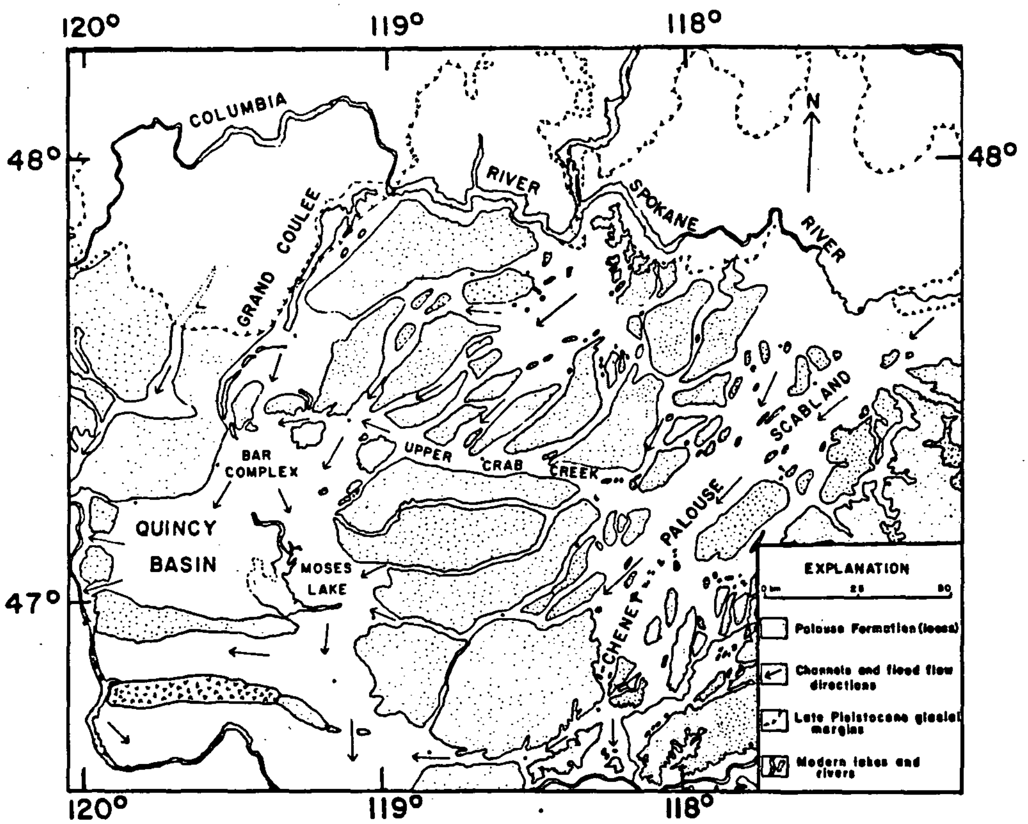

Figure 2.1. Map of the Channeled Scabland in eastern Washington showing the distribution of channels and the general extent of loess (Palouse Formation) that was not stripped away by the last major episode of flooding.

the "Columbia Plateau," but intense folding in the western part produces a series of basins with intervening ridges. The basalt bedrock of the basin has the regional aspect of a structural basin.

The western part of the Columbia Basin, termed the "Yakima Folds," consists of a series of anticlinal ridges extending eastward from the Cascade Mountains. Several of these ridges are transected by the Columbia River. From north to south these are the Frenchman Hills, Saddle Mountains, and Horse Heaven Hills.

The northwest portion of the Columbia Basin, the Waterville Plateau, was not appreciably disturbed by the Neogene folding. A major canyon, Moses Coulee, is deeply excavated into the Waterville Plateau and extends southwestward from its center to the Columbia River Valley.

The far eastern part of the Columbia Basin is characterized by relatively undeformed basalt overlain by as much as 75 m of Pleistocene loess. The loess has been dissected to form a rolling topography known as the Palouse Hills. Elevation of the Palouse Hills declines from about 750 m on the northeastern margin to 100-120 m in the southwest. This gradient reflects the regional dip of the basalt toward the center of the basin.

Between the Palouse Hills and the Yakima Folds, extensive stripping occurred of the loess mantle by the catastrophic flooding of glacial Lake Missoula. The maturely dissected loess presents an abrupt contrast with the black cliffs and ragged appearance of the flood-eroded scabland tracts. Steep scarps on the eroded channel margins resemble wave-cut headlands rising above a desiccated sea. Bretz (1923b) named this region the Channeled Scabland.

The Columbia Plateau lies in the rain shadow of the Cascade Mountains. Precipitation in the western basins (elevation 180-300 m) is less than 20 cm per year, but increases with elevation northeastward, reaching nearly 50 cm per year on the updip margins of the plateau (elevation 850 m). The low precipitation results in a lack of perennial drainage through the huge ancient channels, called "coulees." The Columbia and Spokane Rivers, which are deeply entrenched along the northern and western margins of the plateau, intercept most of the drainage from the areas of higher precipitation in the bordering mountains.

The precipitation pattern is paralleled by a change in Great Soil Groups to the northeast. Sierozem soils form in the dry, southwestern portions of the plateau. Toward the northeast, Brown, Chestnut, and Chernozem soils appear in successively wetter portions of the plateau. These soils occur only on the loess and other fine-grained sediments. The eroded scabland channels are generally devoid of these parent materials for soil formation.

ANTEDILUVIAN EVENTS ON THE COLUMBIA PLATEAU

The Yakima Basalt comprises the bedrock in all but a few parts of the Channeled Scabland. This basalt unit is part of the extensive Neogene eruptions of plateau basalts that cover over 250,000 km² in parts of Washington, Oregon, and Idaho. Most of the lava was erupted during the Miocene. The lava flows are exceptionally thick, and several can be traced over 150 km. Considerable structural and lithologic variation can be found in the basalt sequence, including jointing patterns, pillow-palagonite complexes, sedimentary interbeds (from lakes on the Miocene land surface) and geochemical variation. On the north and east margins of the plateau, the basalt is interbedded with extensive deposits of siltstone and shale of the Latah Formation, deposited as drainages were blocked by the basalt outpourings.

Deformation of the basalt sequence was most extensive during the Pliocene. The entire Columbia Plateau was regionally tilted from an elevation of about 760 m in the northeast to about 120 m in the southwest near Pasco, Washington. Superimposed on the regional structure are the east-west fold ridges described earlier. The upraised northern rim of the plateau is especially significant for the flood history of the Channeled Scabland. Only a truly phenomenal quantity of water could fill the great canyon of the Columbia River between Spokane and Coulee Dam. That filling would be necessary to have water spill over the northern rim of the plateau and flow southwestward, carving the great scabland channels.

During the Pliocene, the great structural basins of the western scablands accumulated a sequence of partly consolidated silt, gravel, and clay known as the Ringold Formation (Flint, 1938a; Newcomb, 1958). The age of this formation was poorly interpreted until Eric Gustafson studied the extensive upper Ringold vertebrate fauna from the White Bluffs area. Gustafson's White Bluffs fauna is early Blancan (Pliocene) in age. The faunal assemblage correlates best with the Hagerman fauna of Idaho, dated at about 3.5×10^6 years B.P.

Gustafson (1973) interprets the Ringold Formation as a sequence of stream-channel conglomerate, point-bar sandstone, and overbank deposits within a major fluvial depositional system. The predominance of browsing forms among the large mammals (interpreted from tooth form) including especially *Bretzia, Platygonus,* and *Megalonyx,* suggests that the Ringold flood plain supported considerable riparian forest. Circumstantial evidence suggests a strongly seasonal climate with yearly rainfall between 25 and 50 cm. Today the rainfall averages 20 cm, and the vegetation is xerophytic (sagebrush).

The Ringold Formation does not occur in the eastern Columbia Plateau, where late Pliocene and Pleistocene sedimentation was largely eolian, as expressed in a complex blanket of loess sheets (Fig. 2.2). The loess units provide a fairly complete Pleistocene chronology that was correlated by Richmond and others (1965) to Richmond's (1965) glacial chronology of the Rocky Mountains.

The oldest loess units are considered to be pre-Bull Lake in age. By the revised glacial chronology of Pierce and others (1976), Bull Lake time correlates to Illinoian, about 125,000 to 200,000 years ago. Two pre-Bull Lake loess units can be recognized in the field by the well-indurated, siliceous petrocalcic horizons that formed on them. The older of the two petrocalcic horizons is colored pinkish by an associated oxidized tuff. These petrocalcic horizons consist of as much as 0.6 m of roughly horizontal carbonate laminae over a calcic horizon of carbonate-plugged loess. This profile is an extremely strong soil horizon that qualifies for the designation "K horizon" (Gile and others, 1966), defined by the presence of 50% or more $CaCO_3$. Fryxell and Cook (1964) describe as much as 3 m of B horizon associated with the pre-Bull Lake loess units.

Figure 2.2. Oblique aerial photograph of a road cut through "Palouse Hills" topography about 5 km west of Washtucna, Washington. The white layers are calcic paleosols in the loess sequence.

Most of the loess on the Columbia Plateau is correlated to Bull Lake Glaciation (Richmond and others, 1965). This loess, which has a thickness of up to 75 m (Ringe, 1970), is formally designated the Palouse Formation (Newcomb, 1961). Richmond and others (1965) recognize three cycles of soil formation in the Palouse Formation, but unpublished observations by the author indicate that more cycles may have occurred. Each soil shows a mature profile with a well-developed textural B horizon. The underlying Cca horizon is strongly calcareous and has a well-developed platy structure. Unlike soils on older loess units, however, this calcic horizon is less thoroughly cemented and does not qualify as a K horizon. Boulders derived from the Cca or B horizons of Palouse Formation soils are commonly found in deposits of the late Wisconsin flooding on the Columbia Plateau.

The Bull Lake age of the Palouse Formation was established by its relationship to glacial deposits in the vicinity of Spokane. It is obvious, however, that the detailed stratigraphic information in this loess sequence is a better record of mid-Pleistocene events than is the glacial sequence to the north. The Palouse Formation should be given detailed study, employing modern techniques as described by Kukla (1975) for the loess of Czechoslovakia. Paleomagnetic and tephrachronologic studies would probably yield an important new stratigraphic interpretation for the Palouse Formation and older loess units on the Columbia Plateau.

A major unconformity separates the Palouse

Formation from younger loess, correlated to the Pinedale Glaciation by Richmond and others (1965). The Pinedale loess is much paler in color than the Palouse Formation. Associated soil profiles have only weakly developed textural B horizons and lack the structural Cca or K horizons of older soils. $CaCO_3$ occurs in veins or nodules but not in continuous plates. In contrast, soils of Holocene age show A-C profiles of minimal development.

The soils within a typical loess hill show periods of stability during the progressive northwesterly accumulation of loess. A section through a loess hill near La Crosse (SE ¼, Sec. 1, T. 15N., R. 39E.) is shown in Figure 2.3. The three Bull Lake loess units have been truncated by a loess unit of a younger age (Pinedale?) which mantles the surface of the whole hill. The entire body of Bull Lake loess has a dull brown color (7.5YR 6/3). Color or textural B horizons were not apparent in this section. The younger loess units are dull yellow orange (10YR 7/3).

PRE-WISCONSIN FLOODING IN THE CHANNELED SCABLAND

The eastern portion of the Channeled Scabland (Fig. 2.4) contains several exposures of pre-Wisconsin flood gravel. The most complete section is exposed in a railroad cut through a gravel bar on the downstream end of a residual loess island about one kilometer west of Marengo, Washington (location 1, Fig. 2.4). The cut exposes a succession of two flood gravel units separated by three layers of loess (Fig. 2.5). These three loess units are each capped by a pedogenic calcic horizon. The lower flood deposit is a poorly sorted mixture of basalt and loess pebbles and cobbles in a matrix of granule-sized basalt grains. The texture is similar to that of the extensive deposits of the last major episode of scabland flooding. Typical of many scabland gravel deposits is the presence of loess cobbles, implying fluvial transport in suspension which prevented destruction of the loess. Cobbles in this deposit are dull orange (7.5 YR 7/4) in color and have black manganese dioxide staining on their surfaces.

The lower flood gravel is overlain by 1 m of dark yellowish brown (10 YR 6/6) loess which is capped by a petrocalcic horizon 60 cm thick. There is also a well-developed argillic horizon on this loess characterized by a pronounced prismatic structure. Grain-size analysis indicates that the prismatic structure is associated with a relatively strong illuvial textural B horizon. Root casts infilled with caliche from the overlying petrocalcic horizon also occur in this argillic zone. The overlying petrocalcic horizon has a platy structure and completely plugs the uppermost part of this loess unit. Although no carbonate analysis was made on this horizon, its morphology is nearly pure carbonate (caliche) with little loess matrix. The horizon therefore qualifies for designation as a K horizon (> 50% carbonate) as described by Gile and others (1966). This pedocal soil appears to have been superimposed on the older, pedalfer soil (textural B horizon) and then eroded to its resistant carbonate layer prior to new loess deposition.

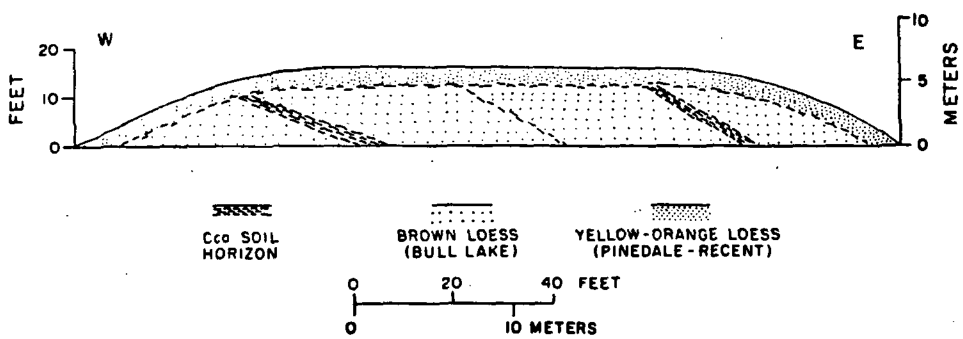

Figure 2.3. Loess stratigraphy exposed by a road cut 2.5 km east of La Crosse, Washington.

The K horizon is overlain by two pale yellowish brown (10 YR 7/3) loess units, each 50-60 cm thick and each capped by calcic soil horizons 10-20 cm thick. These loess layers lack the prismatic structure of the earlier loess and have moderately developed textural B horizons. The thin calcic horizons have a platy structure. Unlike the lower caliche, they do not qualify as petrocalcic horizons because the loess is not continuously cemented and indurated.

The two calcified loess units are overlain by approximately 1 m of granule-to-cobble sized gravel that was deposited by the latest flood to cross the Cheney-Palouse scabland. The loess gravel present in this deposit is predominantly composed of the brownish loess from the Palouse Formation, although a few of the reddish loess cobbles are also present. The gravels are covered by 1.5 m of late Pinedale loess on which a relatively weak A-B-C soil profile has formed. This

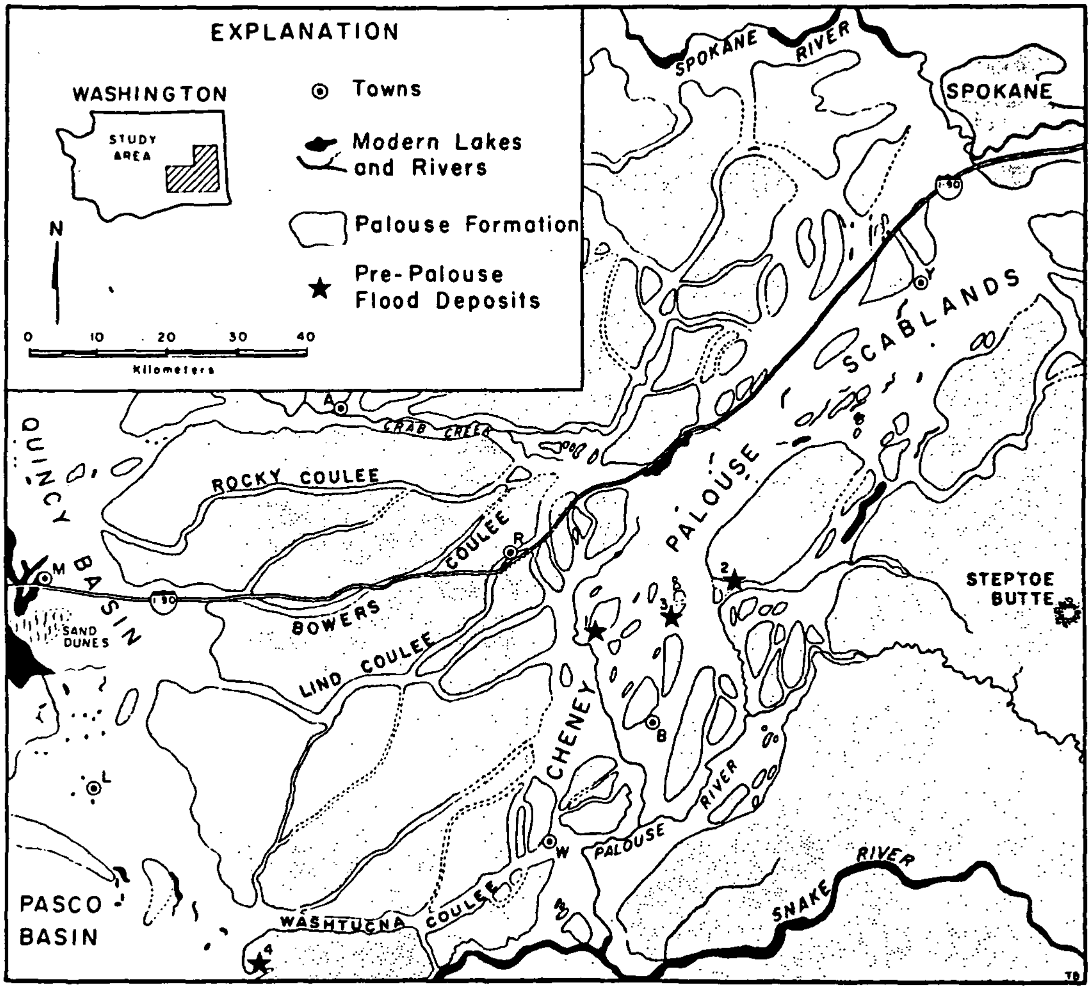

Figure 2.4 Location map showing the generalized distribution of the Palouse Formation in eastern Washington. Areas not covered by Palouse Formation were *scoured by the early Pinedale catastrophic breakout flood* of Lake Missoula. Pre-Palouse flood deposits have been found at the numbered localities: (1) Marengo, (2) Revere, (3) Macall, (4) Old Maid Coulee. Towns are indicated by letters: (A) Odessa, (B) Benge, (L) Othello, (M) Moses Lake, (R) Ritzville, (W) Washtucna, (Y) Cheney.

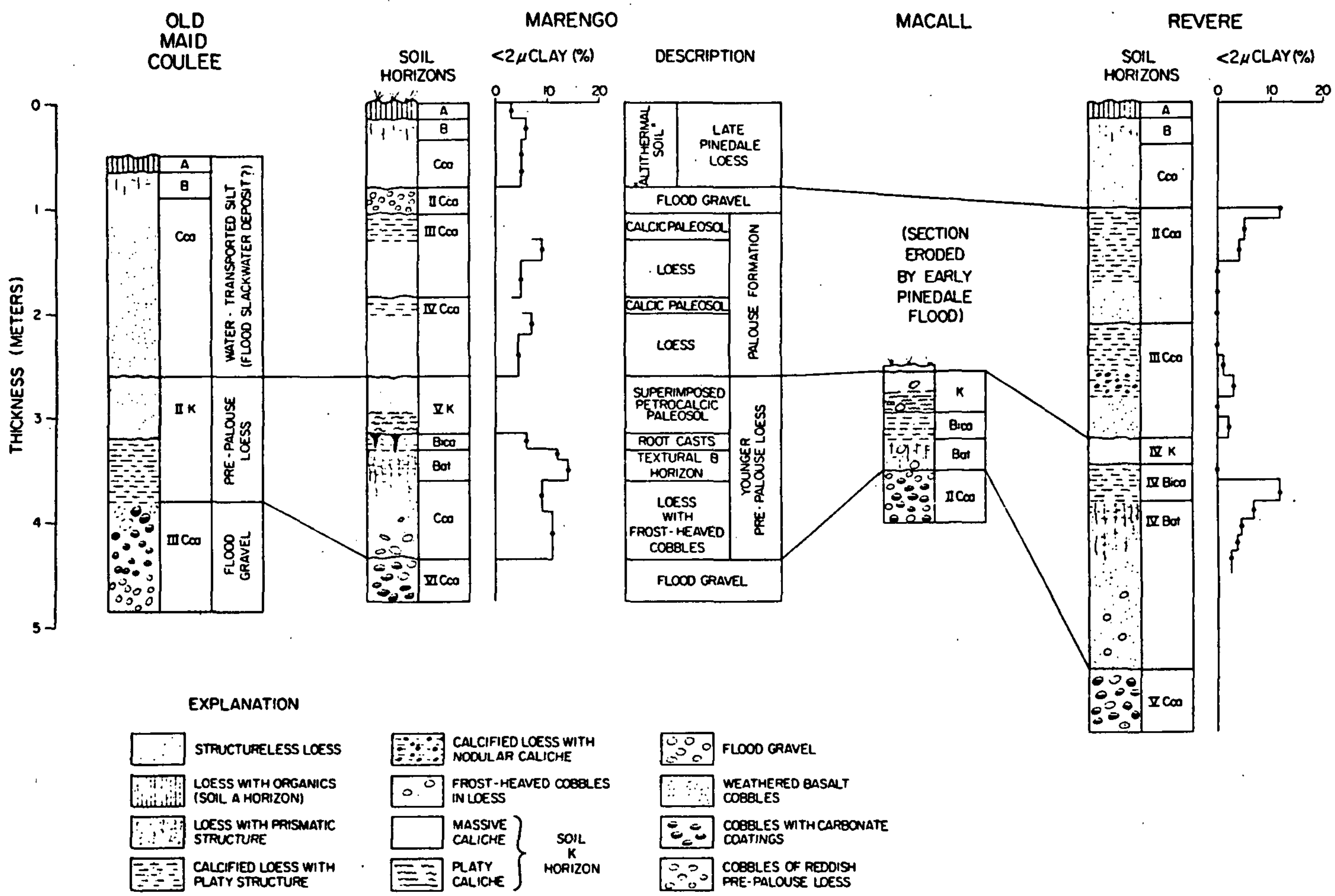

Figure 2.5. *Stratigraphic section at Marengo, Washington and its probable correlation to other sections in the Channeled Scabland.*

latter profile corresponds to the late Wisconsin-early Holocene soil that is common throughout the region on post-scabland loess deposits (Fryxell, 1965; Baker, 1973a). The uppermost gravel unit threfore correlates with the extensive deposits of the last major episode of scabland flooding.

The lower flood gravel at Marengo has also been located at Revere (Locality 2, Fig. 2.4) and Macall (Locality 3, Fig. 2.4). These three new sections probably also correlate with a flood gravel exposed on a divide near the upper reaches of Old Maid Coulee (Locality 4, Fig. 2.4), a section previously interpreted by Bryan (1927, p. 27), by Flint (1938b, p. 518), and by Bretz and others (1956, p. 1006). It consists of foreset-bedded gravel overlain by calcified loess, which is, in turn, overlain by pale, water-transported silt (Fig. 2.6). About 20% of the basalt cobbles in the upper parts of the gravel are completely rotten, indicating prolonged weathering. The loess unit is capped by 60 cm of platy calcium carbonate, a soil petrocalcic or K horizon. The erosion surface at the top of the caliche is overlain by silt that is dull yellow orange color (10 YR 6/3). It lacks the darker coloration and well-developed buried soils of the Palouse Formation. Moreover, its position on a divide and its uniform texture suggest that it may be suspended load from the last major flood that was deposited in a slackwater area. Flint (1938b) believed that the unit was a lake silt. He did suggest that this silt was approximately contemporaneous with the gravel of the main scabland channels, which are now considered to be approximately 13,000 years B.P. in age (Mullineaux and others, 1977).

The loess unit and petrocalcic paleosol capping the old flood gravels at all four sections probably correlates with the younger pre-Palouse Formation loess and soil described earlier. Unlike other pre-Bull Lake flood deposits recognized in the western scablands at George and Winchester (Baker, 1973a), there was a significant pre-Palouse loess accumulation on these early flood gravels prior to a major period of soil formation (an interglaciation?). Moreover, the reddish loess cobbles in the flood gravel probably were eroded from the older of the two pre-Palouse loess units. The first episode of scabland flooding appears to have occurred during a Pleistocene glacial maximum prior to the maximum that produced the Bull Lake Glaciation. Subsequent loess accumulation probably included a relatively humid interval that produced the pronounced textural B horizon, and it was followed by an arid interval that superimposed a petrocalcic horizon on the loess. An erosional episode, probably deflation, then stripped the surface horizons down to the resistant petrocalcic layer. These latter events all characterized the latest pre-Bull Lake glacial and interglacial record on the Columbia Plateau.

Bull Lake time is represented at the Marengo profile (Fig. 2.5) by two of the three known Palouse Formation loess units. This circumstance is typical of the cycles of soil formation, deflation, and lateral accretion that characterize the Palouse Formation (Lewis, 1960; Fryxell, 1966). During the Bull Lake Glaciation loess was accreting behind pre-existing obstacles in a general northeastward direction. Typical exposures through loess hills show that accretion was followed by stability and soil formation. Subsequent

Figure 2.6. Stratigraphic section at Old Maid Coulee.

deflation would remove surficial soil horizons down to the resistant calcic layers. The cycle was then repeated as more deposition occurred upwind of the previous deposit. This pattern of lateral accretion plus the opportunity for stream dissection of the topography make it unlikely to find all three Palouse Formation loess units in one vertical section.

An unresolved problem is the correlation of the pre-Wisconsin Cheney-Palouse flood to the pre-Wisconsin flood deposits of the western Quincy Basin (Baker, 1973a, p. 8). The section there (NW ¼, Sec. 31, T. 19N. R.24E) shows a typical foreset flood gravel containing boulders as large as 1 m in diameter. The uppermost 60 cm of the gravel is capped by a horizontally laminated petrocalcic horizon (Fig. 2.7). This is underlain by 30-60 cm of carbonate-cemented gravel. Local carbonate cementation in the coarser foresets occurs to a depth of 3 m. Carbonate coatings on the underside of cobbles occur to a depth of 4 m. Weathering rinds on the basalt cobbles in the upper 1.5 m of the gravel exceed 7.5 cm in thickness. Many cobbles are completely rotten. The gravel below a depth of 3 m shows no evidence of weathering. This weathering profile is much more intense than that noted on the pre-Wisconsin Cheney-Palouse flood deposits. However, the definitive sequence of loess units is absent in the western Quincy Basin, so a precise correlation remains speculative.

Richmond and others (1965) also recognize a Bull Lake episode of scabland flooding from widely scattered deposits on the Columbia Plateau. The best evidence occurs at Winchester Wasteway in the Quincy Basin, where the Bull Lake flood deposits overlie the older pre-Wisconsin flood deposits seen at George (Baker, 1973a, p. 8).

THE BONNEVILLE FLOOD

Malde (1968) studied the catastrophic flood produced by the overflow and rapid lowering of Pleistocene Lake Bonneville (Fig. 2.8). He traced the course of this flood through the Snake River Plain of southern Idaho to Hells Canyon. Malde interpreted the age of this event to be about 30,000 years B.P., based on a radiocarbon date for molluscan fossils associated with flood debris and on the relict soil profile developed on the flood gravel (Melon Gravel). The soil has a thick calcic

Figure 2.7. Oblique aerial photograph of a sanitary landfill at George, Washington. The white layer is a prominent petrocalcic horizon developed at the top of coarse flood gravel. The gravel was laid down by a deep catastrophic flood that crossed the western rim of the Quincy Basin and flowed southeastward into the Quincy Basin.

Figure 2.8. View of the Snake River Canyon downstream of Perrine Memorial Bridge in Twin Falls, Idaho. Scabland erosion of volcanic rocks, here produced by the Bonneville flood, is very similar to that produced by Missoula flooding of the Channeled Scabland.

horizon extending to depths greater than 2 m (Fig. 2.9). The soil on the Bonneville Flood deposits is believed to have formed during and since the mid-Wisconsin (Bull Lake-Pinedale) interglaciation or within the last 30,000 years.

Downstream from Hells Canyon at Lewiston, Idaho, probable Bonneville flood deposits are overlain by slackwater surge deposits from the last major episode of scabland flooding (Fig. 2.10). Because Bonneville flooding was confined to the Snake River Canyon, it skirted to the south of the Channeled Scabland. Nevertheless, studies in the Pasco Basin should eventually recognize Bonneville Flood deposits in association with Missoula flood deposits.

THE LATE PLEISTOCENE DILUVIAN EVENTS

The late Pleistocene glacial record of northwestern Washington and southwestern British Columbia is very well documented through detailed radiocarbon dating (Armstrong and others, 1965; Fulton, 1971; Easterbrook, 1976). The last major glaciation, called the Fraser Glaciation, extended from about 20,000 to 10,000 years ago. This is also the time interval that produced the last major episode of scabland flooding. At least one and probably two major outbursts of Lake Missoula occurred during this interval. Although the evidence for the floods is overwhelming, the precise dating of the events in this interval is a matter of current controversy. This section will attempt to summarize the diverse arguments.

Glacial Stratigraphy

The Fraser Glaciation is precisely dated on the western side of the Cascade Mountains. During the Evans Creek Stade large alpine glaciers advanced to their maxium extent into the Puget lowland. This was followed by an advance of Cordilleran ice into the lowland from the north sometime after 19,000 years B.P. (Easterbrook, 1976). If the Pend Oreille Lobe of the Cordilleran Ice Sheet correlated with the Puget Lobe, then Lake Missoula could not have formed until after 19,000 years B.P. The Puget Lobe reached its maximum extent 14,000 to 15,000 years ago, during the Vashon Stage.

Figure 2.9. Exposure of Melon Gravel in large boulder bar at mile 161 in the Melon Valley area of the Snake River Canyon, Idaho (Malde, 1968, p. 33). The relict paleosol here is moderately strong, lacking the petrocalcic horizons of pre-Wisconsin flood gravels in the Channeled Scabland.

Figure 2.10. Slackwater facies of the last episode of scabland flooding overlying probable flood gravel of the Bonneville Flood at a gravel pit 5.5 km south of Lewiston, Idaho. The foresets and lithologies indicate that the gravel was deposited by flows coming *down* the Snake River through Hells Canyon. The slackwater deposits were deposited by backwater flooding *up* the Snake River from the mouth of the Palouse River, a distance of over 120 km. This deposit was described as Tammany Bar by Bretz (1969, p. 531-532).

Between 14,000 and 13,000 years B.P. a major recession occurred in the Vashon glacier of the Puget lowland. The interval from 13,500 to 11,500 years ago is characterized by relative stability of Cordilleran ice during the Everson Glaciomarine Interstade, which was followed by a rather minor readvance perhaps between 11,500 and 11,000 years B.P.

The glacial chronology on the Columbia Plateau is severely hampered by a lack of radiocarbon dating. Work on the Waterville Plateau by Hanson (1970) and Easterbrook (1976) shows that an episode of major scabland flooding occurred prior to the major advance of the Okanogan Lobe on to the Columbia Plateau. That advance produced a spectacular moraine, the Withrow Moraine (Waters, 1933; Flint, 1935). This is estimated to have formed about 14,000 to 15,000 years B.P. (Fryxell, 1962; Waitt, 1972a; Easterbrook 1976). Relationships near Jamison Lake show the moraine overlying a Moses Coulee flood gravel bar (Fig. 2.11). Subsequent outwash (Fig. 2.12) and dramatic ice stagnation features record a rapid retreat of the Okanogan Lobe (Hanson, 1970; Waitt 1972a), perhaps contemporaneous to the analogous phase of the Puget Lobe.

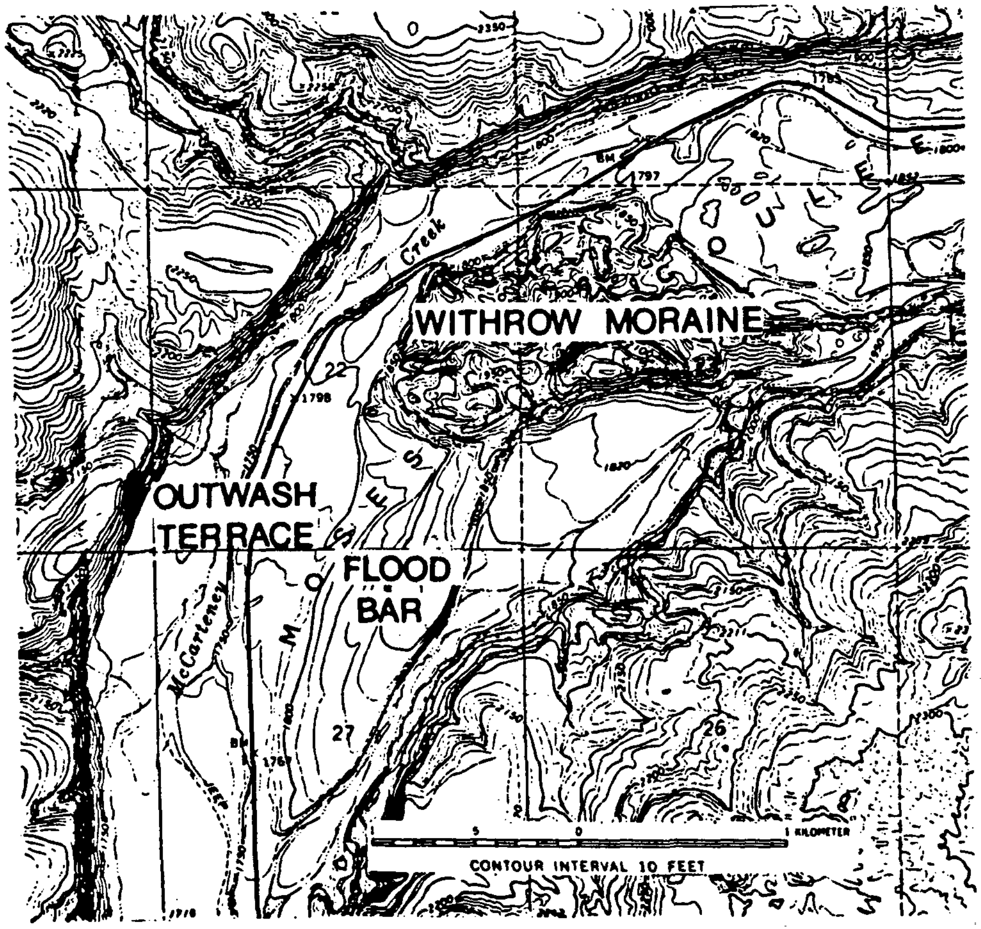

Figure 2.11. Topographic map of upper Moses Coulee, showing the physiographic relationships between the Withrow Moraine, a pendant bar of flood gravel, and an outwash terrace (from Easterbrook, 1976).

Late-Quaternary Vegetation

A pollen sequence in a mire on the Telford-Crab Creek scabland tract near Creston, Washington records the vegetation changes on the Columbia Plateau that followed the scabland flooding of that region (Hansen, 1947). Mack and others (1976) have reinterpreted the pollen sequence at this locality using improved palynological techniques. The postflood vegetation during the Glacier Peak ash fall (12,000-13,000 B.P.) and slightly earlier consisted of steppe vegetation dominated by sagebrush (*Artemisia*) occupying extensive areas of stony patterned ground. Nearby loess hills were occupied by pine forest. A warming trend is indicated to have begun about 9,000 years ago.

The Creston fen accumulated 60 cm of sediment between 9,390 ± 480 and 6,700 years B.P. and 70-100 cm between 12,000-13,000 and 9,390 ± 480 years B.P. (Mack and others, 1976; their Fig. 2). These accumulation rates range from 1.75 to 2.6 cm per 100 years. Since the mire accumulated 30-50 cm of sediment after scour by the scabland flooding and prior to Glacier Peak ash accumulation (12,000-13,000 years B.P.), that flooding must have occurred at least 13,500-16,000 years B.P. Of course, the depositional rates could have been different in this interval, so this very tenuous age estimate must be compared to other data.

Figure 2.12. Oblique aerial photograph of the outwash terrace that extends downstream from the Withrow Moraine in Moses Coulee.

Weathering and Soils

Near Vantage, Washington, wood, dated by radiocarbon at 32,700 ± 900 years B.P., was found in deposits of the last major flood (Fryxell, 1962). The wood was probably derived from a preflood, interstadial bog on a scabland surface (Bretz, 1969). Soil-stratigraphic evidence (Baker, 1973a) implies that the flooding occurred much later than this lower limiting date, but when?

On the Columbia Plateau, one indication of the age of the flooding is provided by the weathering of the basalt boulders on the surfaces of scabland flood bars. Weathering rinds on these boulders never exceed 3 mm in thickness. By contrast, rinds on similar basalt boulders deposited by the Bonneville Flood on the Snake River Plain of Idaho may be greater than 1.2 cm thick (Malde, 1968). Since the Snake River Plain has a similar climate to that of the Columbia Plateau, the Missoula Flood deposits probably considerably postdate the Bonneville Flood, which Malde (1968, p. 10) believes occurred 30,000 years ago.

The relict soils on the scabland floor gravels are poorly developed. There are only a few centimeters of near-surface oxidation. Buried cobbles and boulders lack weathering rinds. The Cca horizon dominates the soil profile, but shows only weak development. Calcium carbonate occurs as coatings on the undersides of pebbles and cobbles to a depth of about 1 m. There is no platy caliche as found on the Bonneville Flood deposits (Fig. 2.9), which were weathered in a similar soil-forming environment.

In the eastern Channeled Scabland the last major flood scoured channel ways through the entire sequence of brown loess units that comprise the Palouse Formation (Fig. 2.13). Small channels eroded into the flood deposits by postflood streams contain pale loess and loess-derived alluvium. The oldest soil profiles on postflood loess deposits show textural B horizons, with the $<2\mu$ size fraction increasing from 8% to 11% (Baker, 1973a).

Bretz' early studies concluded that the eastern Channeled Scabland (Cheney-Palouse tract) was somewhat older than the western Channeled Scabland (Grand Coulee region). The interpretation was developed by physiographic evidence (Bretz and others, 1956; Bretz, 1969). Every scabland channel way entering from the east is clearly blocked by bars that were deposited by later flooding down the Grand Coulee system. Because the weathering of the flood gravel in these bars differs little from that of bars in the Cheney-Palouse scabland, either these floods are consequences of the same outburst event or they are only separated by a few thousand years in time.

Unpublished observations of post-flood eolian silt accumulation on scabland surfaces may be relevant to this question. Bars and giant current ripples in the west scablands typically have 0.15 to 0.30 m of post-flood eolian accumulation on exposed areas and no more than 0.5 in protected swales. In contrast, the eastern scabland bars have 0.5 to 1.2 m of silt in exposed areas and up to 2.5 m in swales. This would appear to indicate greater age for the eastern scablands, but a problem remains. Silt thickness on scabland surface follows a northeasterly gradient such that

$$y = 0.46e^{0.03x}$$

where y is the thickness of postflood eolian silt (ft.) and x is the linear distance (miles) in a northeastward direction from Hanford in the Pasco Basin. Thus, silt thickness may reflect source area and wind-direction controls rather than age.

Post-flood silt is so thick at the TSCR ripple field, in the eastern scablands near Tokio Station, that it completely buries the giant current ripples (Fig. 2.14). The TSCR ripples apparent on aerial photographs are actually the silt accumulations over the former ripple troughs. The textural data indicate that this silt contains two weakly devel-

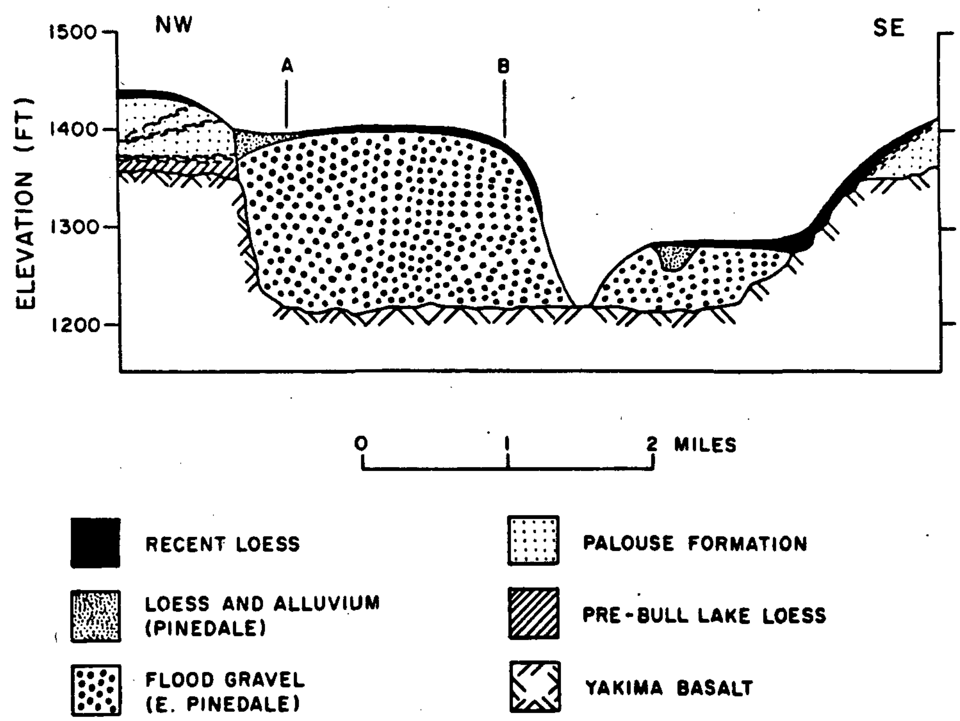

Figure 2.13. Diagramatic cross section showing stratigraphic relationships at Staircase Rapids Bar. Thickness of the loess is slightly exaggerated.

Figure 2.14. Exposure of post-flood eolian silt overlying flood gravel just north of Tokio Station. The gravel pit occurs at SW ¼ sec. 15 T. 20 N., R. 36E.

oped paleosols (Fig. 2.15). This section presents a marked contrast to flood surfaces in the western Channeled Scabland.

Tephrachronology

Eruptions of volcanoes in the Cascade Mountains showered ash over wide areas of the northwestern U. S. during the late Quaternary. Where these ashes can be distinguished, they provide a valuable series of marker beds for dating the associated sedimentary deposits and for tying together the geologic history of widely separated provinces.

Two of the most useful and widespread ash falls for work in the Channeled Scabland are those from Glacier Peak and Mt. Mazama (Fryxell, 1965). Extensive radiocarbon work dates the Mazama ash approximately 7,000 years B.P. (Kittleman, 1973). The Glacier Peak tephra is older than 12,000 and younger than 13,000 years B.P. Lemke and others (1975) designate it as younger than 12,500 years. These two ashes are easily distinguished by their refractive indices (Mazama > 1.502; G.P. < 1.502), but electron microproble analysis of the shards is necessary to avoid confusion of Glacier Peak and certain Mount St. Helens ashes. In the field the Mazama ash is usually granular, while the Glacier Peak ash forms lentils or pods with the texture of tapioca.

Most prolific of the ash sources in the Northwest was Mount St. Helens. The individual eruptions of this volcano have been dated by the detailed work of Mullineaux, Crandell, and Rubin, using the carbonaceous material in the coarse airfall material on the flanks of the volcano. The tephra occur in distinct groups, called "sets." The oldest is set S, erupted between 18,000 and 12,000 years B.P. Mullineaux and others (1975) present arguments for the most widespread members of the set S eruption (layers Sg and So) being dated at about 13,000 years B.P. Mount St. Helens tephra set J was erupted from slightly less than 12,000 to slightly more than 8,000

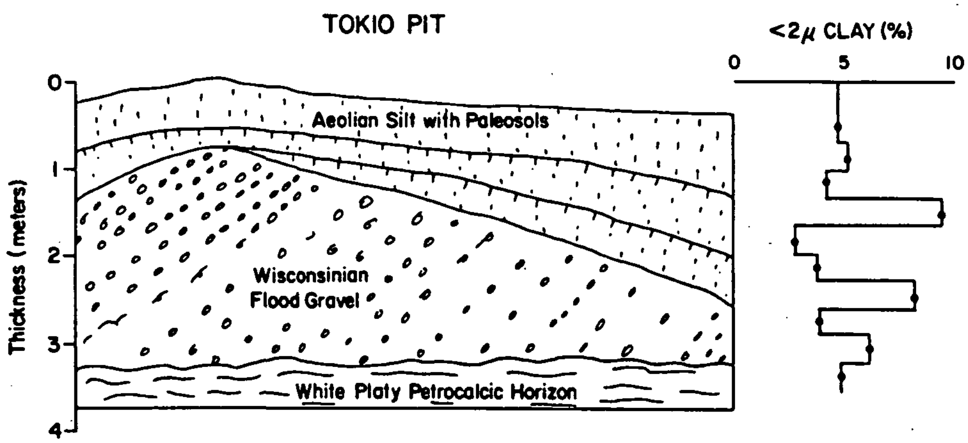

Figure 2.15. Stratigraphic relationships at the Tokio section.

years ago. Tephra set Y was extensively distributed in the Northwest between 4,000 and 3,400 years ago. Set W is the most recent, dated at about 450 years B.P. All these tephra units were carried east of Mount St. Helens and all have now been recognized in Quaternary sediments of eastern Washington (Mullineaux and others, 1975).

The Mount St. Helens set S ash occurs as "couplets" and "triplets" in fine-grained clastic sediments, often designated as "slackwater facies." Moody (1977) finds the triplet north of the Saddle Mountains at sites such as Sentinel Gap, Lind Coulee (Table 2.1), and Lynch Coulee (Fig. 2.16). South of this range only the upper and middle ash are recognized, forming a "couplet." The couplet is well exposed in "slackwater" sediments on the northern slopes of the Horse Heaven Hills, just south of Kiona, Washington. Mullineaux and others (1977) have concluded that tephra set S dates the last major scabland flood. They support the age estimate of 13,000 years B.P. for the flood with a radiocarbon date of 13,080 ± 350 on peat directly overlying the "Portland delta," a Missoula Flood deposit at Portland, Oregon.

Slackwater Sediments

The dating of the last major scabland flood is complicated by the fact that the catastrophic flood deposits occur in two facies. The main-channel facies can always be recognized as unequivocal flood deposition by its coarseness, sedimentary structures, angular boulders, broken rounds, erratics, etc. However, some of the slackwater facies are easily confused with lacustrine deposits. An unequivocal flood origin can be established by continuous tracing between main-channel areas and slackwater areas, as in the Tucannon Valley sequence (Baker, 1973a). At the Tucannon valley one can follow a complete transition from chaotically deposited boulder and cobble gravel in the main channel to rhythmically bedded silt in slackwater areas 15 km up a preflood tributary of the main channel (Baker, 1973a; p. 42-47).

Carl Gustafson (1976) divides "slackwater deposits" into 2 groups, one contemporaneous to scabland flooding and the other postflood. The postflood slackwater sediments contain articulated bivalve mollusks that lived in a lacustrine environment. At Lind Coulee, Gustafson (*in* Webster and others, 1976, p. 15) interprets deposits of the major scabland flood (18,000-20,000 years B.P.), consisting of main-channel and slackwater facies, to be unconformably overlain by younger "slackwater sediments" of lacustrine origin. This lacustrine unit contains the distinctive "triplet" of volcanic ash layers from Mt. St. Helens, which is interpreted as the Mt. St. Helens "S" set by Moody (1977).

Working in the Lower Snake River Canyon, Hammatt and others (1976) interpreted two flood events and one lacustrine phase for the

Figure 2.16. Exposure of the Mount St. Helens ash "triplet" in fine-grained "slackwater" sediment that immediately overlies scabland gravel at the mouth of Lynch Coulee, near its junction with the Columbia River at West Bar. A coarse white ash layer occurs at "B" and a fine-grained thinner layer occurs at "A." The lowest of the three layers occurs discontinuously in the granule gravel "C." The sediment containing the ash (Mt. St. Helens set S) is either time equivalent to or slightly younger than the last major catastrophic flood to affect the Columbia River Canyon northwest of the Channeled Scabland.

Table. 2.1. Correlation of some late-Quaternary deposits and events on the Columbia Plateau.

Age (Approximate years B.P.)	Lind Coulee (various sources)	Creston Mire (Mack *et al.*, 1976)	Marmes Rockshelter (Fryxell)	Okanogan Lobe (Easterbrook, 1976)
	"Holocene Loess"	Pollen Zone II	ROCKFALL —transition— LOESS	Holocene
6700—	XXXXXXXXX	MAZAMA XXXXXXXXXXXXXXX	ASH XXXXXXXXX	
	Unconformity STREAM GRAVEL	(Increase in diploxylon pine; decrease in Artemisia and Picea)	Shell Midden Loess Rockfall	
	Overbank Silt — Archeol. Site		Human Burial	
~9000—		— — — — — —	ROCKFALL	
	St. Helens "J" XXXXXXXXX Unconformity	Pollen Zone I	disconformity	Sumas Stade
12,500—	Overbank Silt — XXXXXX	GLACIER XXXXXX PEAK XXXXXX	ASH XXXXXX	Everson Interstade
	St. Helens "J" XXXXXXXXX Unconformity	(Artemisia on patterned ground; haploxylon pine on loess)		
13,000— ±500	"Slackwater Silt" — St. Helens XXXXXXXXX XXXXXXXXX XXXXXXXXX Set "S"	and		Vashon Stade — Recession
	Unconformity	?	?	
	Flood Slackwater Silt			— — —
14,000— 18,000?	Flood Gravel			Advance

XXXX Volcanic Ash

early post-glacial period. Flood gravel, which they interpret as 20,000 years B.P., is unconformably overlain by thin bedded silt containing Mt. St. Helens tephra. The silt, interpreted as lacustrine, is unconformably overlain by sandy facies interpreted as a flood slackwater deposit laid down approximately 14,000 years B.P. This upper unit forms an undulating mantle over earlier deposits and is characterized by distinctly graded bedding. All three units are cut by clastic dikes.

Sedimentological criteria must be established to distinguish slackwater facies of flood origin (Baker, 1973a) from lacustrine deposits. Work on this problem is currently in progress.

Flood Deposits Northwest of the Channeled Scabland

Waitt (1977b) presents detailed evidence for late Pleistocene catastrophic flooding coming down the Columbia River, through the region that was blocked by the Okanogan lobe until about 13,500 years B.P. Ice-rafted erratics and upvalley-dipping crossbeds in gravel show that this down-Columbia flood was as deep as 400 m at the junction with the Methow River (Waitt, 1977a). These depths require water at Coulee Dam to have had a surface elevation of 760 m. Waitt (1972b, 1977a) suggests that these relationships require hydraulic ponding of a Lake Missoula outburst at the Columbia gorge downstream from Coulee Dam. Because of the chronology of the Okanogan lobe, this flood could only have occurred about 13,500 to 13,000 years ago (Waitt, 1977b).

Flood sediment relationships at Lynch Coulee (Waitt, 1977b, p. 15) suggest that the Columbia flood was approximately contemporaneous to flooding in the Quincy Basin (presumably derived from the Grand Coulee). The Columbia flood deposits indicate transport up Lynch Coulee. These are overlain by flood deposits dipping downcoulee (having flowed from Crater Coulee). The contact shows no weathering. At Moses Coulee, however, the downcoulee deposits came first and even surged up the Colombia (Waitt, 1977b, p. 17). The Moses Coulee flood deposits are overlain by Columbia flood deposits that surged up Moses Coulee near its mouth.

Discussion

The emerging stratigraphic evidence summarized above has not yet completely resolved the number and timing of floods in the last episode of scabland flooding (19,000-13,000 years ago). The tephrachronology suggests that the last major flood affected the Columbia River, northwest of the scablands, and probably the western Channeled Scabland itself. This flood occurred just prior to or nearly coincident with the eruption of Mt. St. Helens tephra set S (approximately 13,000 years B.P.). This event probably coincided with the wastage and breakup of the Okanogan ice sheet on the Waterville Plateau.

It is also probable that another major episode of scabland flooding preceded the Okanogan breakout by several thousands years. This flood probably affected Moses Coulee, the Grand Coulee, the Telford-Crab Creek scabland, and the Cheney-Palouse. The dating of this event is less precise, but includes the following: (1) the Withrow Moranie overlying flood gravel in Moses Coulee (Fig. 2.11), (2) physiographic relationships in the Grand Coulee (Bretz, 1932a, 1969), (3) the amount of sedimentation at the Creston mire prior to the Glacier Peak ashfall, (4) the more extensive silt deposition on the Cheney-Palouse bars, (5) the younger event (13,000 years B.P.?) has bars blocking the mouths of eastern scabland distributaries (Bretz and others, 1956), and (6) stratigraphic relationships in the Snake River Canyon (Hammatt and others, 1976).

A tenative scenario that has yet to be fully tested in the field is as follows. The largest outburst of Lake Missoula occurred just prior to the Vashon glacial maximum, but while the Okanogan lobe was advancing. This flood flowed around the Okanogan lobe through the "Mansfield channels" (Hanson, 1970) and down Moses Coulee. This route was possible because (1) the advancing Okanogan lobe had just recently blocked the Columbia gorge, and (2) the upper Grand Coulee cataract had not yet receded to Coulee Dam (Bretz, 1932a). It is likely that this flood initiated the 250 m cataract in the upper Grand Coulee and that it receded to the Steamboat Rock position during the course of the "early Vashon" flooding. The same flood would also put water into the Telford-Crab Creek and

Cheney-Palouse scabland tracks. Slackwater deposits from this flood would be the lower sequence recognized by Hammat and others (1976) along the Snake River.

The Okanogan lobe advanced to its maximum position, the Withrow Moraine, very shortly after the above flood. Wastage of the Okanogan lobe then led to a second flood that also included a burst from Lake Missoula. Richmond and others (1965) refer to this as the "middle Pinedale" final flood on the Columbia River. However, that flood probably also put water over the Grand Coulee cataract head. During the course of this flood, which was mainly influencing the Columbia River in its early phases, the upper Grand Coulee cataract broke through to the Columbia gorge at Coulee Dam. This sent a final surge of water down the Grand Coulee and into the Quincy Basin. Thus, the Lynch Coulee relationships are explained by the dynamics of the last major scabland flood. That flood affected both the Columbia River and the western Channeled Scabland, but not the eastern scablands. Moses Coulee did not operate because the cataract recession of the Grand Coulee in the earlier flood had pirated the channels that fed its upstream end.

One variation of the above scenario is possible by using Waitt's (1972b, 1977a) hypothesis of hydraulic ponding at Coulee Dam. This would allow completion of the upper Grand Coulee in the earlier flood. Water would then be simultaneously flowing in the Columbia Canyon (free of the Okanogan lobe) and in the Grand Coulee.

POSTDILUVIAN STRATIGRAPHY OF THE COLUMBIA PLATEAU

Post-flood conditions in the Channeled Scabland region have been studied extensively by Roald Fryxell, Carl E. Gustafson, and other workers at the Department of Anthropology, Washington State University. Current data suggest that man entered the region as early as 12,000 years B.P. He encountered a cold dry steppe with considerable lacustrine areas. Deranged drainage conditions and high water tables were the legacy of recently wasted glaciers and catastrophic flooding.

The most intensive study of a post-flood lake in the Channeled Scabland was made by Landye (1973). Lake Bretz, whose shoreline altitude was 350 m (1160 ft.) occupied the closed depression in the lower Grand Coulee upstream from the bar at Soap Lake. The lake probably lasted a few hundred years, long enough to accumulate an abundant population of mollusks and fish. At the maximum extent, the lake extended northward about 30 km from Soap Lake, Washington, to the head of Dry Falls, near Coulee City.

The lake was probably maintained by high ground-water levels in the basalt associated with the recent deglaciation of the Waterville Plateau. The decline in water tables eventually resulted in the lake's demise. Its former basin is now occupied by a chain of somewhat saline lakes, including Deep Lake, Fall Lake, Park Lake, Blue Lake, Lake Lenore, and Soap Lake.

Fryxell (1965) recovered coarse volcanic ash from the sediments of Lake Bretz. This ash was identified as Glacier Peak ash. A radiocarbon date of 12,000 ± 30 years B. P. was obtained on the Lake Bretz mollusk shells. This analysis provides an upper limit to the age of the ash, and it also dates the lacustrine phase. However, new data on the Glacier Peak ash show that this eruption may have been the oldest of several. Mehringer and others (1977) describe two Glacier Peak ashfalls in Montana dated at about 11,250 years B.P. and separated by 10 to 25 years.

Marmes Rockshelter, located near the confluence of the Snake and Palouse rivers, produced a long history of occupation (Fryxell and others, 1968; Rice, 1972). Artifacts discovered include bone needles, bone points, stemmed and lanceolate projectile points, and bola stones. A complex burial sequence was found beneath a shell midden dated by radiocarbon at 7,550 ± 100 years B.P. The midden, in turn, is overlain by Mazama ash (6,700 years B.P.). Carl Gustafson (1972) investigated faunal remains from the site and found bones including those of Arctic fox, large elk, pronghorn antelope, deer, bison, rodents and salmonid fish. These vertebrates belonged mainly to an early postglacial steppe fauna that characterized the region prior to 7,500 years B.P. Marshall (1971) analyzed nearby floodplain sediments and interpreted precipitation and steam run-off to have been greater than now prior to 7,500 years ago. Frost polygons were found in the overbank silt deposits in which the early Marmes cultural material was discovered, but do not form in the area at present. Rockfall frequencies also showed that a cool moist climate was present during the time of the early occupation in the site (Fryxell and others, 1968). Because lacustrine sediment containing Glacier Peak ash was found immediately beneath the cultural zone, it is likely that occupation occurred over a long phase of cool, wet conditions between 12,000 and 7,500 years ago.

The Lind Coulee archeological site near Warden, Washington was the first locality in eastern Washington to yield evidence of paleo-Indian hunters. The early excavation work by Dougherty (1956) was one of the first to use the radiocarbon method, dating the earliest occupation at 8,700 years B.P. A new series of excavations at the site began in 1972. Although those studies are not yet complete, the preliminary results and correlations (Table 2.1) give an excellent late-glacial and post-glacial record. C. E. Gustafson (in Webster and others, 1976) recognizes a catastrophic flood phase unconformably overlain by lacustrine(?) silt containing the St. Helens "triplet" (set S). Inset into this slackwater silt is another silt unit. An overbank unit is set into the silts and contains two discrete ash layers including ashes from St. Helens set J and Glacier Peak (Moody, 1977). Another younger overbank sequence contains another St. Helens J ash dated at 8,700 years B.P. Mazama ash occurs in the eolian silt that blankets the overbank sequence.

CONCLUSION

Based primarily on physiographic evidence, Bretz (1969) has proposed as many as eight separate scabland floods, of which at least four encountered an ice-blocked Columbia River and spilled over the northern margin of the Columbia Plateau. Bretz' other four floods were not diverted by ice and flowed down the Columbia River valley. Bretz (1969, p. 513) believed that the earliest plateau-crossing flood occurred during the Bull Lake Glaciation. Subsequent floods then enlarged and altered segments of this established drainage.

Baker (1973a) noted that some of the physiographic relationships described by Bretz (1969) could be produced during the dynamic progression of a single flood. Only flood deposits recognized in a firm stratigraphic sequence can be considered unequivocal evidence for multiple flooding. The best stratigraphic information to date suggests that there certainly were five major floods in the general vicincity of the Channeled Scabland and perhaps six or seven. One, possibly two, floods were pre-Bull Lake and are overlain either by thick caliche or by Palouse Formation. Another flood occurred during the Palouse deposition (Bull Lake Glaciation). The third flood was the Bonneville event, restricted to the Snake River. The final flood phase, 18,000-13,000 years ago, seems to include two floods. One of these precedes the Vashon maximum, and the other occurred during or just prior to the Mount St. Helens set "S" eruptions.

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Chapter 3

Bedrock Geology of The Northern Columbia Plateau and Adjacent Areas

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ABSTRACT

The Columbia Plateau is surrounded by a complex assemblage of highly deformed Precambrian to lower Tertiary continental and oceanic rocks that reflects numerous episodes of continental accretion. The plateau itself is comprised of the Columbia River Basalt Group, a tholeiitic flood-basalt province of moderate size covering an area of about 2 x 10^5 km^2 with an estimated volume of 2 x 10^5 km^3. The Columbia River basalt, formed between about 16.5 x 10^6 years B.P. and 6 x 10^6 years B.P., is the youngest known flood basalt. More than 99 percent of the basalt was erupted during a 2.5-3 x 10^6 years interval centered about 15 x 10^6 years B.P., building a featureless plateau that sloped toward its center, reflecting concurrent subsidence and volcanism. Eruptions were infrequent between about 14 and 6 x 10^6 years B.P., allowing time for erosion and deformation between successive outpourings. The present-day courses of much of the Snake River, and parts of the Columbia River, across the plateau date from this time. Basalt produced during this waning activity is more heterogeneous chemically and isotopically than older flows, reflecting its prolonged period of volcanism. Most of the flows are thick and ponded behind natural levees. They were erupted from north-northwest-trending linear fissure systems tens of kilometers long, revealed today by dikes and relic vent areas. Eurption rates are estimated for various flows as between 1 km^3/day and 10^{-4} km^3/day per linear kilometer of active fissure, with flow rates of 5 to 15 km/hr down slopes of 1:1,000 considerd typical. Current magma production rates in Hawaii could have produced the basalt in the allotted time. No available models adequately account for the tectonic setting of the province and its relation to coeval calc-alkaline volcanism in the Cascade Range.

INTRODUCTION

The Columbia Plateau is perhaps the best area on Earth to view evidence of catastrophic flooding of truly impressive magnitude. The youngest series of floods produced the famed Channeled Scabland, which we will examine during this field conference. But an earlier episode of repeated flooding left telltale marks of a totally different but equally impressive kind; this flooding, involving lava, not water, deposited the flood basalt that built the Columbia Plateau in Miocene time. It is a remarkable coincidence that two such large yet unrelated inundations occurred in the same area and that evidence for both of them can be examined in many of the same outcrops.

This paper attempts to summarize briefly the bedrock geology of the northern Columbia Plateau and its surroundings, emphasizing the deposits of the lava floods, the Columbia River Basalt Group.

ROCKS BORDERING THE NORTHERN COLUMBIA PLATEAU

The Miocene basalt overlies and encroaches upon a diverse assemblage of Precambrian to lower Tertiary rocks (Figs. 3.1 and 3.2). The depositional, intrusive, and structural histories of these rocks are very complex and poorly known.

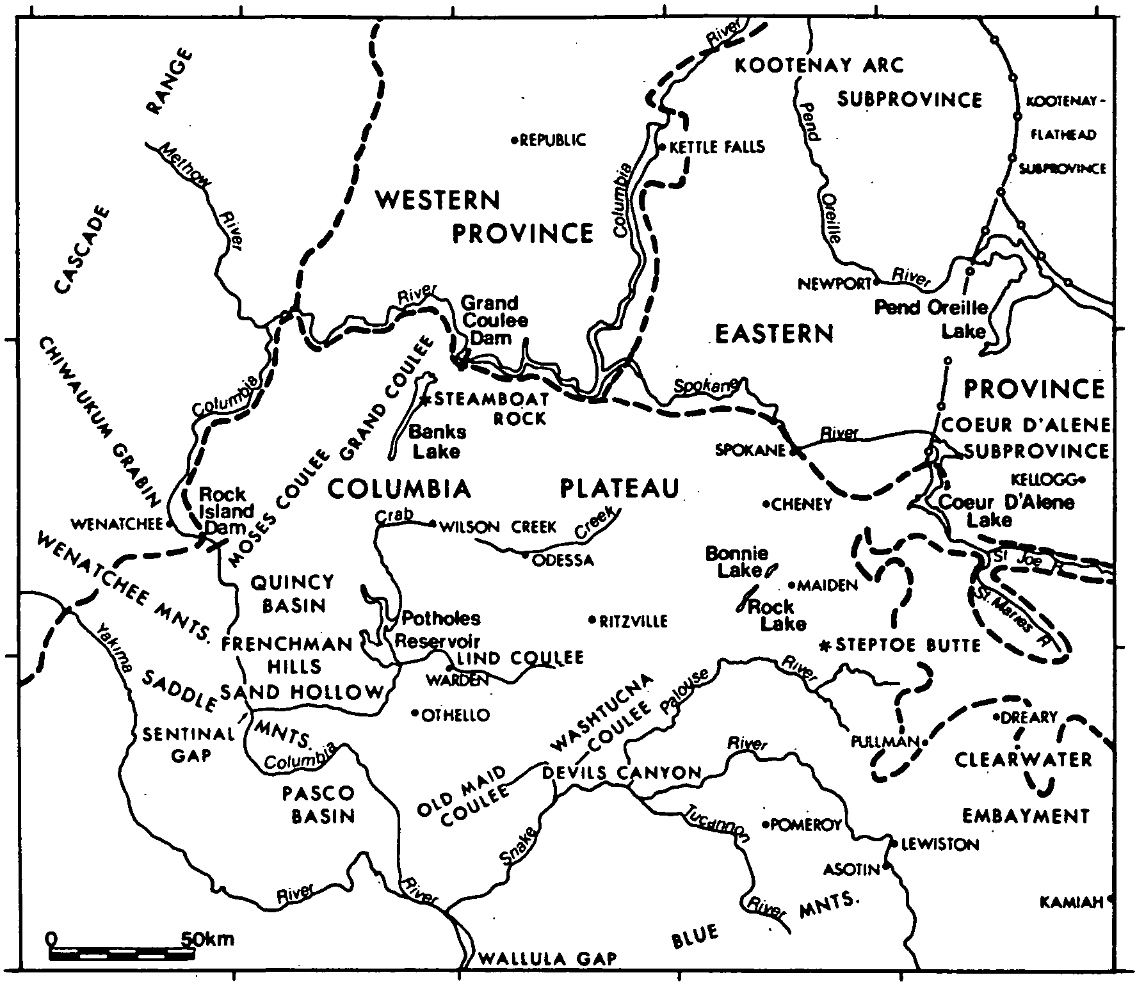

Figure 3.1. Index map showing localities and boundaries of provinces and subprovinces mentioned in text. Boundaries north of Columbia Plateau from Yates and others (1966). Other province boundaries based on Newcomb (1970).

They form highlands that rise abruptly from the margins of the plateau. No glacial floods coursed across them, except between the ice dam across the Clark Fork River and the plateau near Spokane. Glaciers descending from the highlands blocked major drainages and played an important role in scabland development by diverting water onto the plateau.

Yates and others (1966) provide an excellent summary of the pre-Miocene rocks north of the plateau and east of the Okanogan Valley. They have subdivided the region into the geologically distinct Eastern and Western Provinces, separated by the valley of the Columbia River (Fig. 3.1).

The Eastern Province is structurally divisible into three subprovinces, which Yates and others (1966) designated the Kootenay-Flathead, Coeur d'Alene, and Kootenay Arc subprovinces (Fig. 3.1). The *Kootenay-Flathead subprovince* is characterized by subparallel, broad, open folds trending northwest to north-northwest; normal faults with displacements of hundreds of meters or more parallel the folds. The major folding is of Laramide age, but the faults are somewhat

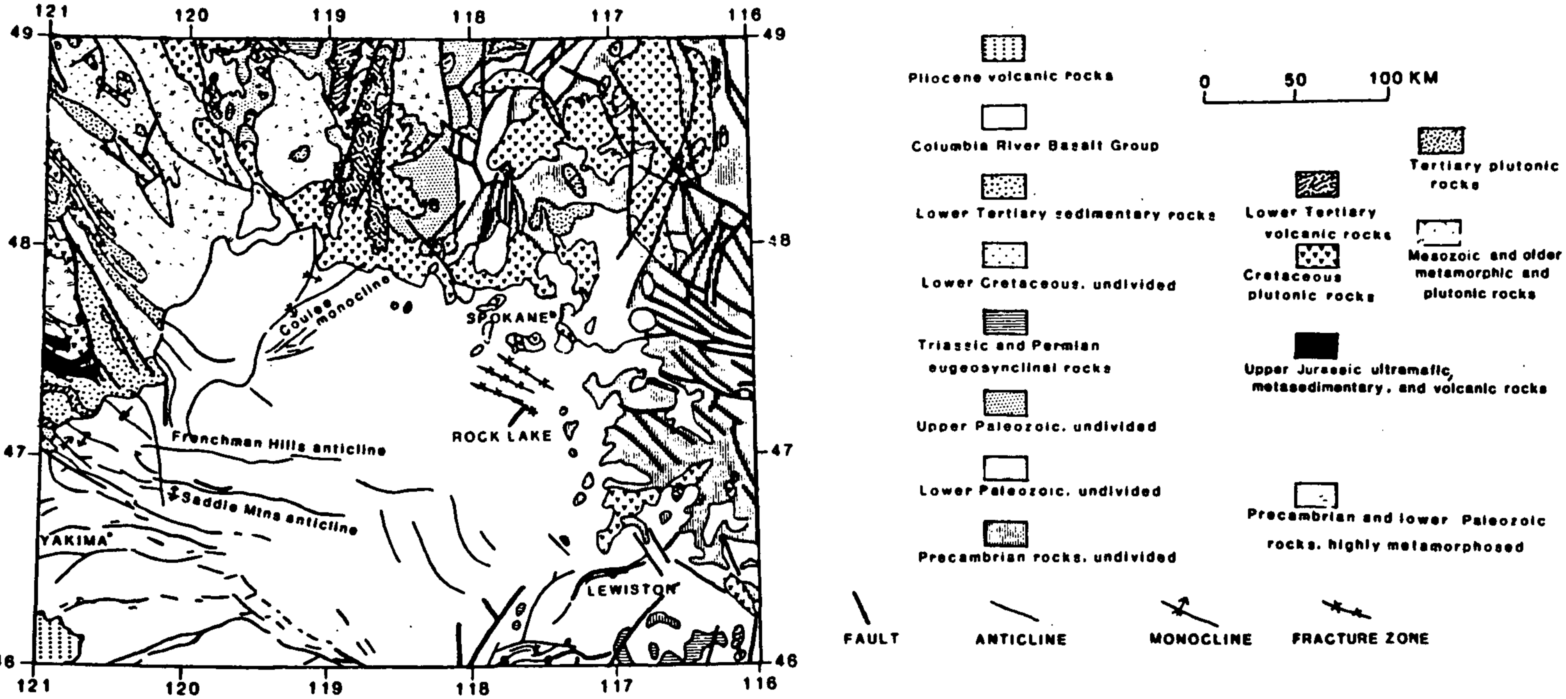

Figure 3.2. Geologic map of northern part of Columbia Plateau and surrounding area, simplified from King and Beikman (1974). Anticlinal axes, shown only on Columbia Plateau, from Newcomb (1970) with additions from Bond (1963), Griggs (1973), Swanson and others (1977), and Tabor and others (1977).

younger. The *Coeur d'Alene subprovince* contains numerous west-northwest-trending strike-slip faults (Fig. 3.2), the most notable being the Hope fault forming the northern boundary to the subprovince and the Osburn fault, both of which have right-lateral displacement of at least 25 km. This set of strike-slip faults defines a part of the Lewis and Clark line, which shows evidence of repeated movement from Precambrian to Cenozoic time (Reynolds and Kleinkopf, 1977). The subprovince contains the famed Coeur d'Alene mining district in the Wallace and Kellogg area of Idaho. The Purcell Trench separates the *Kootenay Arc subprovince* from the other two subprovinces. The Purcell Trench appears to be a major north-trending fault or fault zone separating highly metamorphosed Belt rocks on the west from mildly metamorphosed Belt rocks on the east. This contrast in metamorphic grade can be seen on either side of Rathdrum Prairie northeast of Spokane, just downstream from the site of the ice dam that impounded Lake Missoula (Griggs, 1973). Folds in the Kootenay Arc show an overall northeast trend and are of either Late Jurassic or Early Cretaceous age. The southern part of the Kootenay Arc subprovince is complex, with numerous strike-slip and thrust faults and, locally, large overturned folds cut by high-angle dip-slip faults (Miller, 1975). Deformation here has occurred intermittently since the Precambrian.

The Western Province of Yates and others (1966) contains a eugeosynclinal assemblage of graywacke, greenstone, dark shale and slate, thin chert layers, conglomerates, and pods and lenses of limestone. This assemblage represents a fundamental contrast with the miogeosynclinal character of the rocks in the Eastern Province, but the nature of the boundary between the two provinces is unknown. The age of the eugeosynclinal rocks ranges from Carboniferous to Middle Jurassic. No major unconformities reflecting orogenies during accumulation of these rocks have been recognized in the Western Province. The entire assemblage was folded and faulted, commonly along north to northwest trends, during Late Jurassic or Early Cretaceous time. The Western Province also contains high-grade metamorphic rocks of unknown age; some of these have been interpreted as gneiss domes (Fox and others, 1977; Rinehart and Fox, 1976).

Extensive batholiths intruded both the Western and Eastern (mainly the Kootenay Arc subprovince) Provinces in Late Triassic to Eocene time (Miller, 1975), with perhaps the main pulse of intrusive activity in the middle Cretaceous. These plutonic rocks vary from quartz diorite to quartz monzonite, including a distinctive two-mica quartz monzonite (Miller, 1975). Batholithic rocks comprise more than 50 percent of the exposed terrane over much of the Western Province and the Kootenay Arc subprovince. Similar plutonic rocks underlie much of the area between the Okanogan River and the Methow graben and occur south of the graben as far as the Entiat River drainage system.

Volcanic rocks, chiefly dacite to quartz latite, were erupted principally during the early and middle Eocene (Pearson and Obradovich, 1977) north of the Columbia Plateau. They are associated with clastic sedimentary rocks composed of material eroded from the uplifted batholithic rocks and their wallrocks. These Eocene rocks are best preserved north of Grand Coulee Dam along the north-northeast-trending Republic graben, which was apparently forming during the period of volcanic activity. Volanic rocks of comparable age, which may be erosional remnants of an originally extensive field, occur in Washington in the Toroda Creek graben northwest of Republic (Rinehart and Fox, 1976), along the Okanogan River, at several localities near Kettle Falls, along the Pend Oreille River northwest of Newport, and in a north-northwest belt near the mouth of the Spokane River (Pearson and Obradovich, 1977).

The eastern margin of the northern Columbia Plateau south of the Couer d'Alene subprovince is developed against the Belt Supergroup and, south of about latitude 46° 45′, the Bitterroot lobe (Armstrong, 1975) of the Idaho batholith and its highly metamorphosed wallrock. The Bitterroot lobe, just east of the area shown in Figure 3.2, consists mainly of granodiorite and granite of probable Late Cretaceous age (Hyndman and Williams, 1977). A younger, Eocene suite of granite is volumetrically subordinate. The wallrock of the Bitterroot lobe has generally been considered to be strongly metamorphosed Belt Supergroup, but Armstrong (1975) recently showed that at least the southern and western parts of the lobe intruded pre-Belt high-grade metamorphic rocks about 1,500 x 10^6 years ago. Minor volumes of lower Cenozoic volcanic rocks of intermediate and silicic compositions occur near Deary and Kamiah, Idaho, in the Clearwater embayment (Bond, 1963).

The northwestern and western parts of the northern Columbia Plateau abut against the Cascade Range. The north Cascades, north of about latitude 47° 15′, consist mainly of pre-Cretaceous gneiss and schist intruded by Mesozoic and Tertiary plutons. The parent material for the metamorphic rocks is mostly Paleozoic sedimentary and volcanic rocks (Misch, 1966, 1977). Several periods of metamorphism are known. The dominant structural trends are north to northwest and include major strike-slip, thrust, and dip-slip faults and fault zones (Misch, 1966; Yeats, 1977). Ultramafic rocks, considered by some workers to be parts of ophiolite complexes, occur in a few fault-bonded blocks. Sedimentation and lesser volcanism occurred in the Methow graben from Jurassic to early Tertiary, and in the Chiwaukum graben in early Tertiary times (Gresens and others, 1977; McKee, 1972). South of about latitude 47° 15′, the Cascades are composed dominantly of Tertiary volcanic rocks, calc-alkaline in character, intruded by shallow plutons and overlain in places by Quaternary basalt, andesite and dacite. The Columbia River Basalt Group overlies, interfingers with, and underlies volcanic rocks of Cascade derivation most of which is related to ongoing subduction of the Juan de Fuca or Farallon plate (Christiansen and Lipman, 1972; Dickinson, 1970).

COLUMBIA RIVER BASALT

Introduction

The Columbia River Basalt Group comprises a tholeiitic flood-basalt province of moderate size, covering an area of about 2 x 10^5 km^2 with an estimated volume of 2 x 10^5 km^3 of basalt (Waters, 1962). The province is commonly referred to as the Columbia Plateau, although some physiographers prefer to distinguish the Blue Mountains area in southeast Washington and northeast Oregon, large parts of which are underlain by the basalt, as a separate geomorphic province (see Baker, Ch. 2, this volume).

The group is the youngest assemblage of flood basalt known. Radiometric ages indicate that it was formed between about 16.5 and 6 x 10^6 years B.P. (Watkins and Baksi, 1974; McKee and others, 1977); early Miocene to late Miocene by modern geologic time scales (Berggren and van Couvering, 1974). More than 99 per cent of the basalt was erupted during a short 2.5-3 x 10^6 years interval centered about 15 x 10^6 years B.P. This activity built up a rather featureless plateau whose gentle regional slope was directed toward the central part of the area from all sides, reflecting concurrent subsidence and volcanism. Eruptions were far less frequent between about 14 and 6 x 10^6 years B.P., allowing time for considerable erosion and deformation between outpourings. The present courses of much of the Snake River and parts of the Columbia River were established at this time.

Relatively little erosion, other than along major drainages and in mountainous uplifted areas, has taken place in the last 6 x 10^6 years. Thus the original surface of the plateau is preserved with remarkable fidelity in some areas, especially in the eastern and northern parts of the province, where only the Channeled Scabland lends much variety to the otherwise monotonous Miocene surface covered with a relatively thin blanket of Quaternary loess.

The Columbia Plateau is a typical flood-basalt province. Flows are voluminous and cover large areas. They advanced as sheetfloods, were fed by fissure eruptions, and generally form thick cooling units composed on one or more flows, rather than compound lava flows in Walker's (1972) terminology. Cinder cones were not formed. Small spatter ramparts are present but poorly preserved. In most of these features, the province contrasts with that produced by basaltic plains volcanism, such as the Snake River Plain, discussed by Greeley (1977).

The basalt flows covered an erosional surface of considerable local relief near the margin of the plateau, 1000 m or more in many places. Some of the pre-basalt hills and ridges today still stand hundreds of meters above surrounding lava flows. The geomorphic term, steptoe, a hill of older rocks surrounded by a lava flow, was coined for Steptoe Butte 70 km south of Spokane (Fig. 3.1).

The nature of the pre-basalt surface and the rocks on which it was developed is unknown beneath the central part of the plateau. Sparse evidence suggests that a thick weathered or altered zone caps an older sequence of mafic to intermediate, lower Tertiary volcanic rocks beneath the Pasco Basin (Raymond and Tillson, 1968; Newman, 1970; Jackson, 1975).

Stratigraphy

The Columbia River Basalt Group has recently been subdivided into five formations, three of which are lumped into one subgroup corresponding to the Yakima Basalt of Waters (1961) and Swanson and others (1977). The stratigraphic nomenclature is given in Figure 3.3, with those units that occur within the area of the field conference indicated with asterisk. The Imnaha and Picture Gorge Basalts are restricted to the southwestern and southern margins of the province, respectively (Fig. 3.4), and appear to have a much smaller combined volume than the Yakima Basalt Subgroup. Papers dealing with the Imnaha Basalt are written by Hooper (1974) and Holden and Hooper (1976). The Picture Gorge Basalt is discussed by Nathan and Fruchter (1974) and Fruchter and Baldwin (1975). The formations are not discussed further in this paper.

The Yakima Basalt Subgroup underlines virtually all of the Columbia Plateau in Washington. The Channeled Scabland is developed exclusively in this unit. Representatives of its three formations, the Grande Ronde, Wanapum, and Saddle Mountains Basalts, occur in areas to be visited during the conference (Fig. 3.4). References on the Yakima Subgroup include papers by Waters (1961), Bond (1963; his "upper basalt"), Swanson (1967), and Diery and McKee (1969), on the Grande Ronde Basalt; by Mackin (1961) and Lefebvre (1970) on the Wanapum Basalt; and by Schmincke (1967a) on the Saddle Mountains Basalt.

Grande Ronde Basalt

The Grande Ronde Basalt, equivalent to the lower Yakima basalt of Wright and others (1973), is the most voluminous and areally extensive formation in the group (Fig. 3.4), covering most of the plateau with a volume of more than 150,000 km^3. Its thickness varies considerably depending on buried topography and the amount of erosion. The thickest known section exceeds 1000 m in drill holes in the Pasco Basin. Across the northern plateau, exposed thicknesses are no more than 750 m, generally much less. The formation probably contains hundreds or even a few thousand different flows.

The Grande Ronde Basalt consists mostly of non-porphyritic, fine-grained, tholeiitic basalt. Most flows carry scattered plagioclase microphenocrysts and plagioclase-clinopyroxene microphyric clots. Olivine occurs in small amounts (less than 0.5 percent) in the ground mass of all but the lowest magnesian flows.

Few flows are distinctive enough in the field to serve as stratigraphic markers, except in relatively limited areas. Criteria such as jointing habit and weathering color are tempting but unreliable for flow recognition over long distances. Chemistry is an important adjunct to correlation studies but needs to be examined in an independent framework because of repetitive compositional changes with time in the section (see paragraph on chemistry).

Flows were erupted from dikes found throughout the eastern half of the plateau (Fig. 3.4). One such dike apparently connects with its flow above the highway along the east side of Banks Lake in the Grande Coulee, 8 km southwest of the south end of Steamboat Rock (Swanson and others, 1975b).

The formation can be subdivided into four magnetostratigraphic units based on different magnetic polarities (Fig. 3.3). These four units, mappable in the field with a portable fluxgate magnetometer, are each as much as 350 m thick. They are not fine subdivisions of the formation but do provide both regional correlations and a framework for attempting chemical correlations.

The top of the Grande Ronde is generally well defined by a zone of weathering (a saprolite) or a thin sedimentary interbed separating the formation from overlying flows. For example, a poorly exposed sandstone, the Vantage Member of the Ellensburg Formation, separates the Grande Ronde and Wanapum Basalts along the Grand Coulee. The saprolite and interbed sandstone indicate a significant time break but magnetic polarities above and below the contact are similar.

Series	Group	Subgroup	Formation	Member	K-Ar age (m.y.)	Magnetic polarity
MIOCENE — UPPER MIOCENE	COLUMBIA RIVER BASALT GROUP	YAKIMA BASALT SUBGROUP	Saddle Mountains Basalt*	Lower Monumental Member	6[3]	N
				Erosional unconformity		
				Ice Harbor Member		
				basalt of Goose Island	8.5[3]	N
				basalt of Martindale	8.5[3]	R
				basalt of Basin City	8.5[3]	N
				Erosional unconformity		
				Buford Member		R
				Elephant Mountain Member*	10.5[3]	N,T
				Erosional unconformity		
				Pomona Member*	12[3]	R
MIDDLE MIOCENE				Erosional unconformity		
				Esquatzel Member		N
				Erosional unconformity		
				Weissenfels Ridge Member		
				basalt of Slippery Creek		N
				basalt of Lewiston Orchards		N
				Asotin Member		N
				Local erosional unconformity		
				Wilbur Creek Member*		N
				Umatilla Member		N
				Local erosional unconformity		
			Wanapum Basalt*	Priest Rapids Member*		R_3
				Roza Member*		T,R_3
				Frenchman Springs Member*		N_2
				Eckler Mountain Member		
				basalt of Shumaker Creek		N_2
				basalt of Dodge		N_2
				basalt of Robinette Mtn.		N_2
			Grande Ronde Basalt*		14-16.5[2]	N_2
LOWER MIOCENE						R_2
			Picture Gorge Basalt[4]	(basalt of Dayville / basalt of Monument Mtn. / basalt of Twickenham)[1]	(14.6-15.8)[1,2]	N_1
			?			R_1
			Imnaha Basalt[4]			R_1
						T
						N_0
						R_0?

[1]Information in parentheses refers to Picture Gorge Basalt

[2]Data mostly from Watkins and Baksi (1974)

[3]Data from McKee and others (1977)

[4]The Imnaha and Picture Gorge Basalts are nowhere known to be in contact. Interpretation of preliminary magnetostratigraphic data suggests that the Imnaha is older.

Figure 3.3. Stratigraphic nomenclature, age, and magnetic polarity for units within the Columbia River Basalt Group. N—normal polarity, R—reversed polarity, T—transitional polarity. Subscripts refer to magnetostratigraphic units of Swanson and others (1977). Geologic time scale from Berggren and Van Couvering (1974). * designate units in the area of the field conference.

Moreover, the Grande Ronde and overlying Wanapum Basalts are interbedded at one locality, Benjamin Gulch, 3 km south of Pomeroy, in southeast Washington (Fig. 3.1). Thus, the evidence suggests that the time break occurred within one magnetic polarity interval and probably

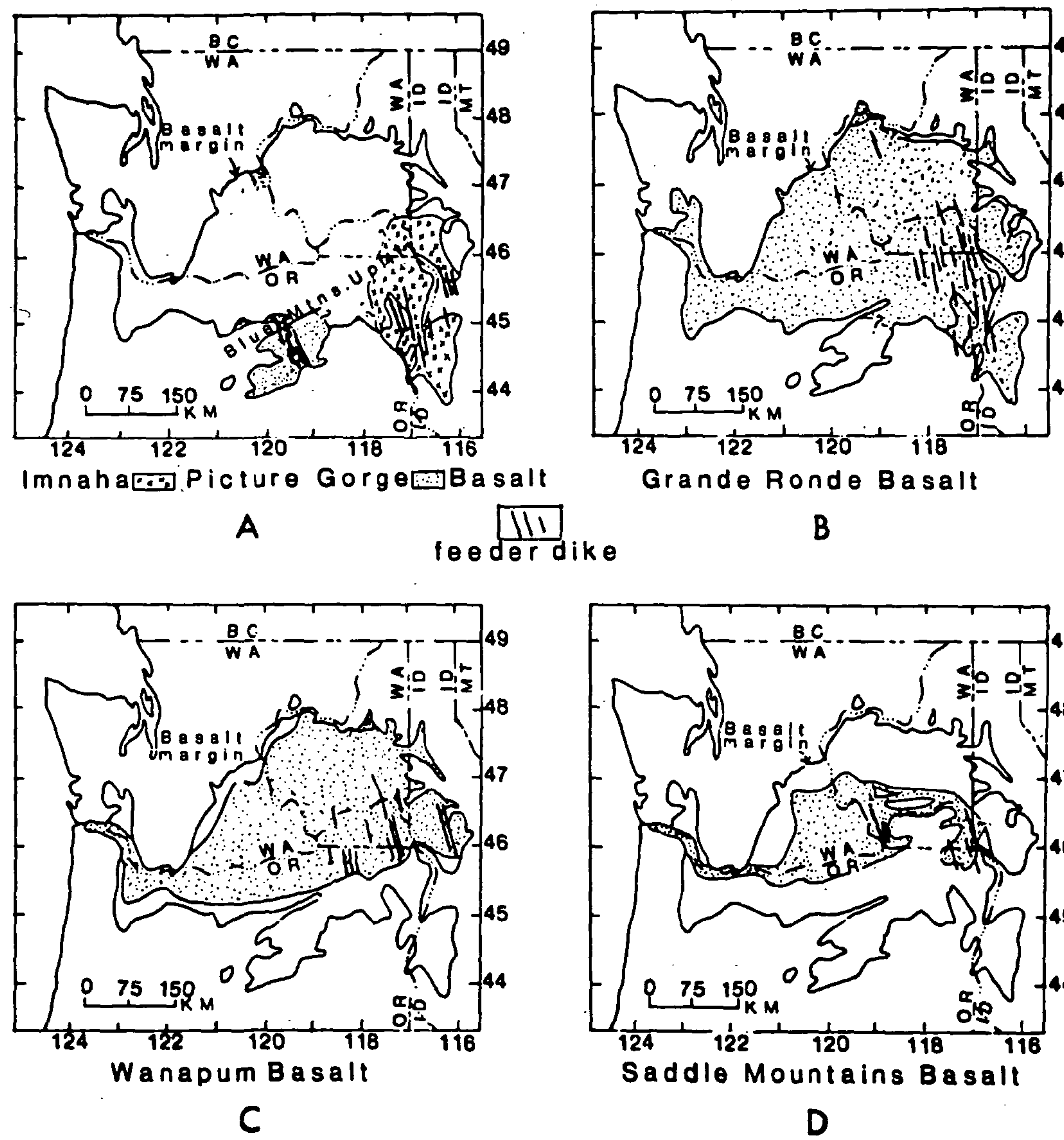

Figure 3.4. Maps showing generalized distribution and feeder dikes for the formations in the Columbia River Basalt Group. Distribution queried where uncertain.

lasted no more than a few tens of thousands of years, with eruptions continuing locally during this period.

The Grande Ronde crops out only in the more deeply eroded parts of the plateau within the area of the field trip. The lower flows in the Spokane area and up the St. Joe and St. Maries Rivers belong to this formation (Griggs, 1976), although previous workers had considered them younger (Pardee and Bryan, 1926; Dort, 1967; Bishop, 1969). The formation crops out along the lower parts of some of the deepest scabland channels southeast of Spokane, such as the channel containing Bonnie and Rock Lakes. The lowest flows all along the Grand Coulee belong to the formation, as do those along Moses Coulee. The valley of the Columbia River is eroded into the Grande Ronde; particularly good sections are exposed in the water gaps across the Saddle Mountains and Frenchman Hills. The formation is well exposed along the Snake River canyon upstream from Devils Canyon.

Wanapum Basalt

The Wanapum Basalt, approximately equivalent to the middle Yakima basalt of Wright and others (1973), covers a wide part of the Columbia Plateau (Fig. 3.4) including most of the area to be seen during this field conference. It is much less voluminous than the Grande Ronde Basalt, probably containing less than 10,000 km^3 of lava flows. On a local scale, the Wanapum overlies the Grande Ronde conformably or with very slight, local erosional disconformity, except for the interbedded relation in Benjamin Gulch already described. On a regional scale, however, the Wanapum overlies progressively older basalt from the center toward the margins of the plateau; this is especially apparent in southeast Washington (Swanson and others, 1977). We interpret this to mean that the central plateau was subsiding during accumulation of the Grande Ronde, so that progressively younger flows of the Grande Ronde were confined to the deepening center of the basin.

The Wanapum is distinguished from the Grande Ronde because it consists of a sequence of generally medium-grained, olivine-bearing flows that commonly contain a few percent plagioclase phenocrysts. Most of the flows have high Fe and Ti contents, in contrast to those in the Grande Ronde.

Feeder dikes for flows in the Wanapum Basalt occur widely in the eastern half of the plateau (Fig. 3.4), although feeders for an individual flow or a sequence of related flows are rather localized (Fig. 3.5).

The formation is divided into four members on the basis of petrography and magnetic polarity (Fig. 3.5). The oldest, the Eckler Mountain Member, does not crop out in the northern plateau except for a possible occurrence near St. Maries, Idaho (Griggs, 1976). The Frenchman Springs, Roza, and Priest Rapids Members are considerably more extensive (Fig. 3.5) and are described below.

Frenchman Springs Member—This is the most extensive member of the Wanapum (Fig. 3.5) and probably contains 3,000 to 5,000 km^3 of basalt. As many as ten flows, generally three to six, of normal magnetic polarity occur in any one section. Mackin (1961) named the three flows in the Quincy Basin, from bottom to top, the Ginko, Sand Hollow, and Sentinel Gap.

Most flows in the Frenchman Springs contain scattered glomerophyric clots of plagioclase a centimeter or more across. The clots are generally very unevenly distributed through a flow and may be hard to find. Many flows were erupted from dikes north and south of Walla Walla (Fig. 3.5).

The Frenchman Springs Member apparently never covered the northeast part of the province (Fig. 3.5). It is not found in the Spokane area or in the Cheney-Palouse scabland north of Rock Lake. The member crops out in some of the deeper channels west of Odessa, for example along Crab Creek. It occurs widely along the upper Grand Coulee, underlying the visitor overlook at Dry Falls, but pinches out to the north about 13 km southwest of the south end of Steamboat Rock. The Frenchman Springs borders the Quincy Basin in the north, south, and west (Grolier and Bingham, 1971) and is well exposed in Lind and Washtucna Coulees, Devils Canyon, and along the Snake River.

Roza Member—The flows within the Roza Member, which overlies the Frenchman Springs, are probably the best known and certainly some of the most widespread of all flows on the plateau

(Fig. 3.5). The following is taken from Swanson and others (1975b).

"The Roza is characterized by abundant plagioclase phenocrysts, averaging more than 5 mm in length; the phenocrysts are distributed quite uniformly, generally about 10 per 150 cm^2 (Lefebvre, 1970), and are mostly single crystals. This texture serves to distinguish the Roza Member from almost every other unit and, when stratigraphic relations are considered, is virtually diagnostic . . . [Work by many geologists shows that] the Roza Member originally covered an area of at least 40,000 km^2 and contained a volume of lava greater than 1500 km^3.

The member typically consists of one or two thick flows . . . [totalling about 50 m thickness] . . . but in some places contains three or more. Thin flow units are abundant only near vent areas. Successive flows can be recognized by the

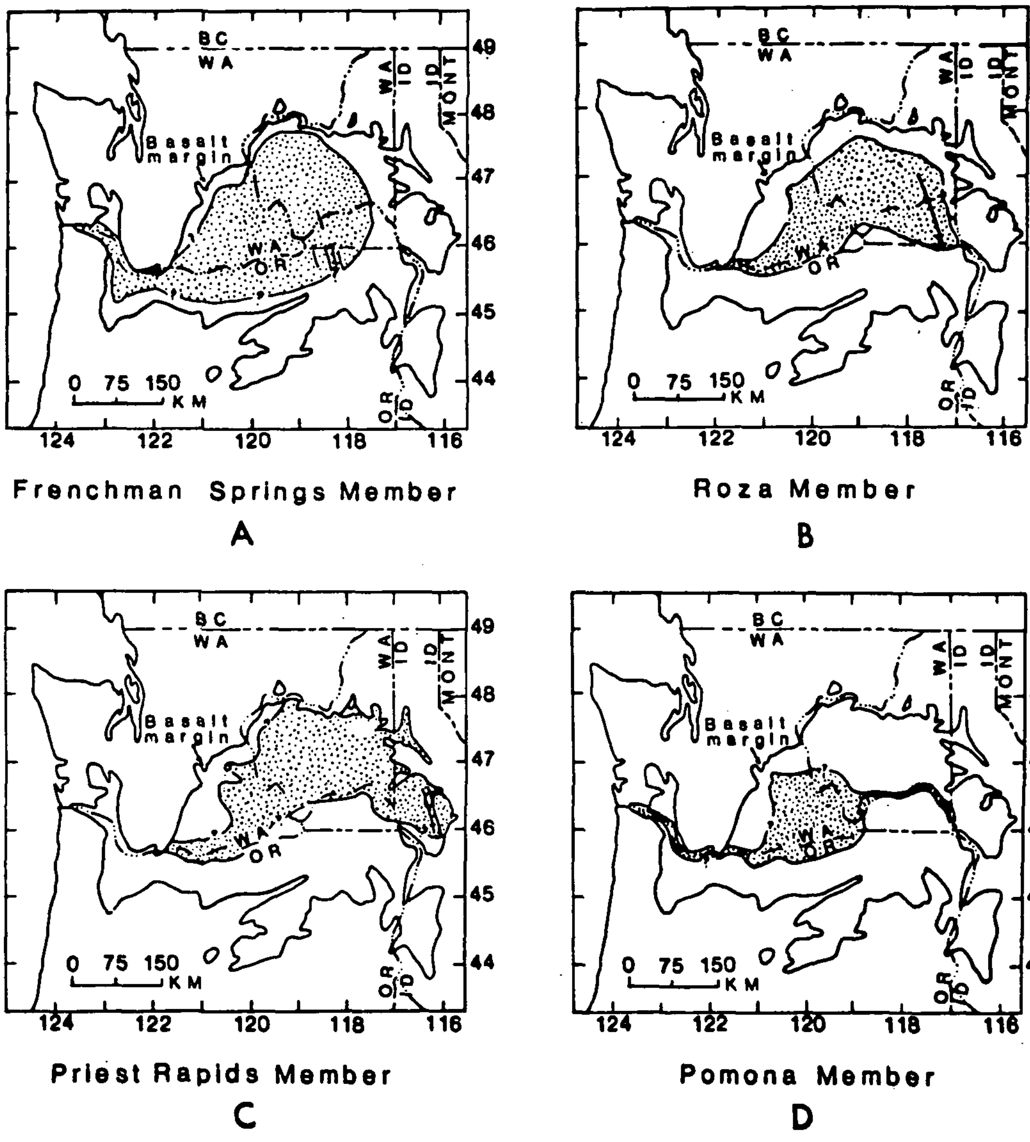

Figure 3.5. Maps showing generalized distribution and feeder dikes for the Frenchman Springs, Roza, and Priest Rapids Members of the Wanapum Basalt and the Pomona Member of the Saddle Mountains Basalt.

presence of intervening vesicular and ropy zones on the upper surface of the lower flow or by small differences in the number and size of plagioclase phenocrysts. Multiple flows may have been overlooked in some places where only one flow has been reported, perhaps because so little time elapsed between successive eruptions that cooling and solidification tended to obscure flow contacts. In general, such terms as simple, compound, single-flow, or multiple-flow cooling units, originally applied to ash-flow tuffs (Smith, 1960a, b), are thought to be equally applicable to the Roza Member and to flows of the . . . [Columbia River Basalt Group] . . . In this scheme, the Roza Member is commonly a simple cooling unit, but in places consists of a multiple-flow, compound cooling unit or several different cooling units.

A variety of evidence suggests that the Roza Member was erupted during a short period of time, perhaps of the order of a few hundred . . . (to a few thousand) . . . years for the entire member and a matter of days for single flows and cooling units. Interflow sediments, except locally derived tephra, have not been found, even though epiclastic sediments of extra-plateau derivation, as well as biogenic sediments, were deposited both before and after Roza time in many places on the Columbia Plateau . . . Evidence of significant erosion between successive Roza flows has nowhere been found, and delicate surface textures are commonly preserved. The transitional magnetic polarity (Reitman, 1966) of the Roza Member is further permissive evidence of its rapid accumulation. Such a polarity is apparently produced during a reversal in the Earth's magnetic field, which is thought to take place rapidly, possibly within 10^3 yrs (Rikitake, 1972) . . . [Recently, a magnetically reversed flow and dike of Roza affinity have been found in southeast Washington; they are interpreted to represent a late stage of Roza volcanism after the magnetic field had completed its change from normal (Frenchman Springs Member) through transitional (Roza Member) to reversed (Roza and Priest Rapids Members) polarity (Choiniere, 1977)]."

We have found numerous vent and near-vent areas and feeder dikes that define a linear, north-northwest trending, vent system in the eastern plateau (Fig. 3.5). The zone is probably less than 5 km wide and at least 130 km long. The northernmost known vents occur in the Cheney-Palouse scabland northwest of Winona (Bingham, 1970). The vent and near-vent areas are defined by accumulations of airfall spatter, some welded or agglutinated, and pumice of Roza lithology. Remnants of spatter cones and ramparts are recognizable in places; some have been bulldozed or burrowed into by a younger flow of the member. Thick mounds of platy, relatively dense flows form low hills at presumed vents in the southern part of the vent system and are interpreted to be some of the latest cooled and/or degassed material to be erupted. This is the material with reversed magnetic polarity.

The Roza, like the Frenchman Springs Member, failed to advance into the Spokane area (Fig. 3.5); its northeast terminus can be observed in one place just south of Bonnie Lake in the Cheney-Palouse scabland. The member is widespread elsewhere in the northern plateau, however. It is well exposed along Crab Creek in the Odessa-Wilson area, where it hosts peculiar ring structures, 50 m to 500 m in diameter, interpreted by McKee and Stradling (1970) as sag flowouts, formed by foundering of crust on a partly solidified flow, and by Hodges (1976) as reflecting interaction of water and lava. The Roza is prominent along the Grand Coulee, where it extends farther north than the Frenchman Springs Member and rests directly on the Grande Ronde Basalt. It occurs throughout the Quincy Basin (Grolier and Bingham, 1971); in the Potholes Reservoir area, where much of the dam consists of riprap from the Roza; along the walls of the Columbia Valley; and throughout the southern part of the Cheney-Palouse scabland.

Priest Rapids Member—The Priest Rapids Member overlies the Roza and contains the youngest flows throughout most of the scabland country (Fig. 3.5). It typically consists of one or two flows totalling 30-50 m thick. The entire member may have a volume of 2,000-3,000 km^3. Flows in the Channeled Scabland typically contain small olivine and scattered small plagioclase phenocrysts. All flows have reversed magnetic polarity.

Flows of two different compositions occur in this member throughout the scabland country. One has high Fe and Ti (Rosalia chemical type of Table 3.1). The Lolo-type flow(s) appear to be younger, but considerably more mapping is needed to definitely establish the age relations. Sources for the Lolo-type flows are known in western Idaho (Fig. 3.5); sources for Rosalia-type flows have not been found.

Table 3.1. **Chemical types of basalt discussed in text and present in Channeled Scabland (oxides in percent; trace elements in ppm)**

	Grande Ronde Basalt		Wanapum Basalt			Saddle Mountains Basalt		
	High-Mg Grande Ronde (one flow)	Low-Mg Grande Ronde (one flow)	Roza	Priest Rapids Member: Lolo	Priest Rapids Member: Rosalia	Wilbur Creek	Pomona	Elephant Mountain
SiO_2	53.56	55.80	50.97	49.96	50.09	54.41	51.77	50.89
Al_2O_3	14.58	14.00	14.01	14.32	13.63	14.51	14.85	13.52
FeO [1]	11.10	11.74	13.85	13.71	14.98	11.07	10.52	14.66
MgO	5.23	3.35	4.37	5.15	4.28	4.51	6.98	4.26
CaO	9.08	6.86	8.42	8.83	8.27	8.32	10.63	8.31
Na_2O	2.90	3.13	2.71	2.56	2.66	2.69	2.35	2.45
K_2O	1.15	1.98	1.21	1.03	1.15	1.77	.64	1.23
TiO_2	1.70	2.26	3.11	3.13	3.54	1.95	1.62	3.51
P_2O_5	.26	.43	.67	.78	.80	.56	.25	.58
MnO	.19	.19	.24	.20	.21	.21	.18	.20
Cr	100.2	12.8	54.5	97.0	15.5	36 [2]	110 [2]	20 [2]
Cs	0.8	1.40	1.0	0.7	.95			
Hf	3.75	5.10	4.3	4.45	5.35	7.1	3.6	6.5
Rb	27.5	45.0	25.0	21.0	28.0			
Ta	.74	1.0	.99	1.18	1.11			
Th	3.50	6.30	3.8	3.55	4.05	6.6	2.6	6.0
Zn	133.0	145.0	171	214	220			
Sc	37.1	31.2	35.45	35.5	36.7	27.0	35	31
La	18.5	28.5	27.0	29.5	33.5	41.0	17	34
Ce	38.5	58.5	54.5	59.5	69.5			
Sm	5.40	7.80	7.8	8.85	9.1	7.9	4.8	9.6
Eu	1.69	2.19	2.34	2.68	2.81	2.4	1.5	2.6
Yb	2.6	3.65	3.20	3.50	4.25			
Lu	.51	.63	.60	.65	.73	0.69	.42	.76

[1] $FeO + 0.9\ Fe_2O_3$.
[2] Trace element data from J. S. Fruchter (written commun., 1977).

The rimrock flows near Spokane, and the younger flows up the St. Joe and St. Maries Valleys, are of the Rosalia chemical type. Similar flows extend far southwest into the Cheney-Palouse scabland and west at least as far as the Odessa area. Farther south, similar high Fe and Ti flows occur near Othello. The high Mg, Lolo-type flows crop out in the lower Cheney-Palouse scabland, for example in the Devils Canyon-HU Ranch area, and within the Quincy Basin. Flows *of unknown composition occur in the member all* along Grand Coulee. The Priest Rapids is commonly hidden beneath loess along the margins of many scabland channels, with the underlying Roza Member the most prominent flow displayed.

Saddle Mountains Basalt

The Saddle Mountains Basalt, approximately equivalent to the upper Yakima basalt of Wright and others (1973), is the youngest formation in the Yakima Basalt Subgroup. It contains flows of diverse chemistry, petrography, age, and paleomagnetic polarity (Fig. 3.3). It was erupted between about $13.5 \pm 0.5 \times 10^6$ and 6×10^6 years ago (McKee and others, 1977; Atlantic Richfield Hanford Company, 1976), during a period of waning volcanism, accelerated folding, canyon cutting, and development of thick but local sedimentary deposits between flows. The Saddle *Mountains Basalt has a volume of only about* 2000 km^3, less than 1 percent of the total volume of the Columbia River Basalt Group, yet contains by far the greatest chemical diversity, including major and trace element and isotopic abundances, of any formation in the group.

The Saddle Mountains Basalt has been subdivided into 10 members based on petrographic, magnetic, and chemical characteristics. These 10 members are listed in Figure 3.3, together with several informally named subunits. Four of the members, the Wilbur Creek, Asotin, Weissenfels Ridge, and Buford, occur principally in extreme southeast Washington and adjacent Idaho and Oregon. The Wilbur Creek and Asotin, however, also are found as intracanyon flows toward the center of the Columbia Plateau (Swanson and others, 1977), probably channeled westward by ancestral valleys of the Palouse River (Figs. 3.1 and 3.4). Two other members, the Esquatzel and Lower Monumental, occur chiefly as intracanyon flows along the ancestral Snake River. They therefore cover only a very small total area but extend for tens of kilometers along the canyon. The Esquatzel spilled from the mouth of the old canyon and covered part of the Pasco Basin. The other four members, the Umatilla, Pomona, Elephant Mountain, and Ice Harbor, cover relatively wide areas and also occur, at least locally, as canyon fills.

Of the 10 members of the Saddle Mountains Basalt, only the Wilbur Creek, Pomona and Elephant Mountain members are likely to be encountered during the conference and are described below.

Wilbur Creek Member.—The Wilbur Creek is fine-grained and has sparse phenocrysts of plagioclase less than 5 mm across. It was apparently erupted in the eastern part of the province, where it is most widespread and flowed westward down an ancestral valley system. The member occurs as a remnant of a valley-filling flow in the Warden-Othello area, where it forms a prominent, sinuous ridge owing to erosional inversion of topography.

Pomona Member.—The Pomona Member consists of one principal cooling unit characterized by small, commonly wedge-shaped phenocrysts of plagioclase (generally less than 5 mm long), together with scattered clinopyroxene and olivine. Some plagioclase phenocrysts are riddled with clinopyroxene inclusions.

Schmincke (1967a) showed that the Pomona occurs as a sheetlike flow throughout much of south-central Washington (Fig. 3.5), and our work indicates its presence in nearly 50 remnants of an intracanyon flow along an ancestral Snake River canyon from Asotin (Fig. 3.1) to the central Columbia Plateau (Swanson and Wright, 1976; Swanson and others, 1977). Figure 3.6 presents an example of an intracanyon flow remnant. The flow presumably advanced down the canyon from a source in western Idaho, emptying from the mouth of the canyon in lower Old Maid Coulee into a broad basin across which the flow moved as a sheetflood.

A peperite, a mixture of sediment and chilled, fragmented lava, is commonly developed where the Pomona plowed into unconsolidated vitric ash near the margin of the flow (Schmincke, 1967b). Good examples of this relation occur along Crab Creek northwest of Othello, about the only place where the member crops out in the area of the field conference.

The Pomona averages about 30 m thick outside of the ancient canyon. Its maximum preserved thickness in the canyon is 110 m near the mouth of the Tucannon River. It covers more than 18,000 km^2 and has a volume of more than 600 km^3. The flow advanced nearly to the Pacific Ocean along the Columbia River Valley, more than 500 km from its suspected but unproven source area in western Idaho (Fig. 3.5). It is truly one of the monumental basalt flows on Earth.

Elephant Mountain Member.—The Elephant Mountain Member overlies the Pomona and consists of several nearly non-porphyritic and generally fine-grained flows that are physically nondistinctive but chemically recognizable (Table 3.1). The member occupies more or less the area covered by the Pomona, although it did not advance west of the Cascades. Flows advanced down the ancestral Snake River Canyon from sources in southeast Washington and adjacent Idaho and Oregon (Swanson and others, 1975a). Its thickness averages about 30 m, in intracanyon remnants reaching 150 m, and its volume is 200-250 km^3. The member is the youngest flow exposed south and west of Othello.

Physical characteristics of flows

Flows within the Grande Ronde, Wanapum, and Saddle Mountains Basalts range from a few tens of centimeters to more than 100 m thick, averaging 30-40 m. The thick flows generally record ponding in pre-basalt valleys, in structurally controlled basins that developed during volcanism, or in narrow canyons eroded into older flows; such intracanyon flows are common only in the Saddle Mountains Basalt. Even the thinner flows generally show evidence of being ponded. This evidence con-

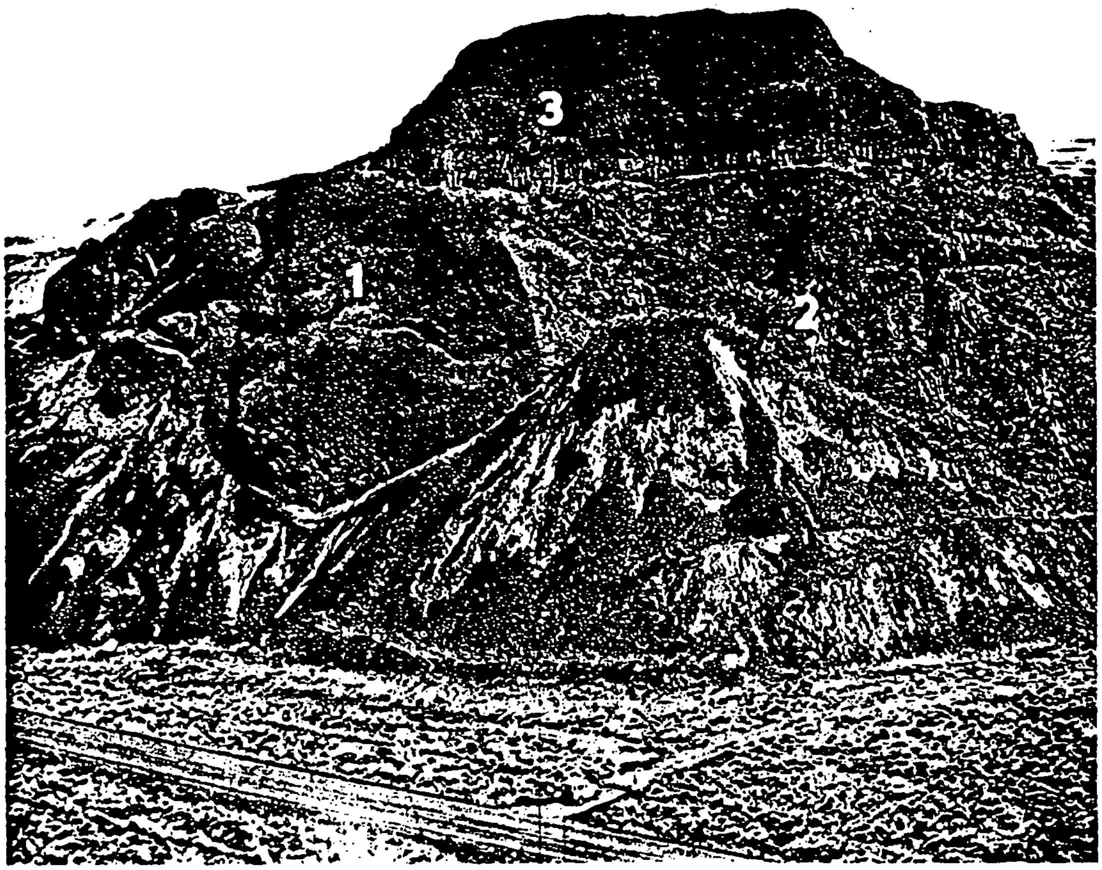

Figure 3.6. Photograph showing cross section of an ancient canyon of the Snake River and three flows of Saddle Mountains Basalt that were successively deposited in it. 1 designates the oldest intra-canyon flow which partly filled a canyon eroded into the Frenchman Springs Member, which is visible in the background at the extreme left edge of the photo. This oldest intracanyon flow was itself incised, and flow 2, the Pomona Member, makes up the columnar basalt against the steep wall of that paleocanyon. The Pomona (2) and flow 1 were then beveled by erosion to a nearly flat surface, and flow 3, in the Elephant Mountain Member, moved down-canyon (toward observer) and covered both older flows. Exposure is on the east wall of Devils Canyon and is about 100 m high.

sists of the columnar-jointed nature of the basalt (Fig. 3.7). Such columns can apparently form only under static cooling conditions; their development therefore implies that the lava had ponded. What impounded the lava can rarely be determined. Natural levees several meters high have been observed in places and probably account for most of the ponding. Elsewhere, flows could have pinched out against opposed topographic slopes.

Flows that cooled under stagnant conditions contracted and developed a characteristic jointing habit, shown in idealized form in Figure 3.7. The terms colonnade and entablature were borrowed from classical architectural usage by Tomkeieff (1940). Columns in the colonnade are from 10 cm to 5 m in diameter, averaging about 1 m, and can be as long as 50-75 m although generally 5-10 m. Most are straight, but curved columns are rather common and generally unexplainable in terms of simple cooling models. Columns in the colonnade are commonly subdivided into prismatic blocks by cross or blocky joints, and platy joints may form in the coarsest-grained part of a flow.

The colonnade-entablature contact is relatively sharp, the change commonly taking place within 1-2 cm. The contact is traceable in many places for several kilometers before other complexities obscure it. The glass content of the groundmass increases abruptly from the colonnade to the entablature for an unknown reason (Swanson, 1967). The entablature consists of columns of smaller diameter, generally less than 25 cm, and less consistent orientation than those in the colonnade. Columns in many entablatures are bundled into fan-, synclinal-, tent-, or other unusually shaped arrangements. Most columns in an entablature are highly segmented by irregular cross joints, so that the columns can be readily broken in fist-size pieces. The entablature generally comprises about 70 percent of the thickness of a flow but can make up 100 percent (one example that we know of) to zero percent. The upper part of the entablature is scoriaceous and commonly merges into a zone of short, wide, generally poorly defined columns that some workers call the upper colonnade. A rubbly, clinkery zone occurs at the top of some flows locally; such a zone is, in our experience, much more common near vent areas than elsewhere.

Idealized jointing patterns can be satisfactorily explained by existing theory for the cooling of bodies of igneous rock (Jaeger, 1961), but such patterns are seldom found in nature. Acceptable thermomechanical explanations for the typically complex jointing patterns, particularly in the entablature, are not available despite considerable descriptive information (Tomkeieff, 1940; Waters, 1960; Mackin, 1961; Spry, 1962; Swanson, 1967; Schmincke, 1967a). Problems such as mutual interference of columns growing inward from irregular contacts, ponding of water on a flow surface and percolation down joint planes during solidification, the influence of chemical composition on tensile strengths and heat conduction, and inadequate knowledge of rock mechanics under high temperature-low pressure conditions are some of the difficulties that plague attempts at analysis of natural jointing habits.

Some flows have a tiered appearance defined principally by alternating layers of vesicular and relatively nonvesicular rock rather than by joints. These layers may record separate gushes or thin flows that piled up and solidified as a single compound cooling unit.

Many flows entered water and formed pillows. Recent studies (Jones, 1968; Moore, 1975) have demonstrated conclusively that pillows are nothing more than the subaqueous equivalent of pahoehoe toes. Many of the pillowed flows occur near the margin of the plateau at the time of eruption, apparently because lakes resulted from flows ponding rivers draining marginal highlands. Other pillowed flows are much more extensive, perhaps signifying entry into shallow lakes standing on the plateau surface. An example of such a flow is one of the Priest Rapid flows, which is pillowed throughout an area of tens of square kilometers in the Cheney Palouse scabland, although the

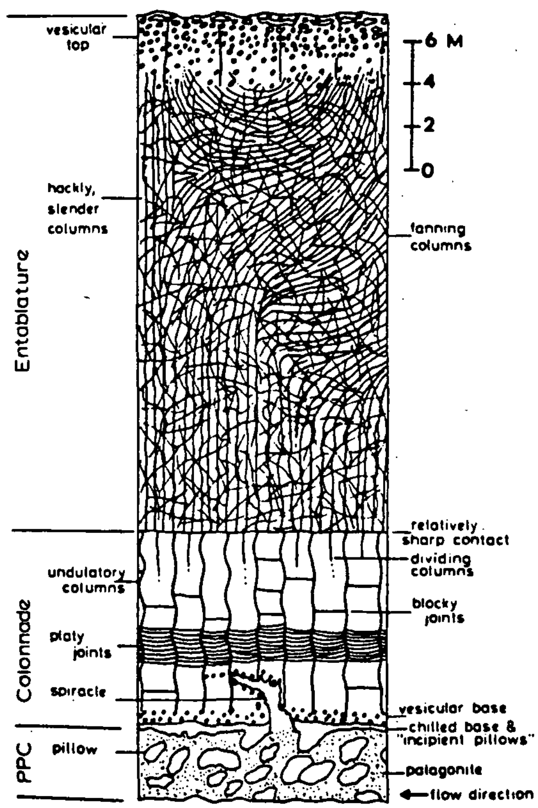

Figure 3.7. Cross section of typical flow in the Yakima Basalt Subgroup showing, in idealized form, jointing patterns and other structures. PPC, pillow-palagonite (hyaloclastite) complex, present at base of flows that entered water (from Swanson, 1967).

pillows were stripped away by the Pleistocene floods in many places and are not very apparent.

In places, lava deltas (Fuller, 1931; Moore and others, 1973) formed as lava poured into shallow lakes and ponded streams. The direction of dip of foreset "bedding," defined by elongate pillows and thin sheet flows, in the lava deltas indicates the local flow direction of the lava. Some classic lava deltas can be seen near Malden south of Spokane (Griggs, 1976), near the mouth of Moses Coulee (Fuller, 1931), and at the mouth of Sand Hollow south of Vantage.

Other criteria for defining flow directions include inclined pipe vesicles, which plunge upcurrent, and bent spiracles (Fig. 3.7), formed by steam blasts beneath a flow, which tail out downcurrent. Flow directional data for basalts must be treated in the same way as those for current-produced structures in sedimentary rocks—carefully. A few data in a small area show the local direction but say little about regional patterns. Nonetheless, careful studies by Schmincke (1967c) and ongoing work by others are succeeding in defining patterns of lava advance within the plateau.

Basalt flows burrowed into deposits of unconsolidated sediments, forming peperites and even sills, in many places on the plateau (Schmincke, 1967b). Excellent examples of such flows, called invasive flows (Byerly and Swanson, 1978), are exposed in the diatomite mines in the Quincy Basin (Roza Member), near Rock Island Dam southeast of Wenatchee (Fig. 3.8; the Hammond sill in the Grand Ronde Basalt; Hoyt, 1961), and in many areas near Spokane where flows burrowed into sediments of the Latah Formation. In our experience, nearly half of the examples of basalt-sediment contacts on the plateau record invasion by the basalt and inversion of the stratigraphy.

Invasive flows are particularly common in the Wenatchee-Ellensburg area, where they have been studied by Byerly and Swanson (1978). Some of these 5-120 m thick flows cover hundreds of square kilometers. Their tops have thick glass selvedges, generally are nearly planar, and contain few vesicles. Locally, thin dikes and sills sprout from the top and intrude the host sedimentary rock. All sill-like bodies intrude 3 to 20 m thick sedimentary deposits; none cuts an

Figure 3.8. Tongues of basalt invading silt and fine sand (now lithified). Satellite sill associated with the Hammond invasive flow near Rock Island Dam.

older flow, nor have any feeder dikes been found. Exposures show lateral gradations over hundreds to thousands of meters from surface flows through pillow-hyaloclastite complexes and peperites into invasive flows. Flow directional data show that the lava flowed toward the sill-like bodies. Microprobe analysis of glassy selvedges confirms correlations made across these facies changes and shows that the invasive flows fit perfectly into the chemical stratigraphy established for the basalt in nearby areas. Likewise, magnetic stratigraphy shows no anomalies; all flows and interlayered invasive flows in the lower and upper parts of the section have normal polarity, all in the middle part have reversed polarity. Each invasive flow apparently formed prior to the next higher flow, as lava advanced into a low area and burrowed into unconsolidated sediments.

Chemistry

Study of the chemistry of the Columbia River Basalt Group has gone hand in hand with field work. Major element chemistry has been obtained for hundreds of samples collected in the field (Brock and Grolier, 1973; Hooper and others, 1976; T. L. Wright and D. A. Swanson, unpub. data, 1978), and trace element chemistry has been obtained for selected sections and also for representative samples of all parts of the stratigraphic section (Osawa and Goles, 1970; Nathan and Fruchter, 1974; T. L. Wright and D. A. Swanson, unpub. data, 1978). Sr-isotope data have been published by McDougall (1976), and studies are being completed by D. O. Nelson (unpub. data, 1978; Nelson and others, 1976). Lead isotope data are available from S. E. Church (written commun., 1975).

We and others have used major element chemistry to define the variation between and within stratigraphically defined flows in the Yakima Basalt Subgroup. Flows in the Saddle Mountains Basalt all have distinct major compositions termed chemical types. Flows in the Wanapum and Grande Ronde Basalts are chemically distinguished from each other, but flows of similar chemistry are repeated within each formation. Representative chemical types from each unit are shown in Table 3.1 and representative intra-flow chemical variation is shown in Figure 3.9. Within the three formations, trace element and major element compositions correlate closely. Some flows at different stratigraphic levels in the Grande Ronde Basalt have virtually identical major and trace element compositions, indicating that magmatism was broadly cyclic. A few flows of Saddle Mountains Basalt are similar in major element abundances to flows of Wanapum or Grande Ronde Basalt but differ significantly in minor and trace element abundances; one example is shown in Table 3.2.

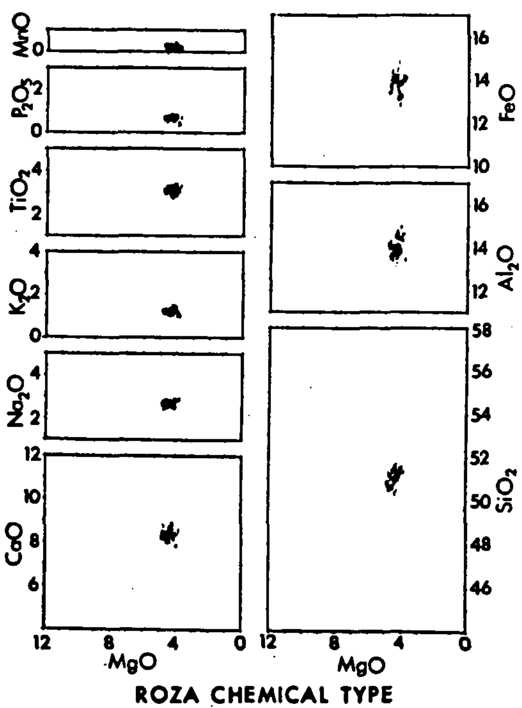

Figure 3.9. MgO-variation diagrams for the Roza Member, showing representative intraflow variation, to a large degree due to analytical uncertainty and alteration. Abundances are in weight per cents.

Table 3.2. Comparison of two chemical types with similar major element but different minor and trace element compositions (oxides in percent; trace elements in ppm)

	Chemical type	
	Intermediate-Mg Grande Ronde (Grande Ronde Basalt)	Wilbur Creek (Wilbur Creek Member of Saddle Mountains Basalt)
SiO_2	54.45	54.41
Al_2O_3	14.43	14.51
FeO [1]	11.53	11.07
MgO	4.49	4.51
CaO	8.14	8.32
Na_2O	2.99	2.69
K_2O	1.44	1.77
TiO_2	1.79	1.95
P_2O_5	0.30	0.56
MnO	0.20	0.21
Cr	17.2	36 [2]
Hf	4.1	7.1
Th	1.36	6.6
Sc	35.9	27.0
La	19.5	41.0
Sm	6.1	7.9
Eu	1.84	2.4
Lu	0.54	0.69

[1] FeO+0.9 Fe_2O_3.
[2] Trace element data from J. S. Fruchter (written commun., 1977).

Chemical compositions are used extensively for purposes of flow correlation. A scheme by which flows are identified using their major element chemistry to define chemical types has been recently published (Wright and Hamilton, 1978). Major element chemistry is routinely used to back up field identification of samples, and with few exceptions is of equal or greater effectiveness than comparisons made on the basis of trace elements, which are commonly determined with less precision. Chemical compositions alone are insufficient for flow correlation within the Grande Ronde Basalt, however, and reliable correlations involve consideration of stratigraphic position, magnetic polarity, and petrography. Such correlations in the Grande Ronde indicate that individual flows in southeast Washington are of relatively restricted extent, traveling at most about 30 km from their source. Another application of major element chemistry to field studies is the correlation of dikes and flows. This is successful insofar as the chemistry is distinctive, but there is ambiguity within the Frenchman Springs and Grande Ronde Basalts, where stratigraphically separated flows of similar chemistry exist.

Sr isotopes are uniformly low (.704-.705) in the Grande Ronde and Wanapum Basalts and variably higher (.707-.715) in the Saddle Mountains Basalts, suggesting a fundamental change in the source material during the waning stages of volcanism on the Columbia Plateau. Lead isotope data are consistent with the Sr data, becoming more radiogenic in the Saddle Mountains Basalt.

We have an incomplete understanding of the petrogenesis and evolution of the Columbia River basalt, but modeling of the chemistry leads to the following broad generalizations:

1. Crustal fractionation is not an important process, and magma storage at high levels in the crust is probably minimal.

2. The lava chemistries, assuming that they represent relatively unmodified partial melts, are consistent with a mantle that is heterogeneous both with respect to mantle compositions and mineral proportions.

3. Chemical compositions are more consistent with derivation from a pyroxene-rich (e.g., garnet pyroxenite) rather than olivine-rich (e.g., peridotite) source rock and could represent moderate (10-20 percent) degrees of partial melting in a nondepleted source.

Mode of Eruption

The Columbia River basalt flows were erupted from fissures, now exposed as dikes (Fig. 3.10). Several hundred such dikes have been found, and many others must be buried by younger flows or unexposed in forested areas. Taubeneck (1970), who has done the best work on the dikes to date, estimates that a minimum of 21,000 dikes are present. This number seems somewhat excessive but even if accurate would not necessarily imply 21,000 separate eruptions, as many flows were doubtless fed by several offset dikes extending along the length of the fissure system.

Feeder dikes are known to occur within the eastern two-thirds of the province (Fig. 3.4). Many more have been found in the southeast part of the area than elsewhere, where they comprise the Chief Joseph dike swarm (Taubeneck, 1970), a name for the combined Grande Ronde and Cornucopia dike swarms of Waters (1961). Topography is much more rugged in this area, however, so that more of the basalt section is exposed than elsewhere; this alone may account for the seeming concentration of dikes in this area. We have recently discovered scattered feeder dikes for all formations within the Yakima Basalt Subgroup west and north of the Chief Joseph dike swarm, despite generally subdued topographic relief and relatively poor exposure. One such dike occurs along the east wall of the Grand Coulee 8 km south of the south end of Steamboat Rock; some others are listed in Swanson and others (1975b). We feel that the evidence, evaluated in

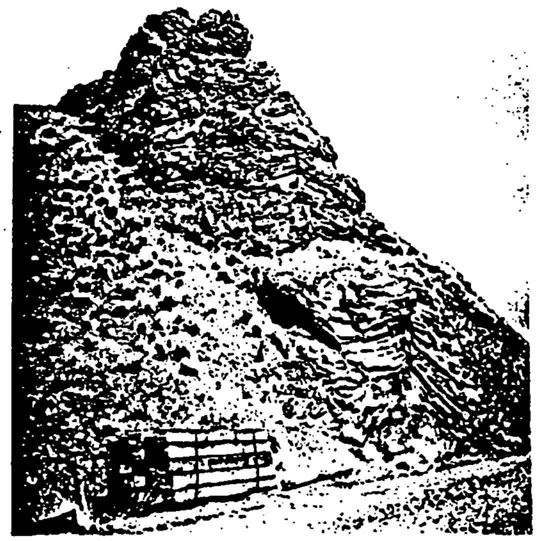

Figure 3.10. Feeder dike for flow in Frenchman Springs Member, along the Snake River about 14 km downstream from the mouth of Devils Canyon. Concave-upward columnar joints reflect influence of cooling at the ground surface, which was only about 40 m above level of exposure at the time of intrusion. Dike is about 10 m thick.

terms of quality of exposure, suggests that dikes occur more or less evenly distributed beneath the eastern two-thirds or more of the province. The Picture Gorge Basalt was erupted from a separate swarm, the Monument dike swarm in north-central Oregon (Waters, 1961; Fruchter and Baldwin, 1975).

No dikes of Columbia River basalt have been found in the Cascade Range, as discussed by Swanson (1967) and confirmed by subsequent work. This is of some importance, as the Cascades were erupting calc-alkaline rocks at this time. The distinction between the Cascades and Columbia Plateau clearly has a magmatic as well as physiographic basis, with no known overlap.

The feeder dikes average about 8 m wide but vary from a few centimeters to more than 60 m. They may tend to thin upward, but this is far from certain. The dikes cannot be traced far laterally, in part because of exposure problems. Obviously related dike segments, offset a few meters to form an en echelon pattern, form systems extending tens of kilometers. Compound or multiple dikes, consisting of two or more pulses of magma related to the same intrusive event, are common, but composite dikes, containing two or more phases of contrasting compositions, have not been reported. In other words, each fissure was utilized just once, not repeatedly.

The chance of finding a dike connecting with a flow it fed is small, owing to problems of exposure and to the fact, observed repeatedly at Kilauea, that lava drains back down a fissure after the end of an eruption, thus tending to break the contact between dike and flow. Nonetheless, several examples have been found of such a connection. In most cases, the top of the dike is rubbly, apparently consisting largely of slabs of crust once floating on a flow before it poured back into the fissure (for example, Plate 1a in Swanson and others, 1975b). In one example lacking such rubble, the dike merges imperceptibly with its flow of the Frenchman Springs Member (Fig. 1, Number 6, in Swanson and others, 1975b).

Many other dikes can be correlated with particular flows, or at least sequences of flows, on the basis of chemical and magnetic-polarity similarity. In this way, dikes have been found for most of the named stratigraphic units.

It is possible to reconstruct the nature of the vent systems for some units from the distribution of feeder dikes as well as accumulations of pyroclastic material and, in places, the preservation of spatter cones and ramparts. The results show that eruptions of single flows or related flows took place from fissures concentrated in long, narrow vent systems on the order of tens of kilometers long and several kilometers wide (Swanson and others, 1975b, and later work).

Attempts have been made to estimate the rate of eruption and advance for single flows. The estimates take into account the observation that flows, even those that advanced tens to hundreds of kilometers from their sources, quenched to a crystal-poor sideromelane glass when they entered water; this indicates little cooling during transport and hence rapid advance, since the lava apparently moved as sheet floods rather than through insulating tube systems. Application of rheologic models, developed in part from this observation by Shaw and Swanson (1970), to vent systems of known dimensions suggests eruption rates of about 1 km^3/day per linear kilometer of active fissure for the largest flows, such as those in the Roza Member, and about $10^{-4} km^3$/day/km for the smaller flows (Swanson and others, 1975b). For flows of "average" volume, probably several tens of km^3, rates of 10^{-1} to $10^{-2} km^3$/day/km may be inferred. By comparison, sustained rates of eruption at Kilauea and Mauna Loa are 10^{-3} to $10^{-4} km^3$/day/km. Using observed dike widths, theoretical modeling suggests that such eruption rates could indeed have been sustained by supply from depth (Shaw and Swanson, 1970). Such eruptions probably lasted a very few days. Flow rates of 5 to 15 km/hr down slopes of 1:1000 are calculated from the model, adequate to allow thick flows to move far with little cooling.

Rapid eruption rates do not necessarily imply rapid melting rates in the mantle. Flows were erupted only once every ten thousand years or so on the plateau during even the peak of volcanic activity, as estimated by counting the number of flows in a magnetostratigraphic unit of assumed duration based on comparison with seafloor magnetic anomalies of roughly comparable age. Calculations show that continuous melting at the Hawaiian rate, $10^{-1} km^3$/yr (Swanson, 1972) or a little less (Shaw, 1973), could account for the volume of Columbia River basalt in the allotted time. Unusually rapid melting events cannot be excluded but are not required.

If melting progressed at the Hawaiian rate, then large, deep storage reservoirs are required in order to account fo the large volume of single flows. This contrasts with the Hawaiian situation, where eruptions are much more frequent and lava "leaks" to the surface more or less continuously. The presence of large, deep storage reservoirs may be a principal and distinguishing characteristic of flood-basalt provinces in general.

SEDIMENTARY DEPOSITS

Sedimentary deposits are interlayered with the basalt in places, particularly near the margins of the plateau. Schmincke (1967c) conducted the best study of these interbeds to date, finding two dominant provenances. One is the older rocks that surround the plateau. Rivers draining these highlands carried detritus out onto the flat constructive surface of the plateau, dumping it there and then having it covered by the next flow. The second main source was the erupting calc-alkaline volcanoes in the Cascade Range. Large volumes of pyroclastic debris were blown, carried by lahars, or, most commonly, transported by rivers and distributed across the western part of the plateau in both Oregon and Washington. The thickness of such interbeds increases with decreasing age, reflecting either the slowed rate of basalt outpouring, possibly accelerated rate of Cascade volcanism, or both. A third, minor source of interbeds was the collection of diatom tests in shallow lakes that stood for thousands of years on the plateau surface. A fourth, minor source was the basalt itself, which was eroded particularly during the period of uplift during Saddle Mountains time (about 13 to 6 x 10^6 years ago).

Interbeds in the Spokane area are assigned to the Latah Formation. They are subarkosic and derived from nearby pre-basalt hills. These interbeds present a certain geologic hazard, as they frequently have failed beneath the heavy basalt layers and given rise to large landslides along steep-walled valleys. Griggs (1976) describes these slides in some detail.

A particularly important interbed is the Vantage Member of the Ellensburg Formation, to which all interbeds in the northwestern part of the province are assigned. The Vantage, primarily a subarkosic sandstone, separates the Grande Ronde and Wanapum Basalts. It is poorly exposed because of talus cover, but its erosion leads to a prominent stripped structural surface on top of the Grande Ronde in Grand Coulee, along the Columbia River, and elsewhere.

An extensive diatomite deposit occurs in the Quincy Basin and may be observed during the conference (e.g. Stop 14, Ch. 7, this volume). It is being mined north of I-90 and west of George. This diatomite was apparently deposited on top of the Frenchman Springs Member and burrowed into by the Roza Member, forming a peperite. Blocks of this diatomite were carried a short distance eastward by an old Missoula flood and deposited with gravel just east of George, as seen at Stop 13 (Ch. 7, this volume).

None of the interbeds is as extensive as most of the basalt flows. Moreover, each interbed displays lateral facies changes that make correlation based on lithology difficult. Thus, the interbeds are of little use in regional stratigraphic studies, although they are good marker beds locally. The interbeds deserve considerably more attention than they have received recently, because they hold the key for unraveling ancient drainage patterns that, in turn, reflect the tectonic disturbance of the province.

Most known faults on the plateau are closely associated with sharp folds and are considered to have formed in response to the folding. These faults approximately parallel fold axes and have thrust-type displacements. A few normal faults and fault zones, however, are neither spatially nor geometrically related to steep folds and may indicate independent zones of rupture (Fig. 3.2).

The Columbia River cuts across the anticlinal ridges in a series of spectacular water gaps, as does the Yakima River near Yakima. These gaps themselves are structurally controlled by shallow north-south synclines that cross the ridges. Some of these water gaps began to form at least 8-12 x 10^6 years ago, as shown by fluvial conglomerates of those ages confined to the area of the gaps (Wallula and Sentinel Gaps (Fig. 3.1) are good examples); later uplift of the ridges carried these deposits several hundred meters above present river level.

DEFORMATION AND TECTONIC SETTING

The northern Columbia Plateau is far from the flat, featureless area implied by its name. The plateau is actually a structural basin, with its low point in the Pasco Basin near the center of the province. Stratigraphic evidence indicates that this basin was forming during basalt emplacement and continued to subside into Pliocene and possibly Pleistocene times. The eastern half of the plateau has not sagged appreciably since 6×10^{6} years ago, however (Swanson and others, 1975a). Regional slopes toward the Pasco Basin are variable but average about 2.5 m/km from the north and east. Slopes from the south are steeper because of the Blue Mountains uplift (Swanson and others, 1977).

The western part of the northern plateau is creased by a series of gentle to sharp, even overturned, folds with structural amplitudes as great as 1,800 m but averaging about 1,000 m (Fig. 3.2). These folds, so young that the anticlines form ridges and the synclines valleys, have trends varying from almost due north to about 20 degrees of east-west (Newcomb, 1970). The folds were forming during Saddle Mountains time, as demonstrated from field relations in a number of places (Waters, 1955, 1961; Schmincke, 1967a; Bentley, 1977); some, such as the Naneum Ridge anticline in the Wenatchee Mountains (Fig. 3.2), were probably active as early as Grande Ronde time.

Anticlines and synclines are the dominant structures, but monoclinal flexures are common. One of the most prominent monoclines is the Coulee Monocline (Fig. 3.2) which controlled the erosional etching of the lower Grand Coulee (Stop 8, Ch. 7, this volume). Its structural relief is more than 300 m. Many of the folds are boxlike, consisting of two or more monoclinal flexures with opposed directions of dip. This type of structure is thought to form under little confining pressure, consistent with the evidence for the youth of the structures and the lack of indication that much overburden has been stripped off the plateau. Bentley (1977) contends that "in gross character, the anticlines are 'drape' folds caused by vertical breakup of basement rocks probably coincident with the major . . . uplift of the Cascades."

The regional tectonic setting of the Columbia Plateau is complex and poorly understood. The plateau occupies a position inland of the partly coeval Cascade Range and in this manner resembles marginal basins of the western Pacific. Furthermore, east-west extension, indicated by dike orientations, is consistent with such an analogy. However, this zone of extension apparently continues, with local breaks, north into interior British Columbia and south into the Basin and Range province far beyond the limits of Cascade volcanism. Moreover, other flood-basalt provinces on Earth appear to have formed in areas of incipient continental rifting, not marginal basins. By analogy, then, the Columbia Plateau may occupy part of an incipient or slowly developing continental rift. Neither of these settings, however, explains adequately the complex contemporaneous deformation of the province. We have a long way to go before sense can be made of the tectonic setting of the plateau, and it will only be understood after integrated studies of the Cenozoic volcanic and tectonic history of the entire Pacific Northwest.

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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 dis-

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tinctive 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) postflood 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

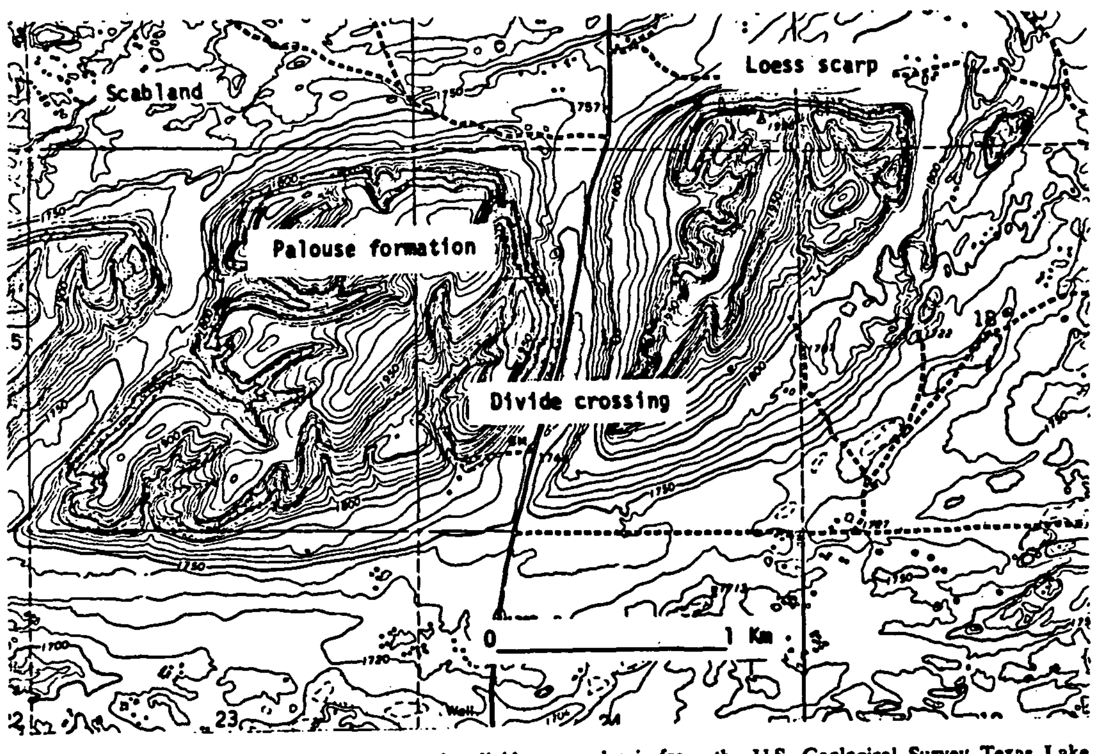


Figure 4.1. Loess scarps, scabland, and a major divide crossing in the Cheney-Palouse scabland tract. Topography is from the U.S. Geological Survey Texas Lake Quadrangle (7.5 minute, 10 feet contour interval).

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.

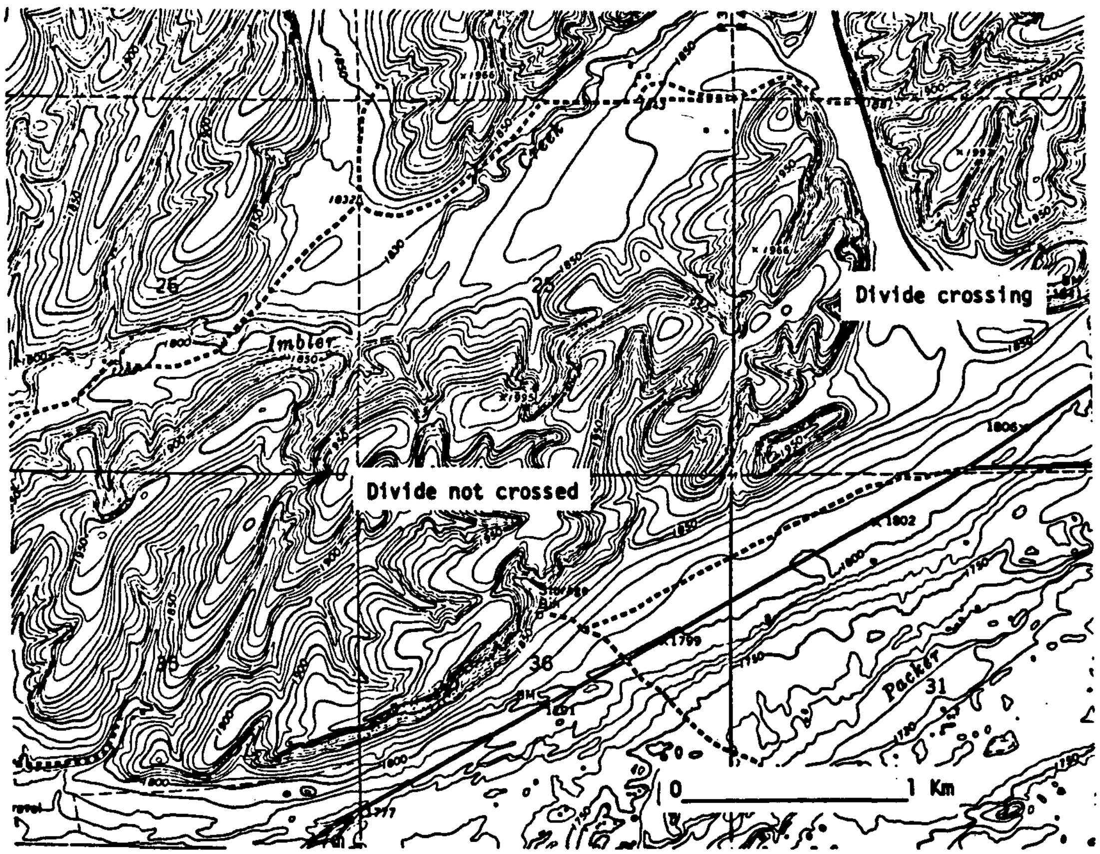


Figure 4.2. Examples of crossed and uncrossed loess divides in the Cheney-Palouse scabland tract. The divide crossing depicted is a definite example with a linear, flat-floored trough. Topography is from the U.S.G.S. Texas Lake Quadrangle.

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

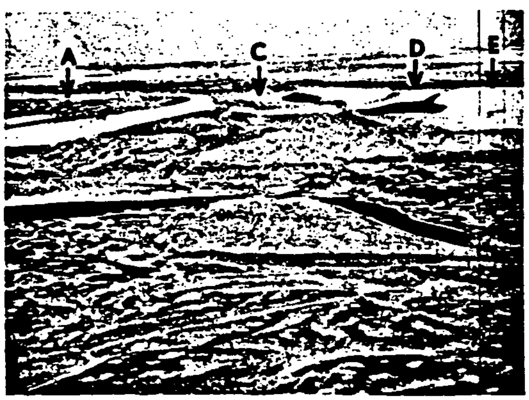


Figure 4.3. Oblique photograph of minor divide crossings (arrows) cut through a loess divide between Crab Creek and South Fork (Sections 26, 27, and 28, T.21N, R.36E.). The floors of the divide channels A, C, and D are at 600, 588, 594 meters (1970, 1930 and 1950 feet) elevation respectively. The uncrossed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

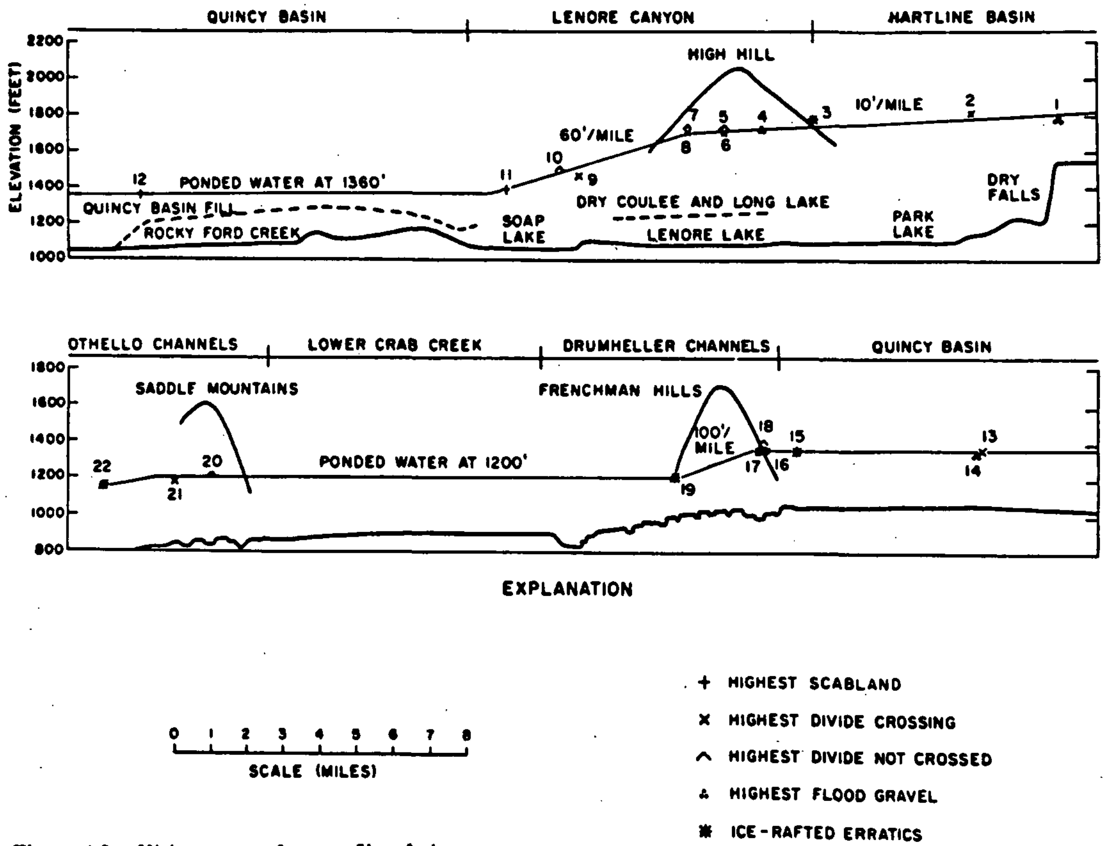


Figure 4.5. High-water surface profile of the western Channeled Scabland. The numbered high-water marks are individually described by Baker (1973a).

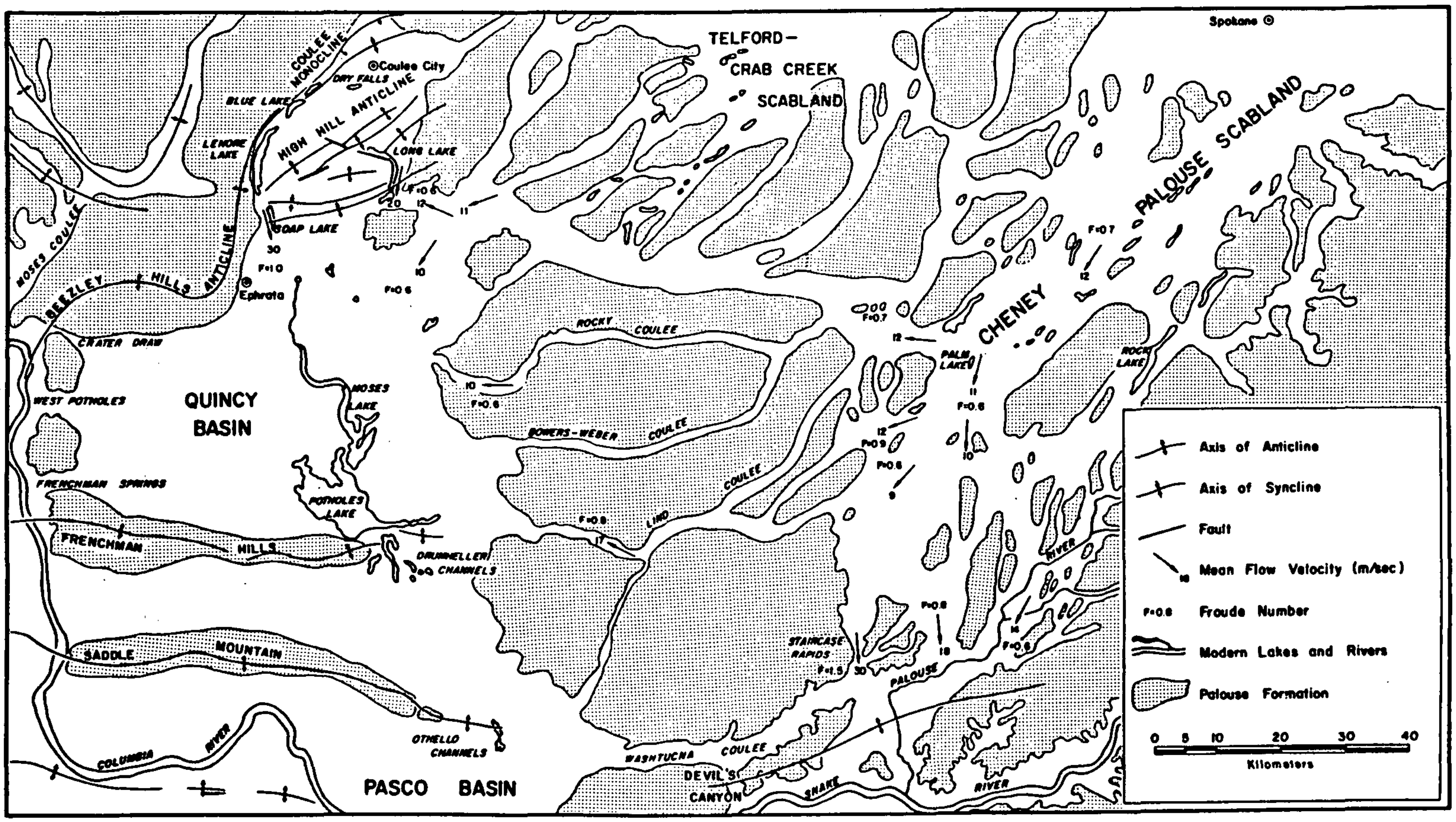


Figure 4.6. Regional paleohydraulic features of the Channeled Scabland. Mean flow velocities were determined from high-water mark evidence and channel geometry by Baker (1973a). Also shown is the regional structural pattern on the Columbia Plateau. The Palouse Formation is not shown on High Hill and 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 21.3×10^6 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 2.13×10^6 m^3/sec. Even at this phenomenal discharge, the lake volume of 2.0×10^{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 9.1×10^6 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 western Channeled Scabland (Fig. 4.5), water was ponded 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.

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.

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 4.10. 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-

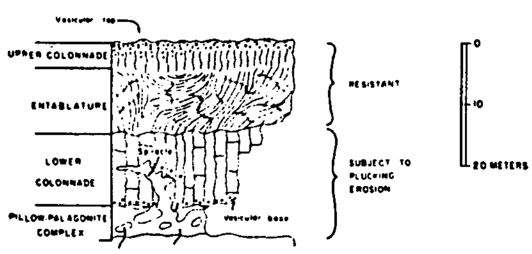
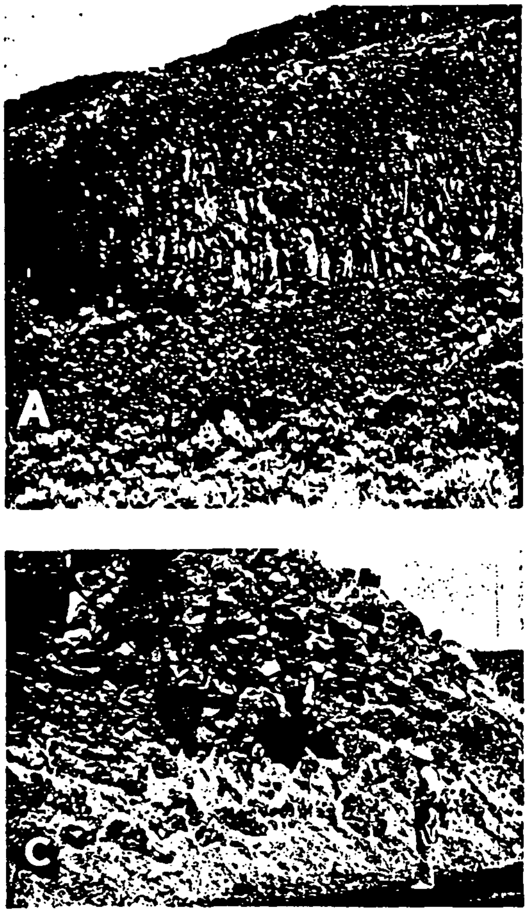


Figure 4.12. Cross section of an idealized Yakima basalt flow showing structural features important to flood erosion processes. The upper colonnade and the pillow-palagonite complex are only present in some basalt flows. Diagram is modified from Swanson (1967) and Schmincke (1967a) and uses the terminology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



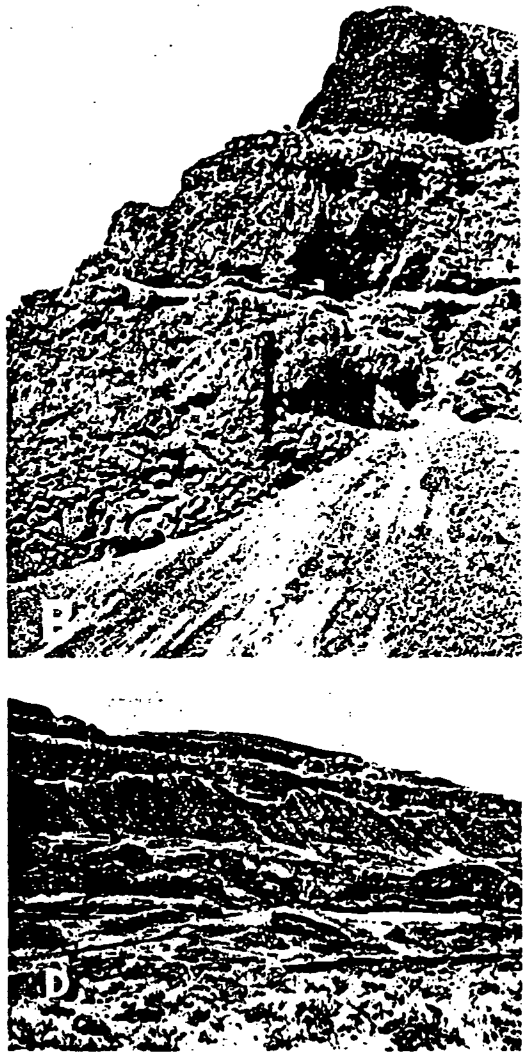


Figure 4.14. Structural characteristics of Yakima Basalt flow units. A. Entablature forming a resistant bench in flood-eroded scabland near Lenore Caves. B. Pillow lavas with overlying columns exposed in the wall of Moses Coulee near Appledale, Washington. C. Pillow-palagonite (Rosa Member) exposed in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

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.

HYDRODYNAMIC EROSION

Cavitation

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

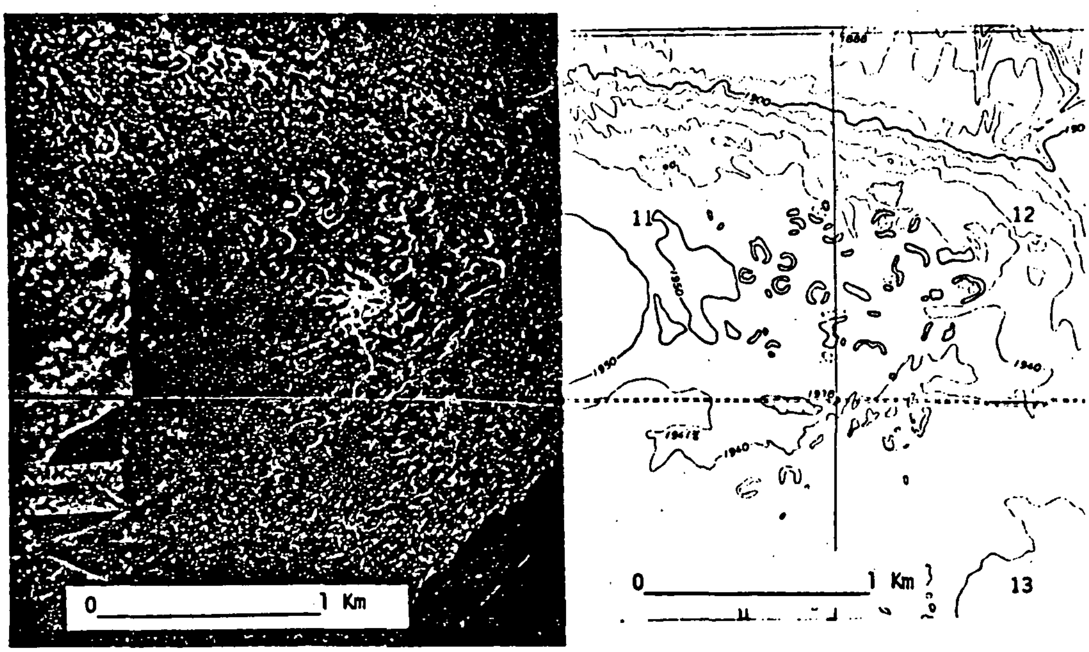


Figure 4.17. Vertical aerial photograph and topographic map (10 feet contour interval) of the ring structures shown in Fig. 4.16. These rings occur at 40 km east of the prominent rings described for the Odessa area by McKee and Stradling (1970) and by Hodges (1976).

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).

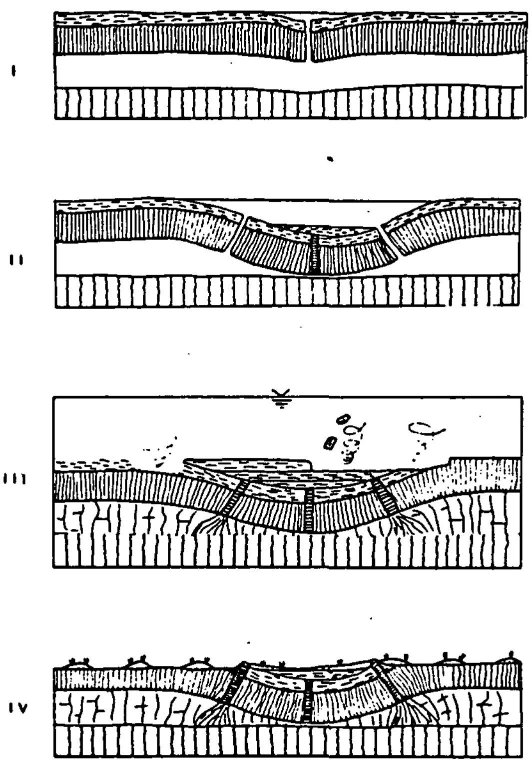


Figure 4.18. Hypothetic sequence of development for basaltic ring structures (modified from McKee and Stradling, 1970). I. Flowout of lava from the molten interior of a crusted lava flow. II. Sagging of the lava flow crust induced by loading from a point source induces circular fractures that allow additional 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:

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

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), γ 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

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

For terrestrial conditions near sea level, this equation can be further reduced by assuming $g = 9.81 \text{m/sec}^2$, $P_a = 1$ atm., $P_v = 0.024$ atm. (21°C), and $\gamma = 9.8 \times 10^3$ N/m³. The equation for the critical cavitation velocity V_c (m/sec) is then a simple function of flow depth (m):

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

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

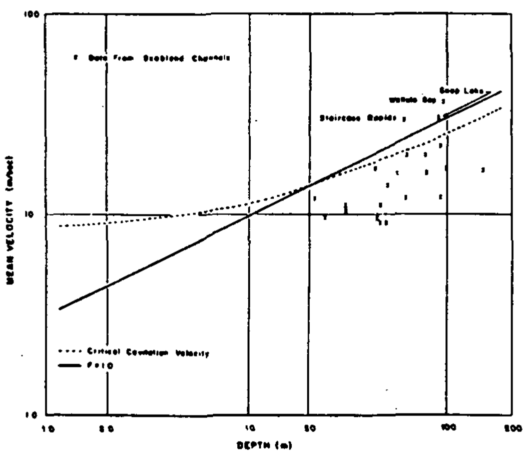


Figure 4.19. Influence of flow depth on the critical mean flow velocity necessary for cavitation to occur according to Barnes (1956, p. 499). Hydraulic data from scabland channels are plotted for comparison. Most of the flood flows were sub-critical and below the critical depth-velocity combinations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

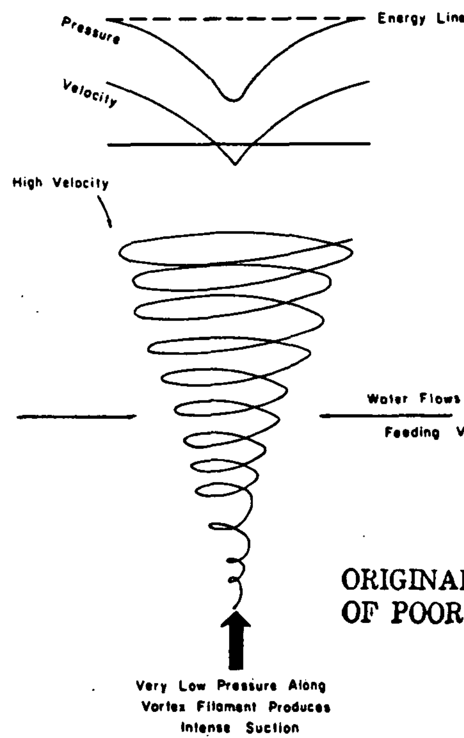


Figure 4.20. Characteristics of a hydraulic vortex ("kolk"). Such macroturbulent effects are set up by flow obstructions. The intense pressure and velocity gradients of the vortex produce a hydraulic lift. Alternatively, Jackson (1976) ascribes "kolks" to the turbulent bursting phenomenon.

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

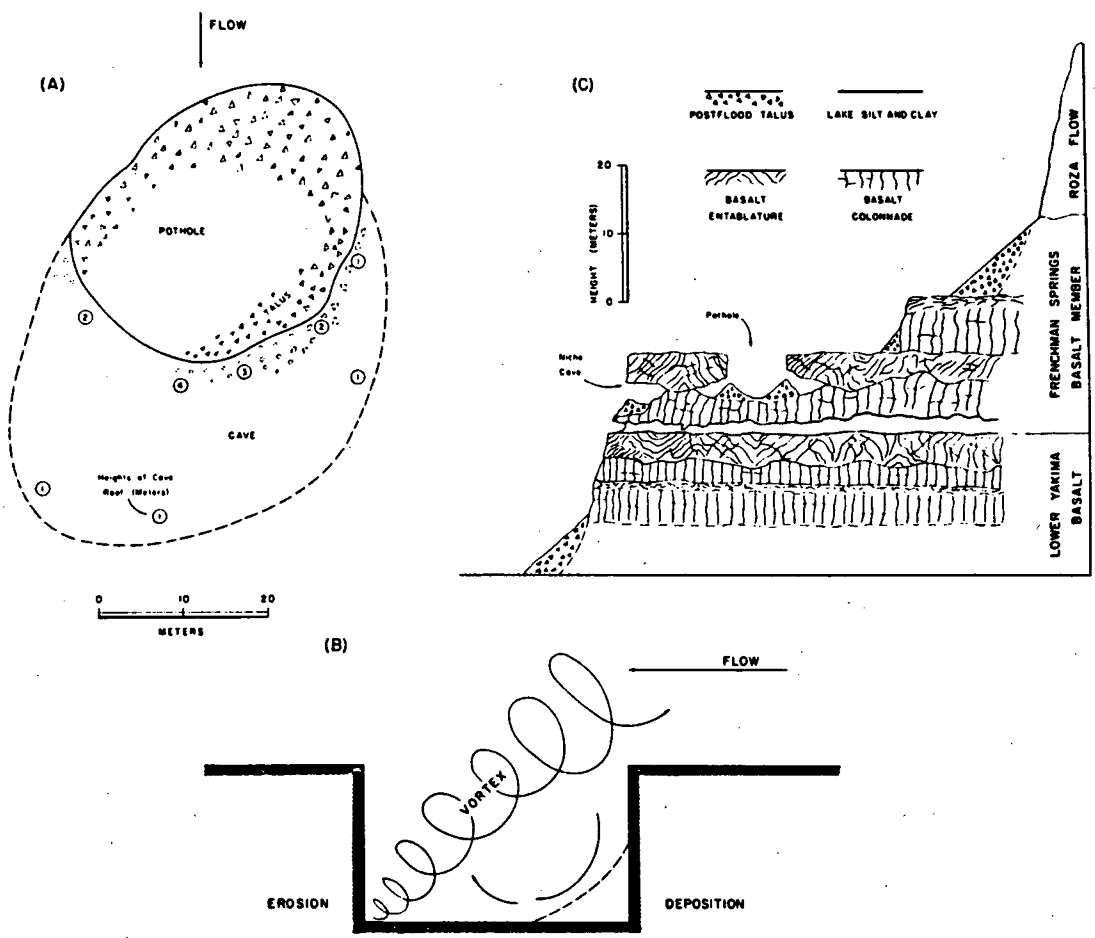


Figure 4.21. Pothole erosion at Lenore Caves, 15 km north of Soap Lake in Lenore Canyon. The sketch map (A) shows undercutting of the pothole rim on the downstream side. This is consistent with experimental observations of vortex scour (B). The relationship of pothole to the local basalt section is shown 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 τ is the mean shear stress for initiating particle transport, γ is the specific weight of the transporting fluid (9.8×10^3 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:

$$D = 65\tau^{0.54}, \tag{4-7}$$

where D is the intermediate particle diameter (mm) and τ 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-

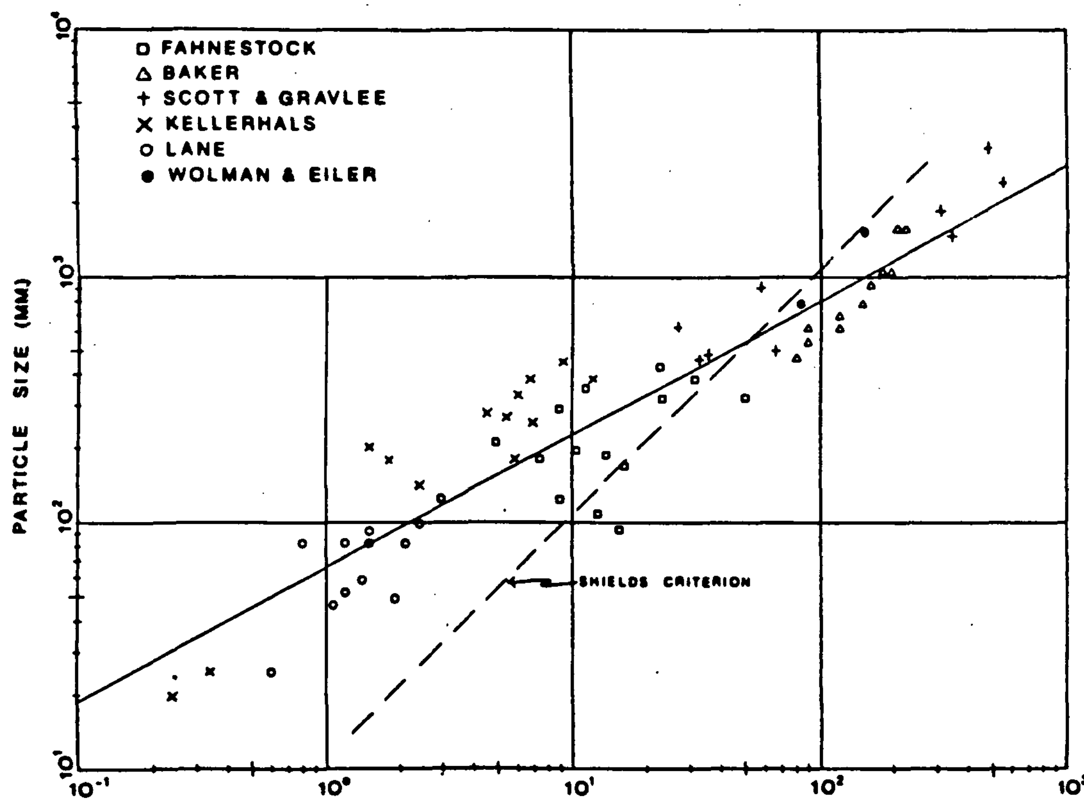


Figure 4.22. Intermediate particle diameter versus shear stress for extremely coarse bedload. The triangles indicate data for boulder sizes and shear in relatively uniform scabland reaches.

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

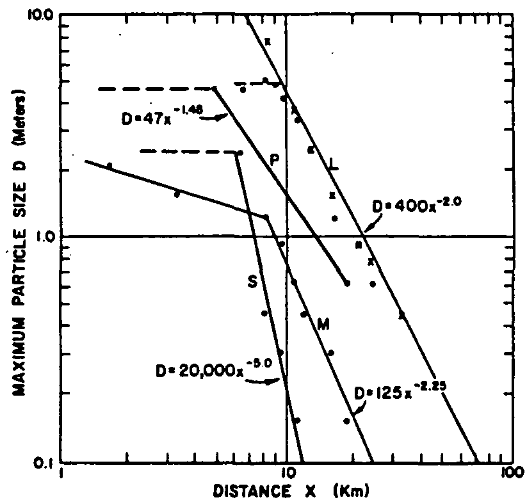


Figure 4.23. Downstream decrease in intermediate diameter of the largest boulders deposited below constrictions by catastrophic terrestrial floods. Data were obtained from the following sources: (L) Lenore Canyon, Channeled Scabland (Baker, 1973a), (P) Portland area below Columbia Gorge (Trimble, 1963), (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.

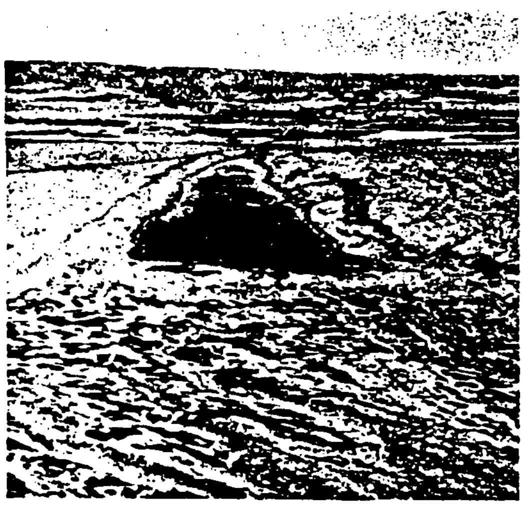


Figure 4.24. View north across the Ephrata Fan toward the Soap Lake constriction.

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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Chapter 5

Large-Scale Erosional and Depositional Features of the Channeled Scabland

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ABSTRACT

The Channeled Scabland is a great anastomosing complex of highly overfit steam 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'

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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") or bedrock streams. The scabland anastomosis is deeply cut into rock. Anastomosis occurs in the Channeled Scabland because preflood valleys did not have the capacity to convey the Missoula Flood discharges without spilling over preflood 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 channnel 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 preflood 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.

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).

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 Penteado-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

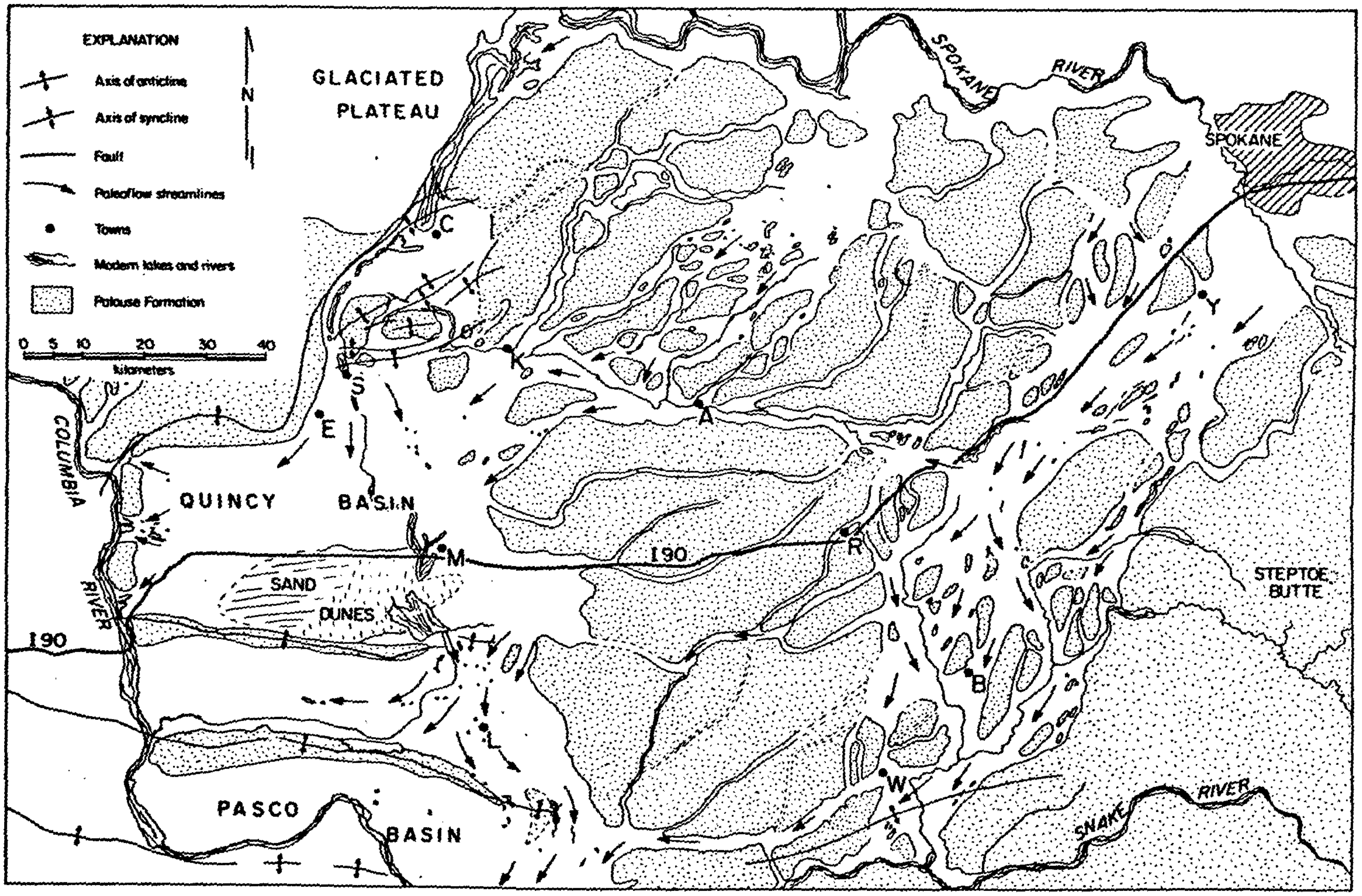

Figure 5.2. Regional pattern of the Channeled Scabland as mapped from LANDSAT imagery (E-1039-1843-5 and E-1004-18201-7). (Y) Cheny, (R) Ritzville, (B) Benge, (W) Washtucna, (M) Moses Lake, (E) Ephrata, (S) Soap Lake, (C) Coulee City.

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 m^3/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 m^3/sec (Baker, 1973a, his Plate 1). Bretz (1928b) noted that the flooding could not tolerate the leisurely preflood 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 preflood 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 preservation 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 preflood valley of Crab Creek. The bar is developed for 3 km downstream of the trenched spur buttes (T) that mark the preflood 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

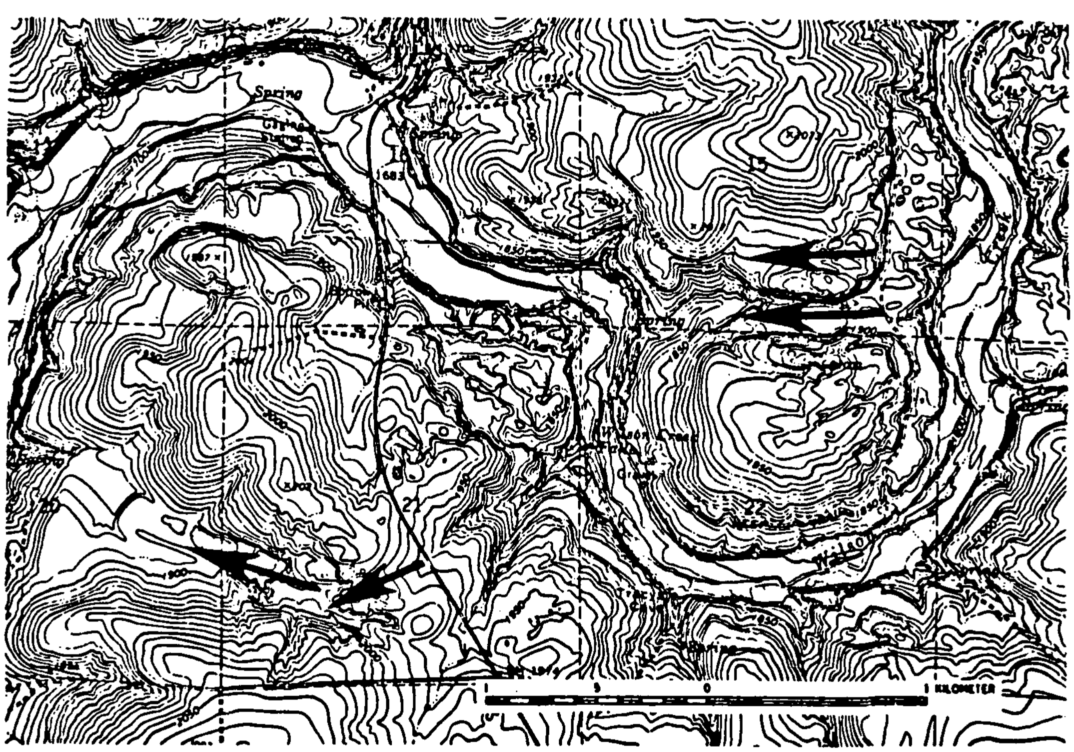

Figure 5.4. Topographic map of the upstream part of Wilson Creek (U.S.G.S. Almira, Wash., 7.5-minute quadrangle). Contour interval is 3 m (10 feet). Arrows show major divide crossings.

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. Cateracts.	Scour Marks.	Large-scale Transverse Ripples (Giant Current Ripples).
MICROFORMS	Scallop Pits.	Not Preserved.	Small-scale Ripple Stratification. (Restricted to Slackwater Facies.)

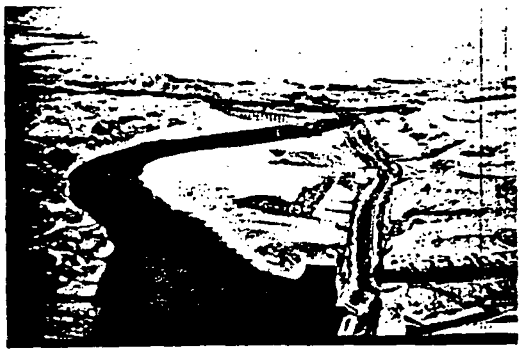

Figure 5.5. Excellent association of depositional macroform (bar) with superimposed mesoforms (giant current ripples). The bar occurs along the Snake River just downstream from its junction with the Palouse River (top center). The Snake River is impounded by a dam just downstream from the reach depicted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

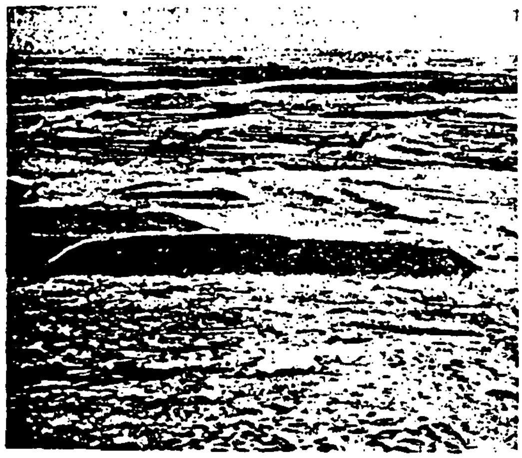

Figure 5.6. Oblique aerial photograph of the upstream and of a quadrilateral loess residual on the Palouse-Snake divide. Note the steep scarps that were eroded by flood water streaming over the divide. Washtucna Coulee is visible in the background.

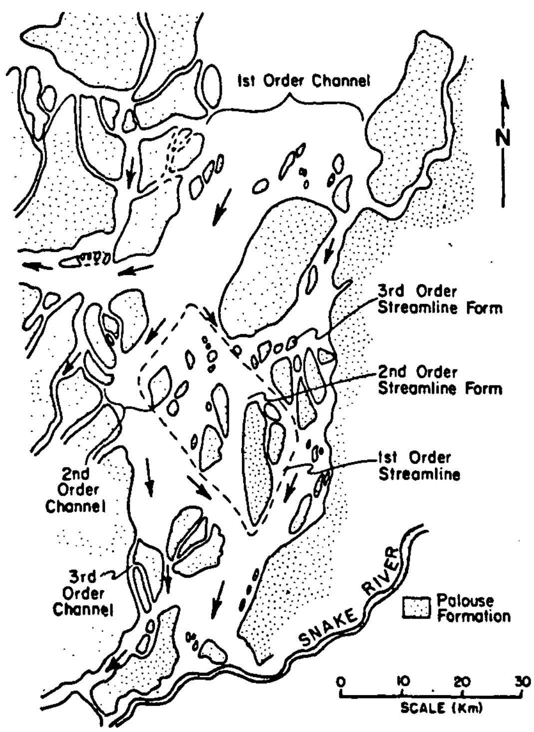

Figure 5.7. Sketch map of the Cheney-Palouse scabland showing the hierarchial relationship of channels and interchannel residual areas (mantled by Palouse Formation). An arbitrary section has been ordered to show these relationships.

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 factors 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, ungullied 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:

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

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

$$P = l \cos K\theta, \qquad (5\text{-}2)$$

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

$$A_t = lw. \qquad (5\text{-}3)$$

However, from Figures 5.12 and 5.13 we have:

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

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

and

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

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:

$$A_t = l^2.$$

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

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

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:

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

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

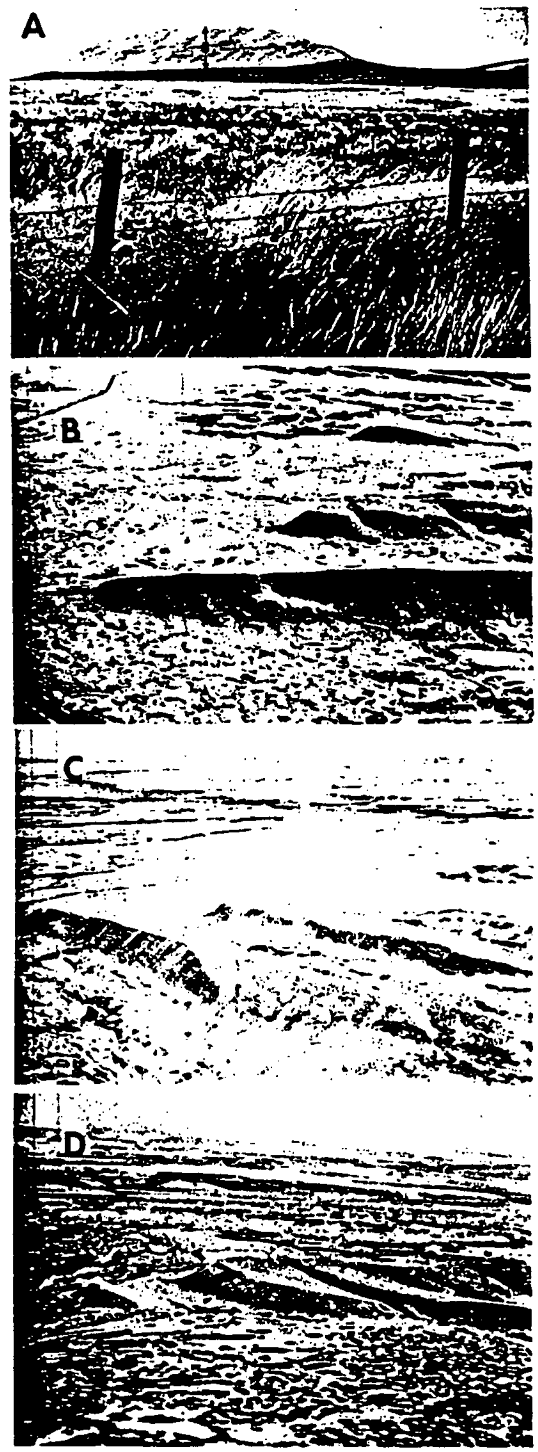

Figure 5.8. Streamlined residual forms in the Cheney-Palouse scabland: A. Ground-level view showing height(s) of the flood-eroded loess scarp. B. "Upstream" end of the loess hills on the Palouse-Snake crossing. Note the prominent prows formed by convergent hillslopes pointing upstream for the flood flows. 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 increasing 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:

$$R = \frac{Q R \rho}{A \mu}, \tag{5-9}$$

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.

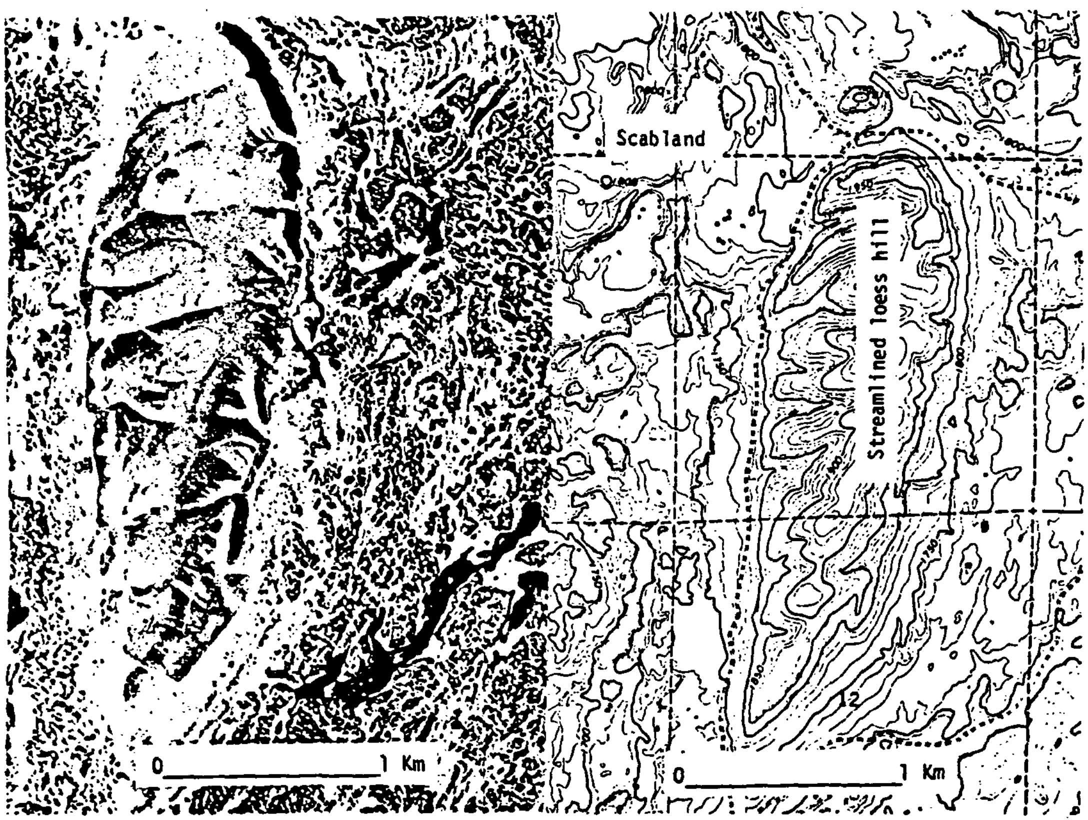

Figure 5.9. Residual loess hill streamlined by flood erosion. A small cataract, heading an inner channel, has worked its way around the blunt upstream end of the hill (on map). Longitudinal grooves and butte-and-basin topography can be seen in the marginal scablands. Water depths and velocities averaged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

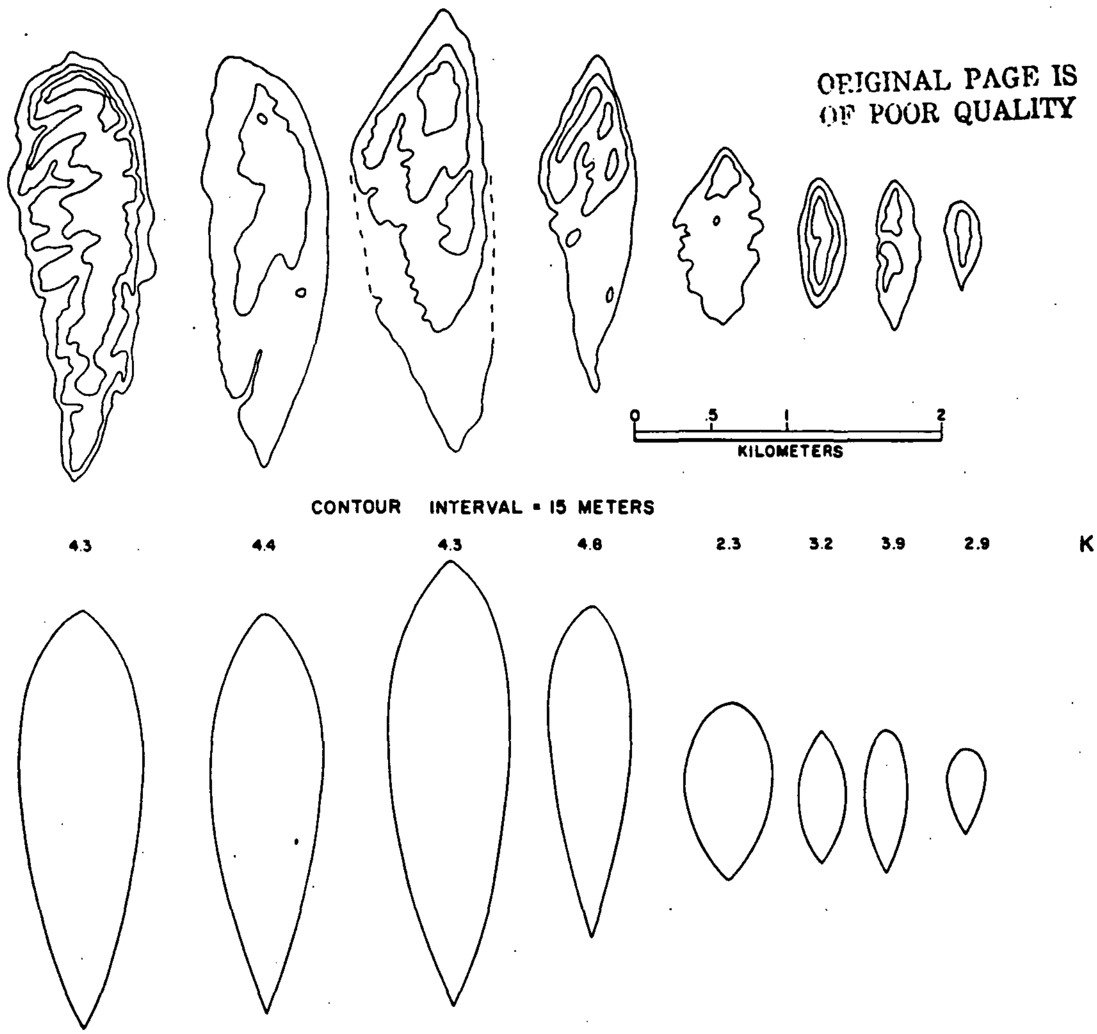

Figure 5.10. Comparison of typical streamlined scabland loess hills to equivalent lemniscate loops as calculated according to Chorley's (1959) analysis of drumlin shapes. Water flowing over and around the easily scoured loess caused the streamlined shapes, permitting less resistance to flow. Subsequent gulleying of the soft loess has somewhat modified the streamlined shapes.

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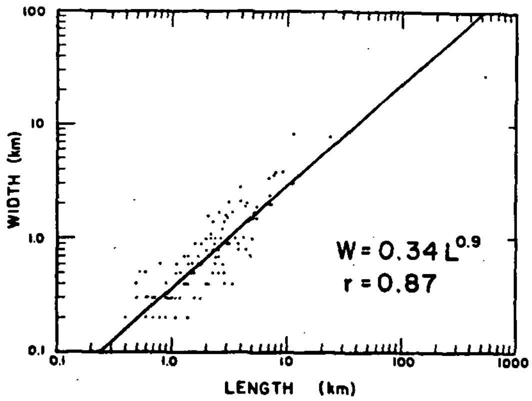

Figure 5.11. Maximum width of scabland streamlined forms versus the length of the form.

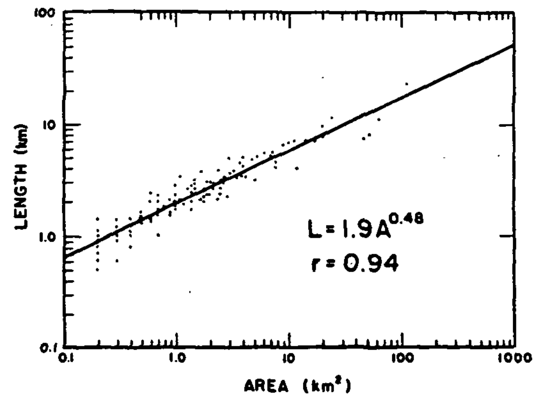

Figure 5.12. Length of scabland streamlined forms versus form area.

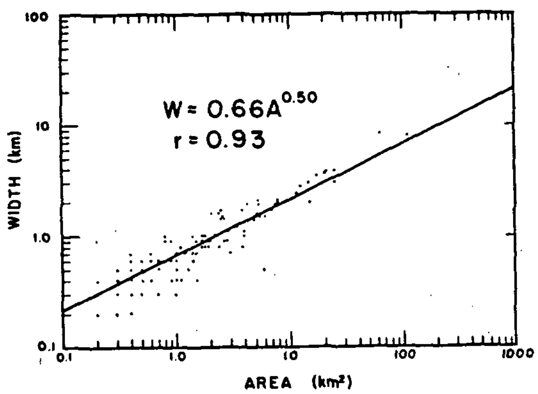

Figure 5.13. Maximum widths of scabland streamlined forms versus form area.

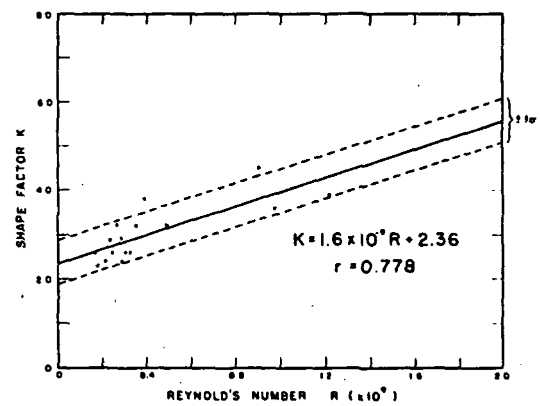

Figure 5.14. Streamlined shape factor K versus the Reynold's number for maximum scabland flood flows associated with scabland streamlined forms. The dashed lines indicate one standard error.

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

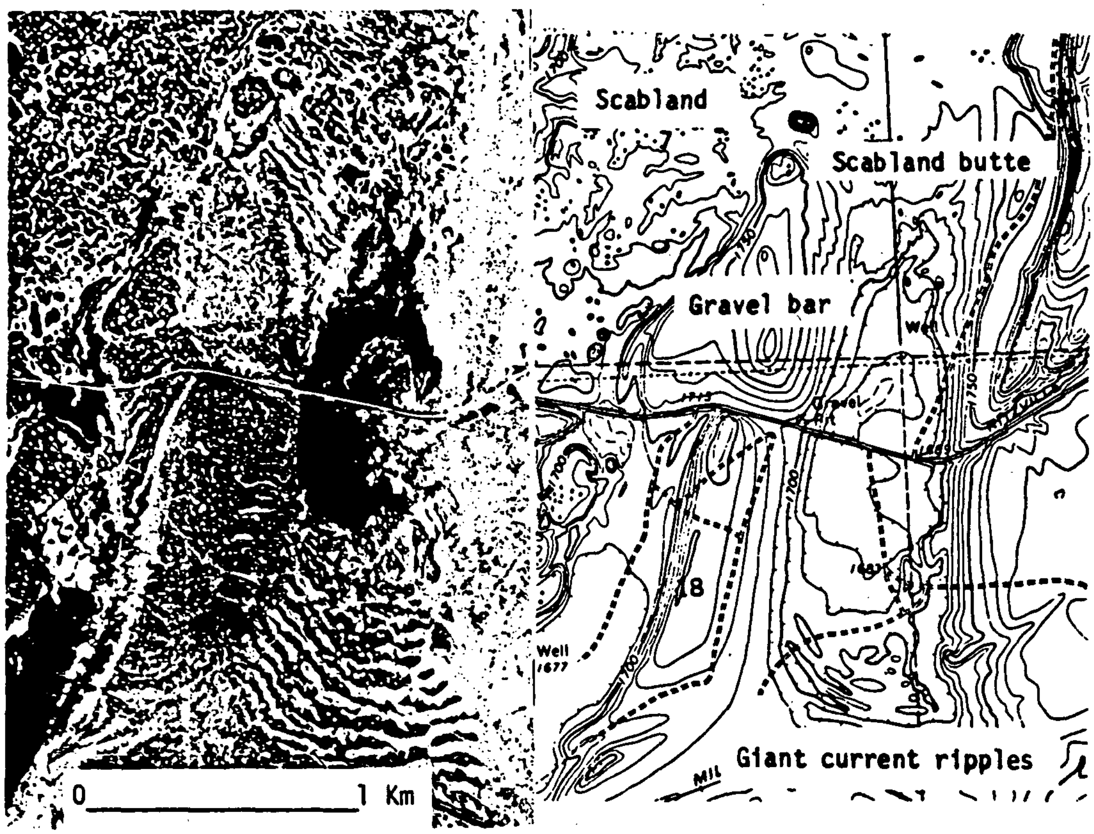

Figure 5.15. Vertical aerial photograph and topographic map of a small pendant bar in the Cheney-Palouse scabland near Macall, Washington. Contour interval is approximately 3 m (10 feet). The bar accumulated downstream from a residual butte of basalt. The bar occurs in sections 7 and 18, T. 18N., R. 38E.

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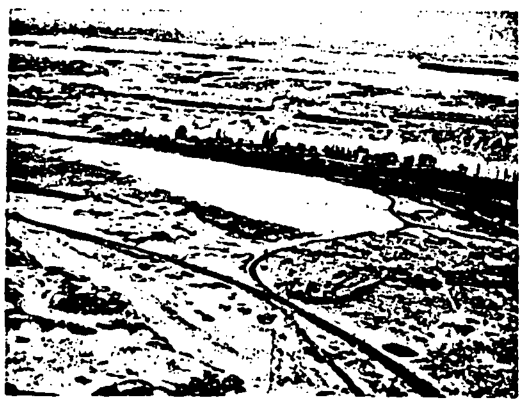

Figure 5.16. Oblique aerial photograph of Bar 2 near Wilson Creek, Washington. This is a relatively small bar about 1 km in length.

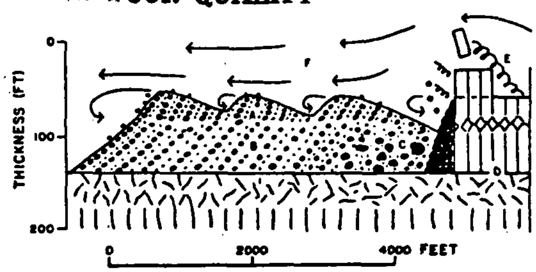

Figure 5.17. Schematic development of a hypothetical pendant bar. Bedload is transported across the surface of the bar by giant ripples and deposited at the downstream end of the bar. Explanation: A—foreset bedding associated with giant current ripples, B—foreset bedding within the bar, C—chaotically 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

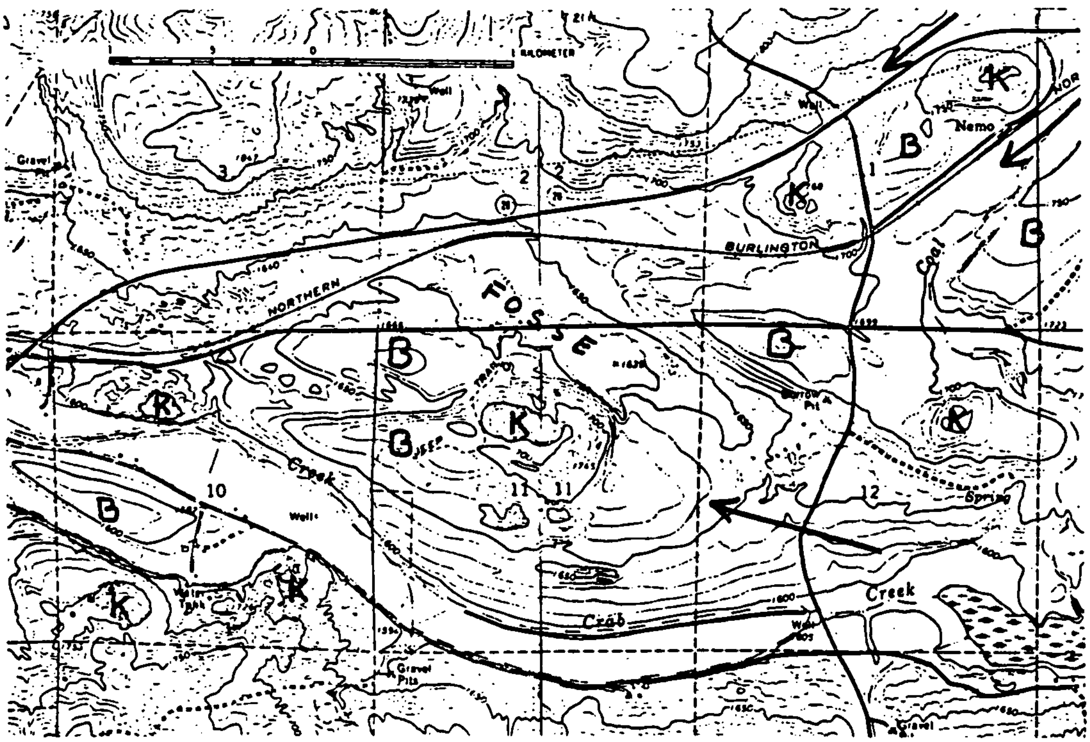

Figure 5.20. Topographic map of pendant bars (B) near Odessa, Washington. The prominent basalt knobs (K) result from the resistance of circular dikes intruded into the jointed lava flows. Contour interval is 3 m (10 feet). Topography is from the Odessa and Sylvan Lake, Washington, 7.5-minute quadrangles.

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)

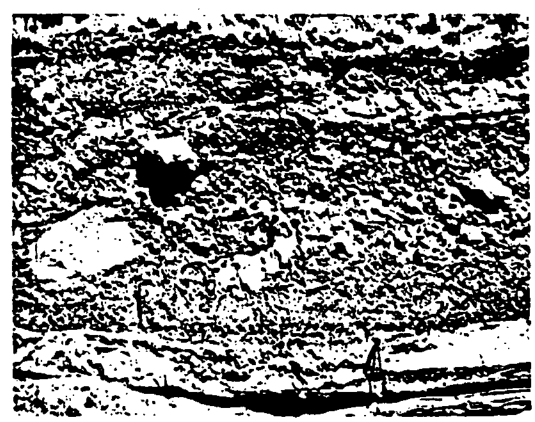

Figure 5.21. Eddy bar deposits at the mouth of the Tucannon River. The lowest layer of horizontally bedded sand is overlain by about 60 cm of laminated silt. This is overlain by poorly sorted, angular to subangular flood gravel containing percussion flaked boulders and cobbles. The prominent boulders 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 eddys 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 featuers 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 subfluvially 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 horseshoe-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,

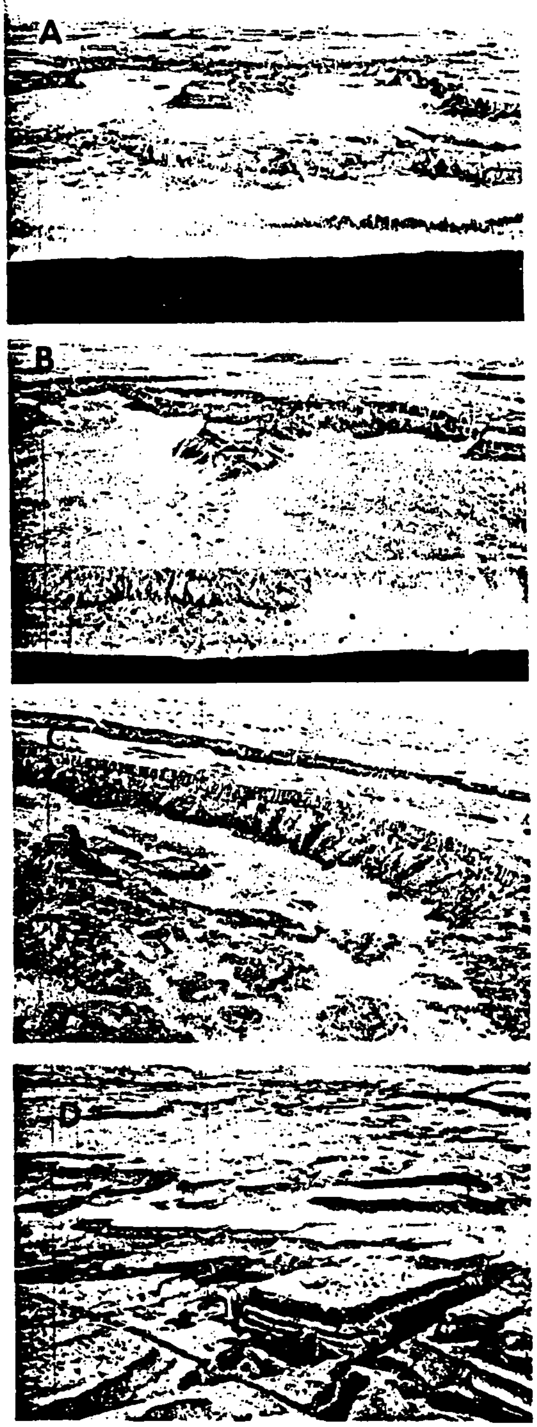

Figure 5.22. Oblique aerial photographs of scabland cataracts and inner channels. A. West Potholes cataract on the western rim of the Quincy Basin. B. Frenchman Springs cataract also on the western rim of the Quincy Basin. C. Hudson cataract at the head of Hudson Coulee in the Hartline Basin. D. Palouse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

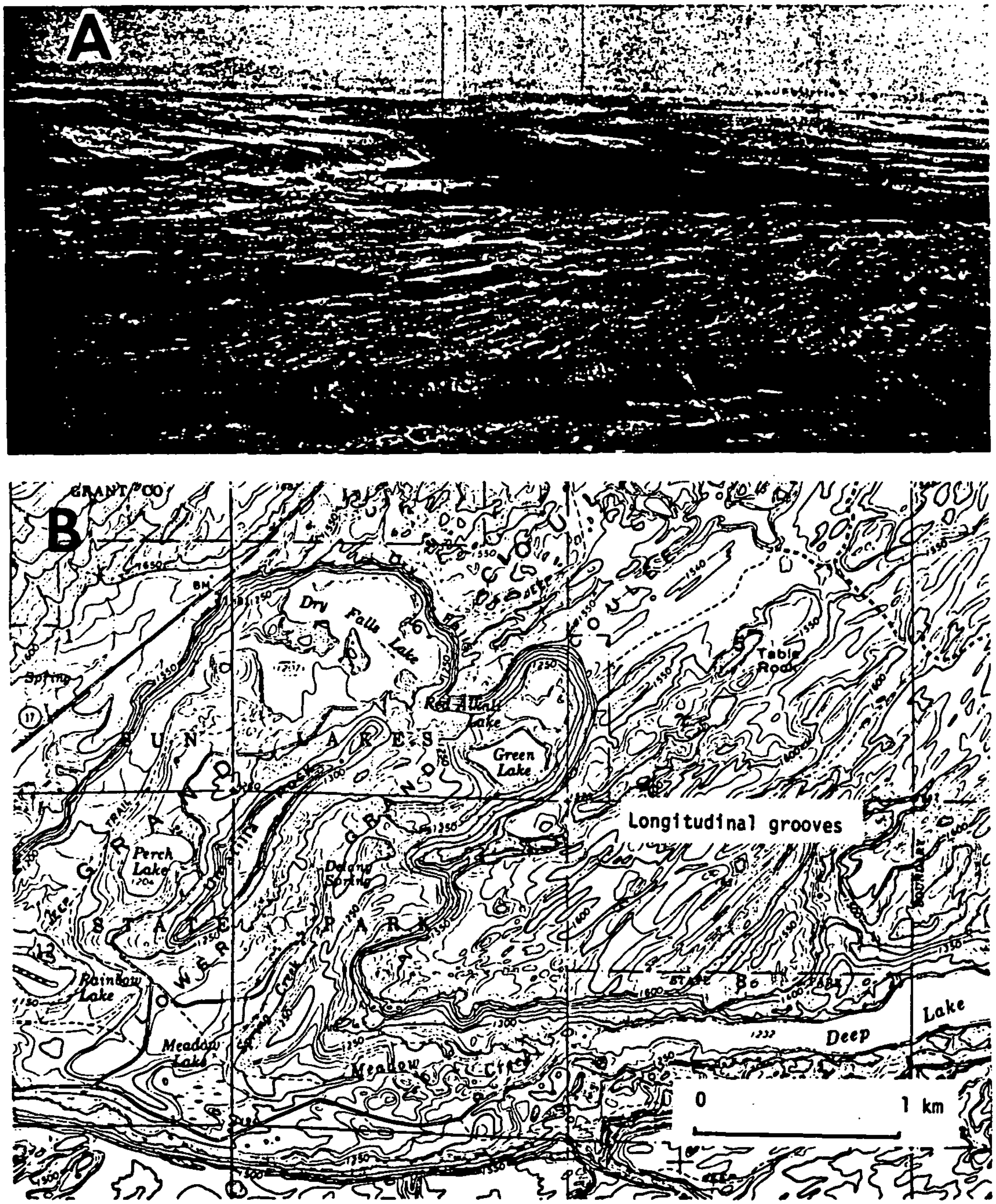

Figure 5.23. A. Oblique aerial photograph of the Dry Falls cataract group. The cataract is 120 m high and 5.5 km wide. Longitudinal grooves are visible just upstream from the cataract head, and the upper Grand Coulee (containing Banks Lake) extends to the horizon. B. Topographic detail of the Dry Falls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

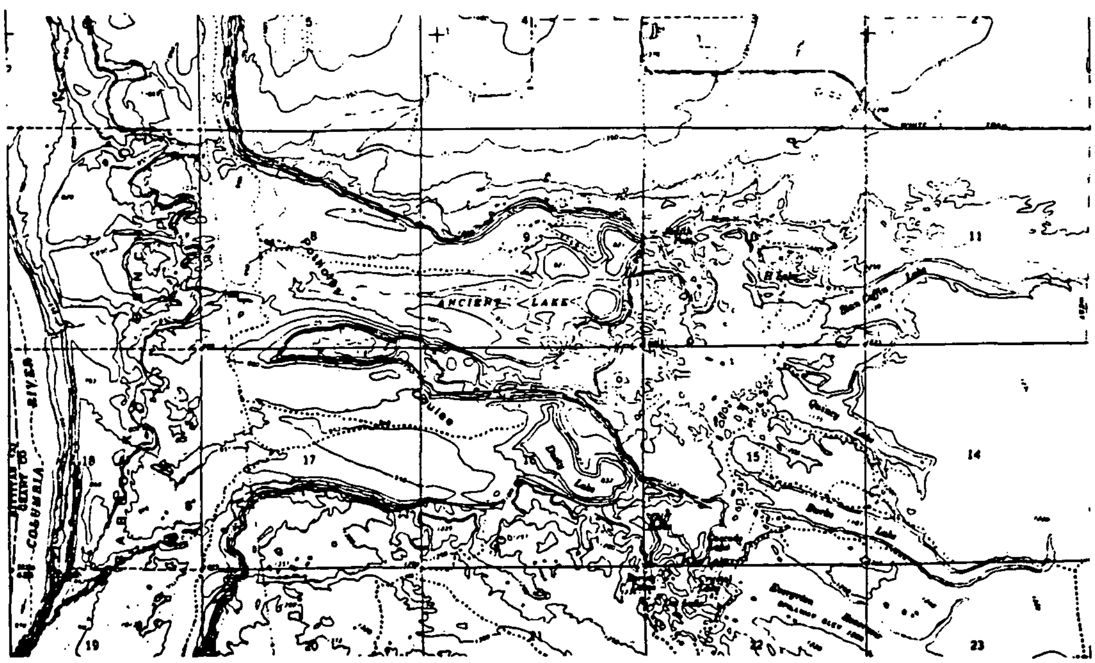

Figure 5.25. Topographic map of the Potholes Cataract. Topography is from the Babcock Ridge 7.5-minute quadrangle.

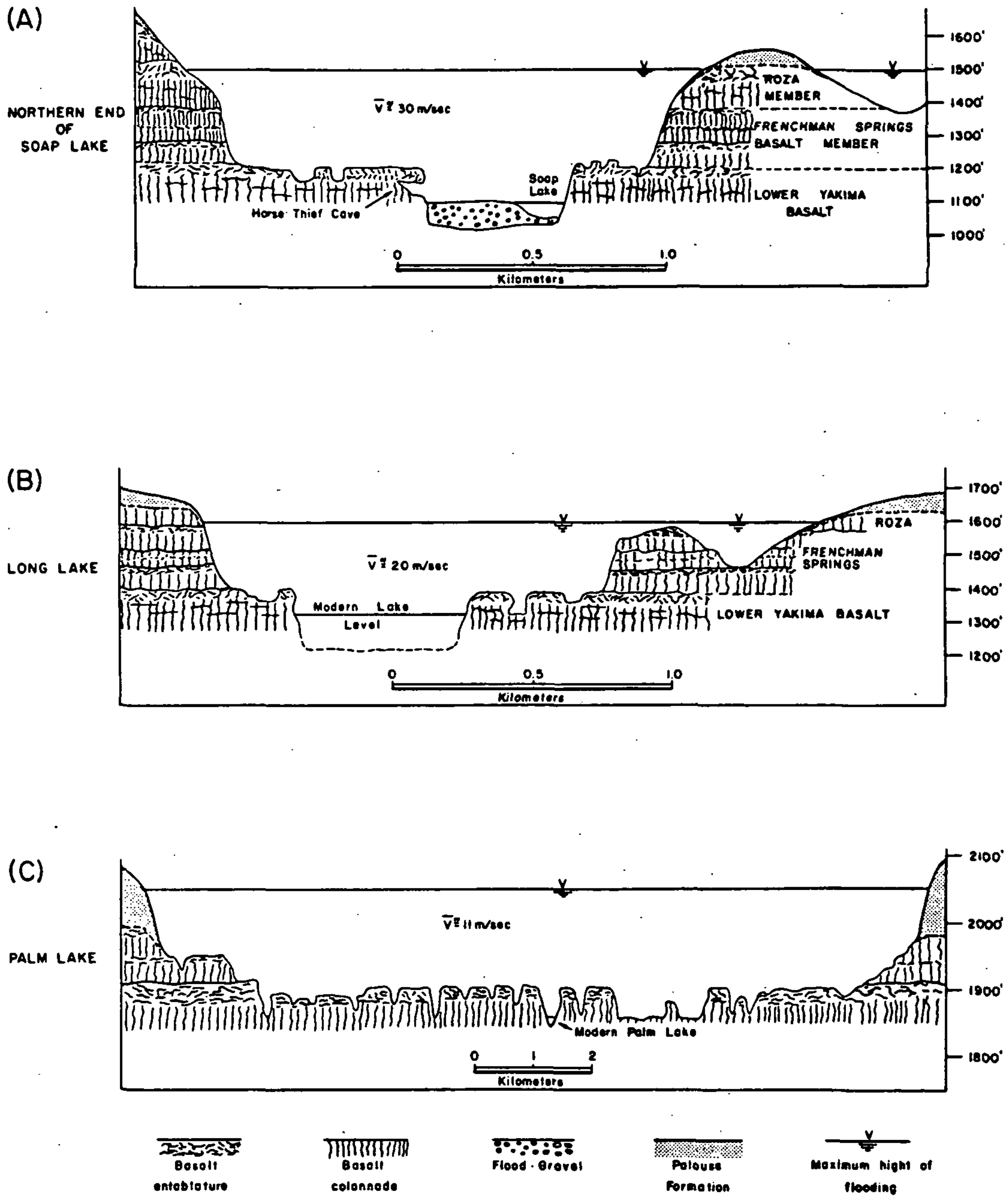

Figure 5.26. Representative cross sections of scabland channels showing structural characteristics of basalt, erosional features, and flood stage plus maximum mean flow velocity ($\overline{V}$). The sections at Soap Lake (A) and Long Lake (B) show inner channels with flanking butte-and-basin scabland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

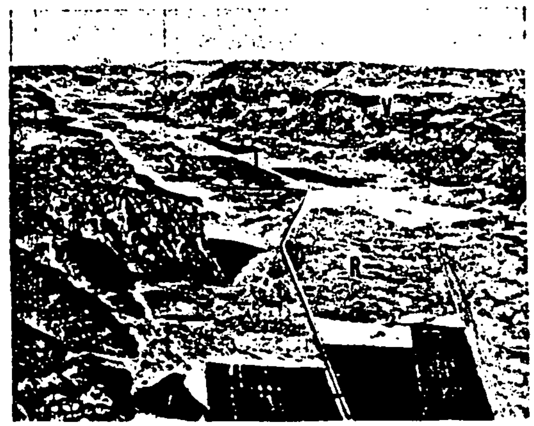

Figure 5.27. Inner channel development in Moses Coulee. This oblique aerial photograph shows a distinct inner channel (I) on the coulee floor with marginal butte-and-basin scabland (S). Giant current ripples (R) occur on the surface of a flood gravel bar at the mouth of the inner channel. Preflood tributaries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

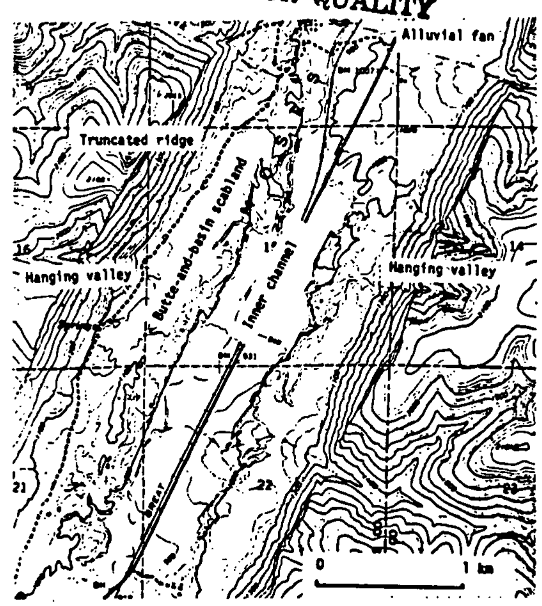

Figure 5.28. Topographic map of lower Moses Coulee showing the relationships discussed in Fig. 5.27. The contour interval is 10 feet (elevations in feet).

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 bizzare 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)

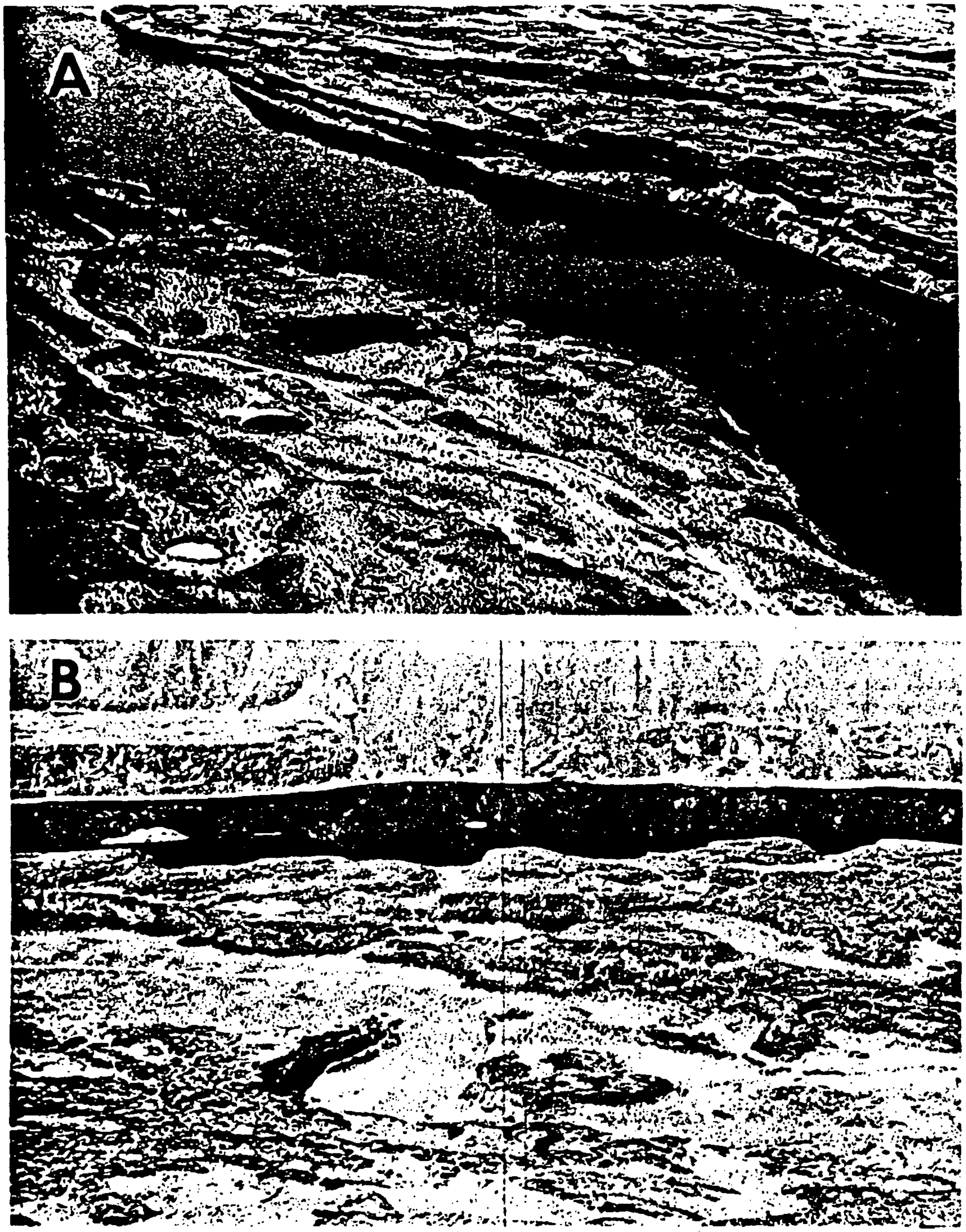

Figure 5.30. Butte-and-basin topography near Long Lake (A) and Blue Lake (B). Potholes show undercutting of their entablature rims particularly on the downstream side.

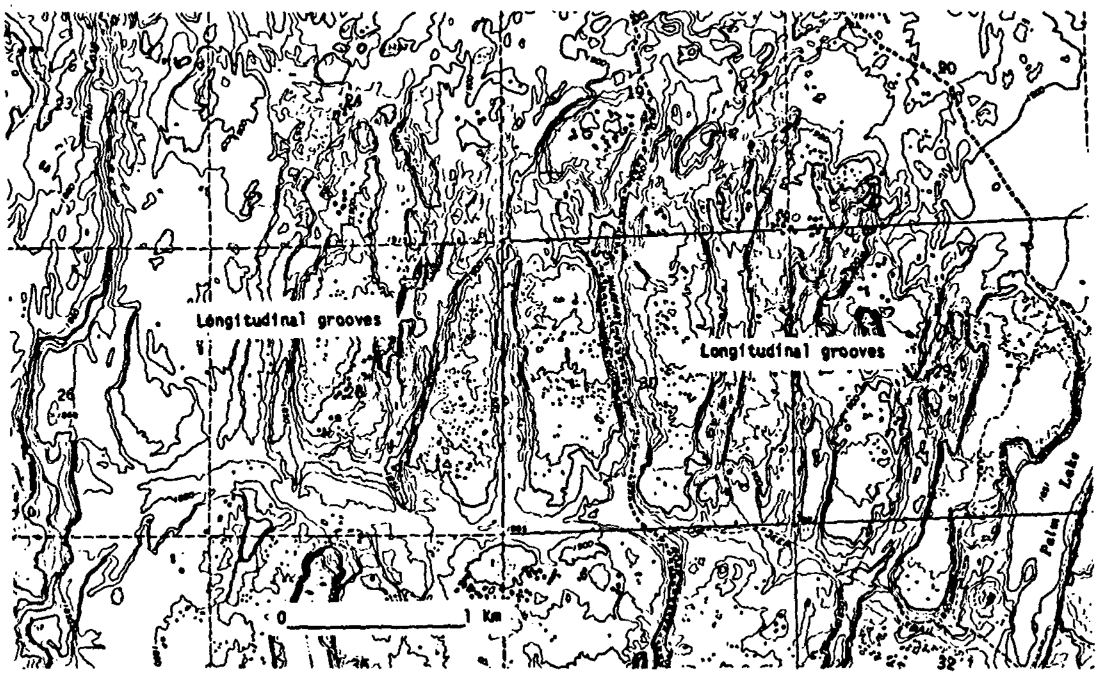

Figure 5.31. Longitudinal grooves developed on the scabland surface near Palm Lake. The small closed contours are mostly mounds of post-flood silt. The larger closed depressions are scabland basins. Topography is from the Palm Lake, Washington, 7.5-minute quadrangle.

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.

I

II

III

IV

V

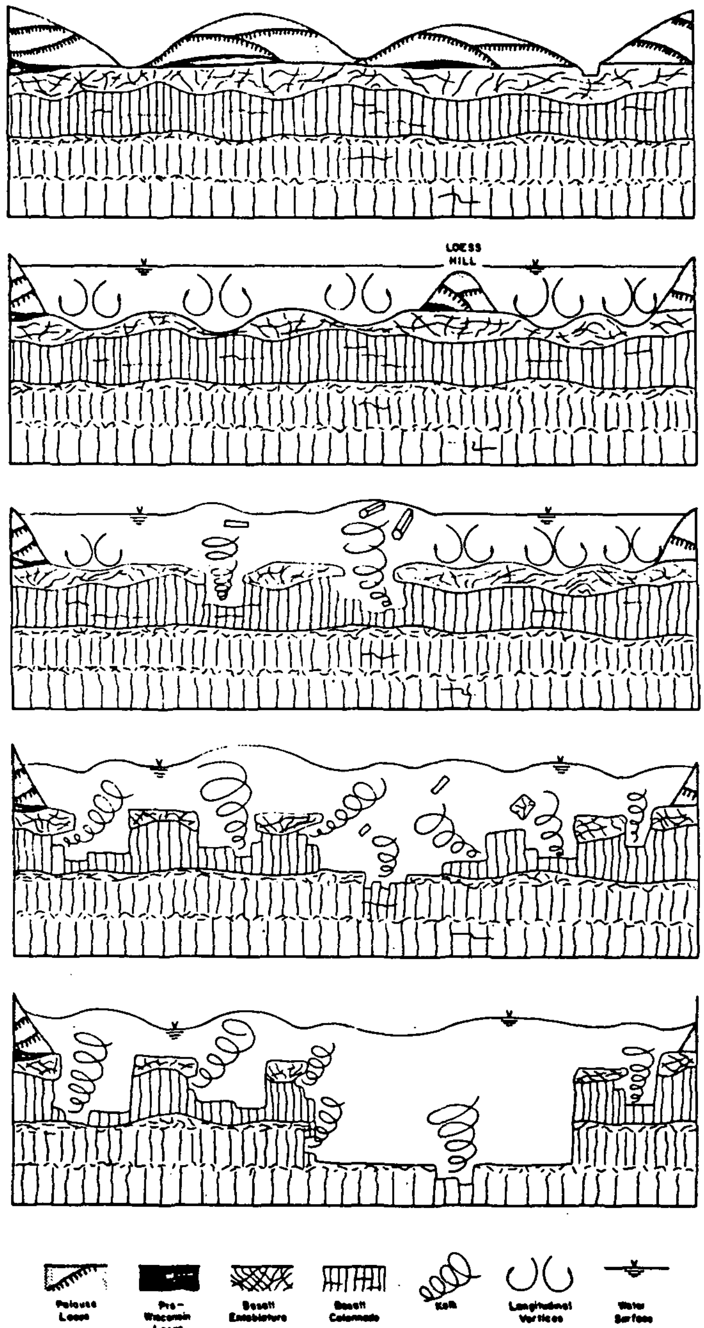

Figure 5.32. Hypothetical sequence of flood erosion for a typical scabland divide crossing. See text for discussion.

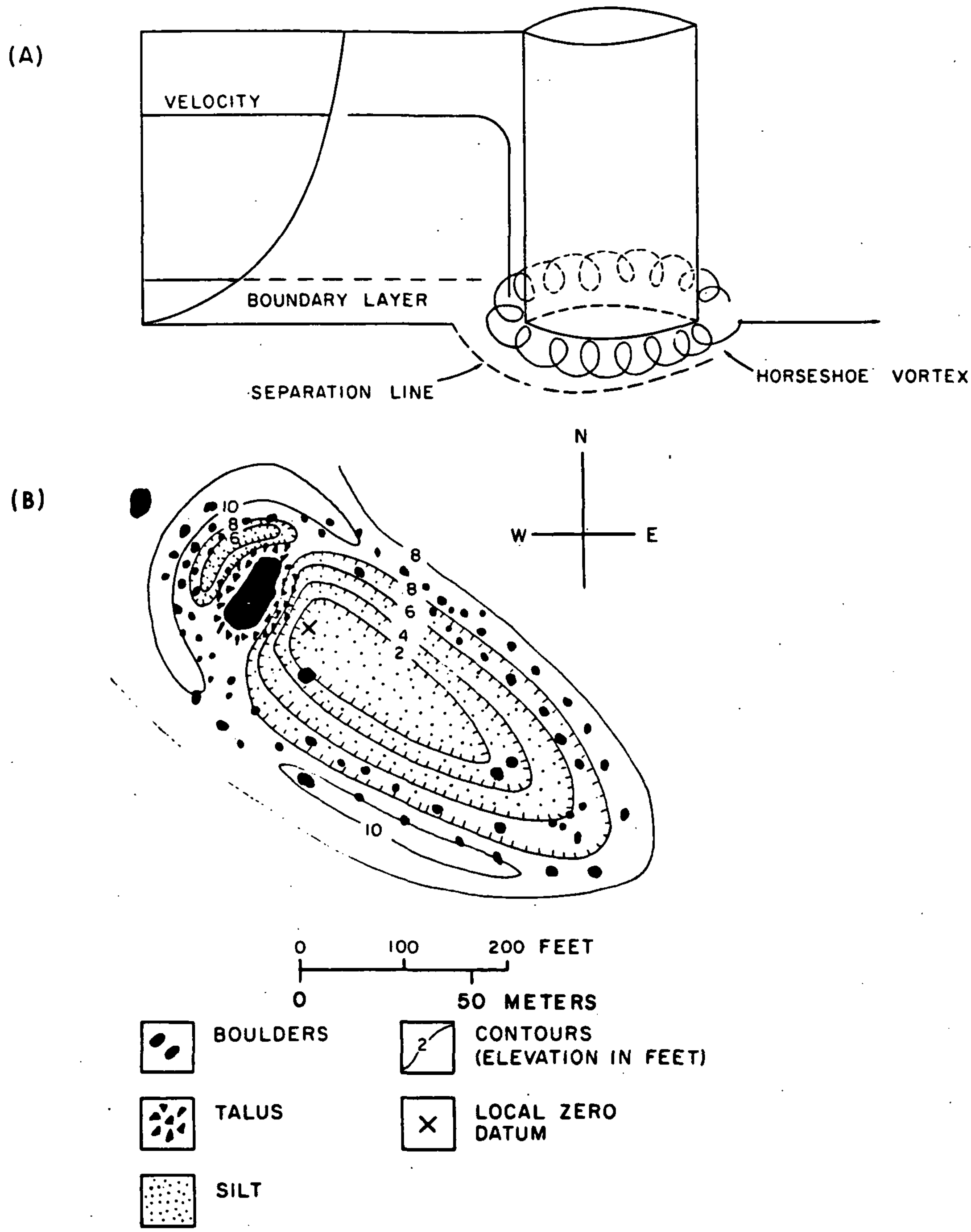

Figure 5.33. A. Formation of a horseshoe-vortex system at the front of a vertical cylinder mounted on an experimental flume bed (after Moore and Masch, 1963). B. Scour hole development near an 18 x 11 x 8 m boulder (Fig. 5.15) 2.5 km west of Rocky Ford Fish Hatchery, Ephrata, Washington (Baker, 1973a).

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 cuspate 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*

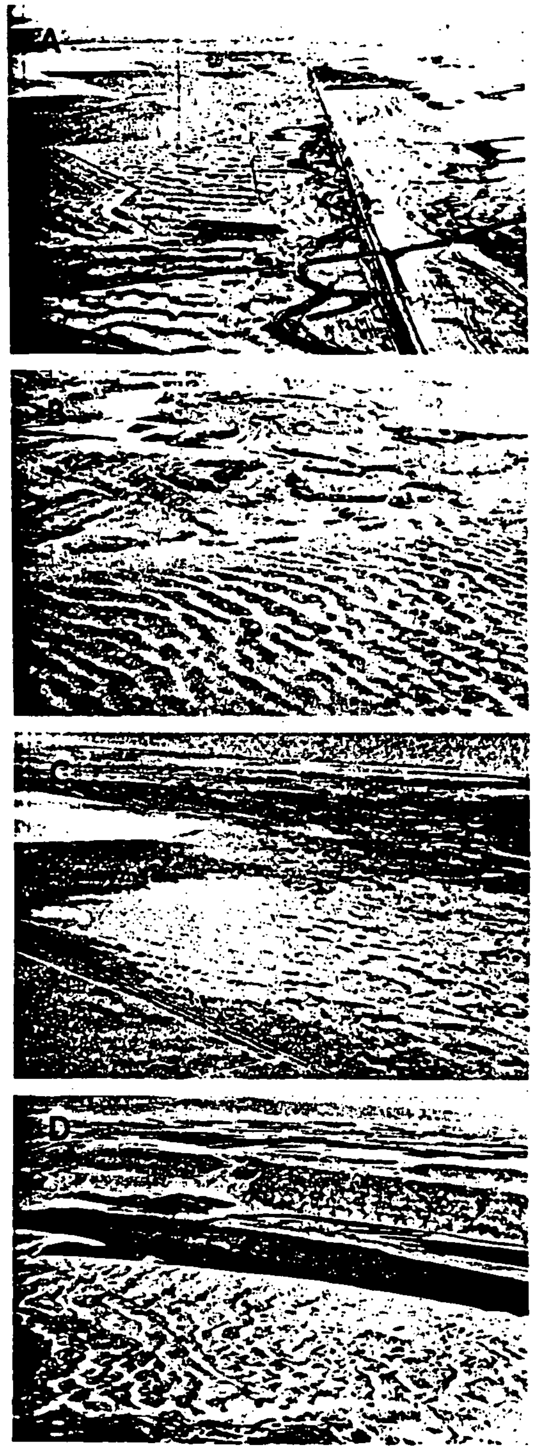

Figure 5.34. Typical sets of giant current ripples in the Channeled Scabland: A. Lind Coulee, B. Marlin, C. Artesian Lake, D. West Bar. Locations for these ripple trains are given in Baker (1973a).

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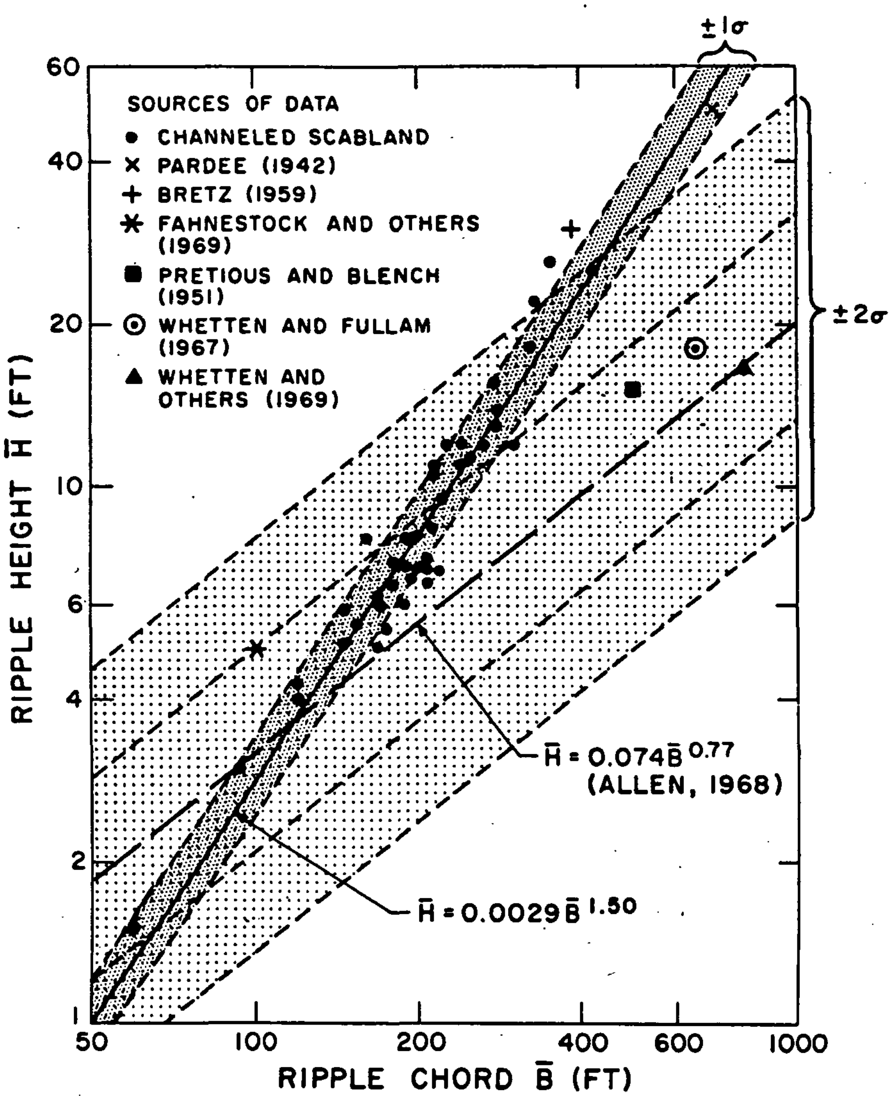

Figure 5.35. Logarithmic relationship of height as a function of chord for 40 sets of giant current ripples. The general relationship for large-scale current ripples determined by Allen (1968) is shown by a dashed line. Five additional measurements by various authors are also plotted. Standard errors 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.

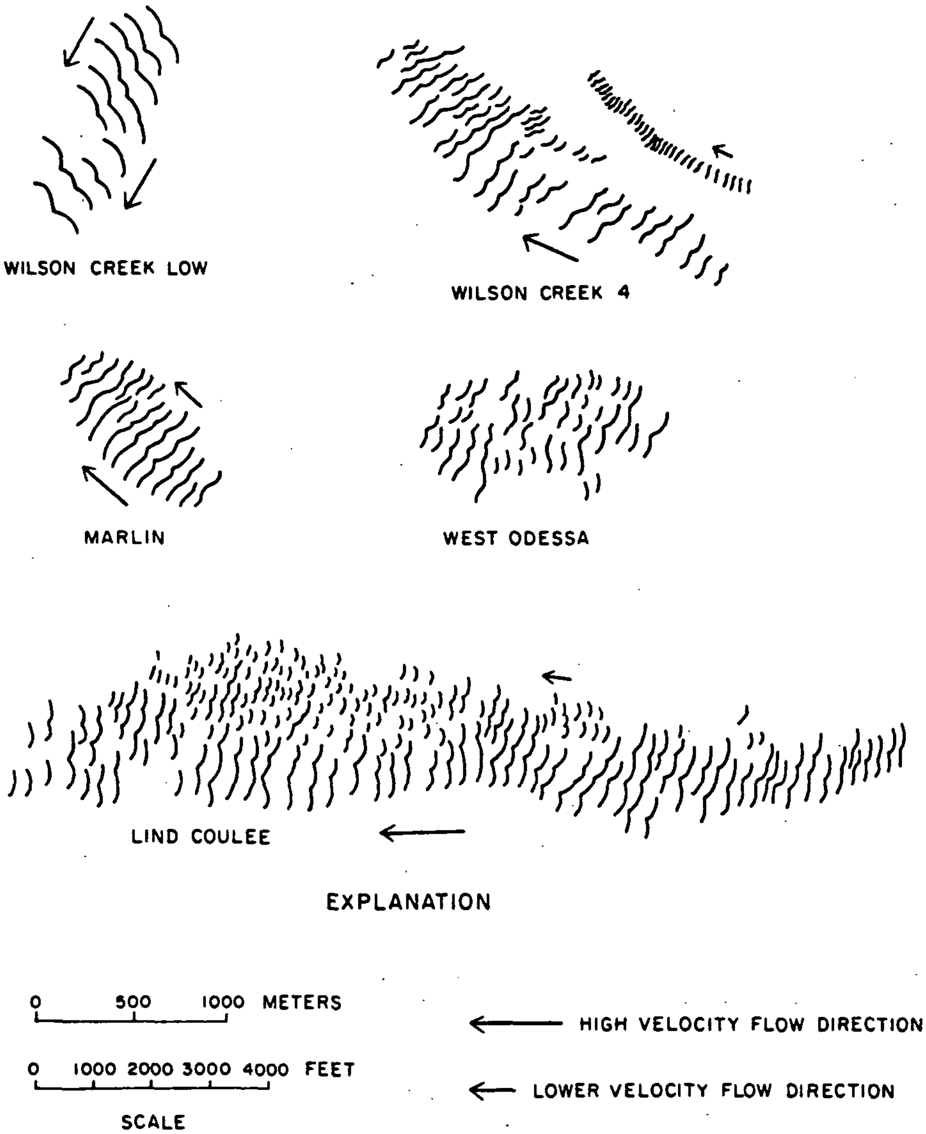

Figure 5.37. Giant current ripple trains viewed in plan. These patterns of ripple crests, traced off vertical aerial photographs, show a decrease in bed form size away from the thread of maximum flow velocity.

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 cuspate 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 (9.8×10^3 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:

$$\overline{H} = 24.6\ (DS)^{1.17} \qquad (5\text{-}10)$$

may be used within the standard error indicated

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

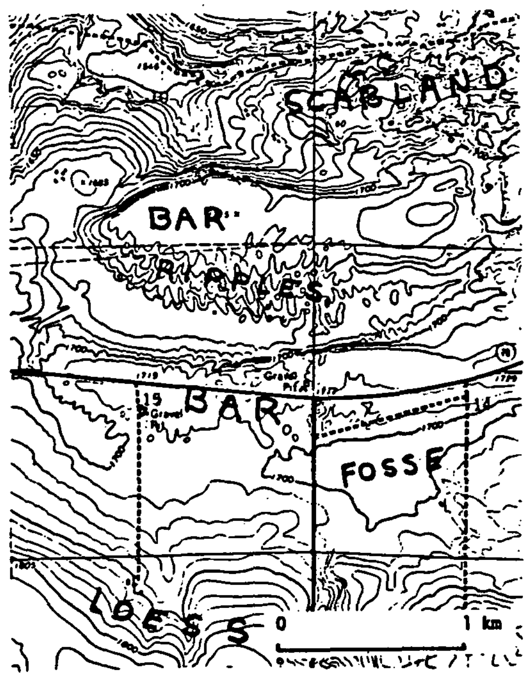

Figure 5.41. Grain-size distribution for sediment comprising giant current ripples in the Channeled Scabland. The numbers for the ripple sets are keyed to locations and descriptions in Baker (1973a, Appendix II).

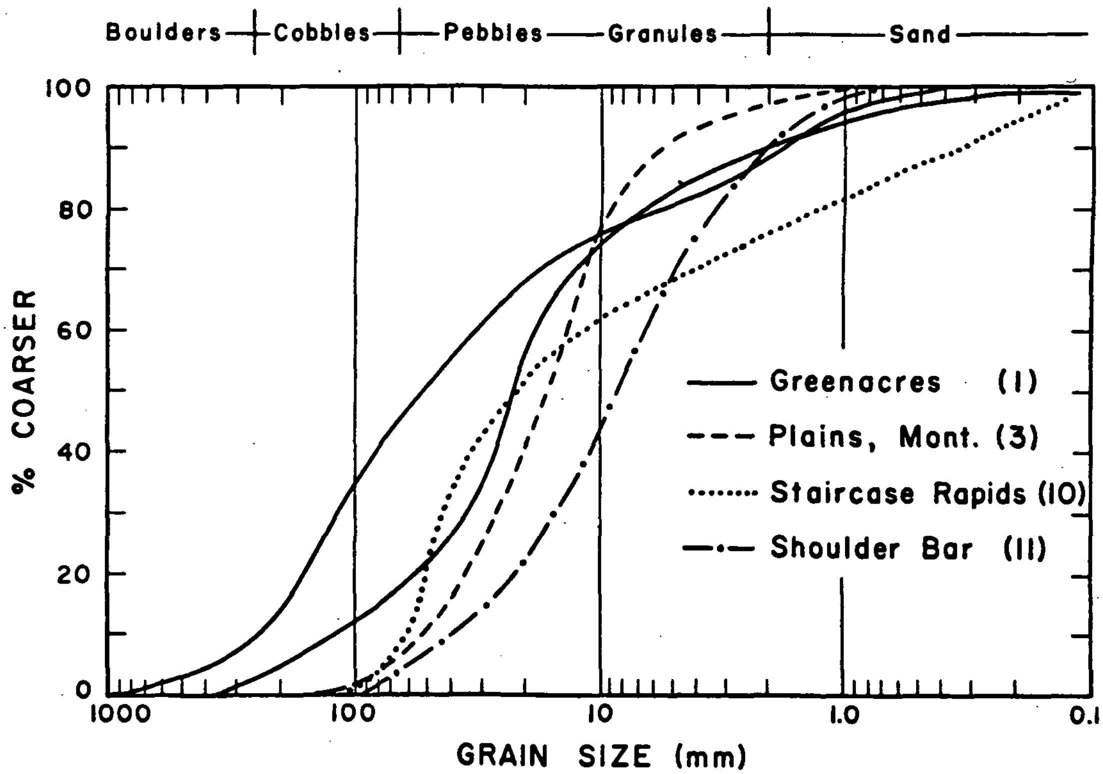

Figure 5.42. Foreset bedding exposed in a gravel pit cut through a Lind Coulee ripple oblique to the probable current direction. Apparent dips of the bedding vary from 10 degrees on the left to 25 degrees on the right portion of the photograph.

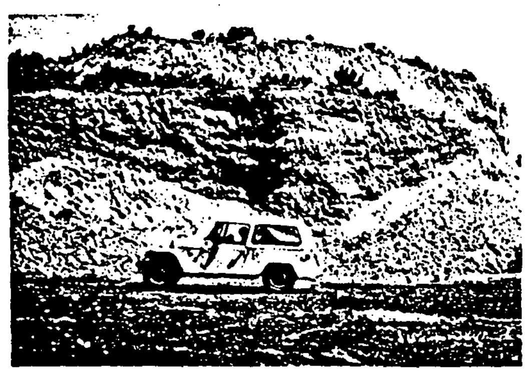

Figure 5.43. Trough-filling cross stratification in flood gravel exposed near Coulee City, Washington.

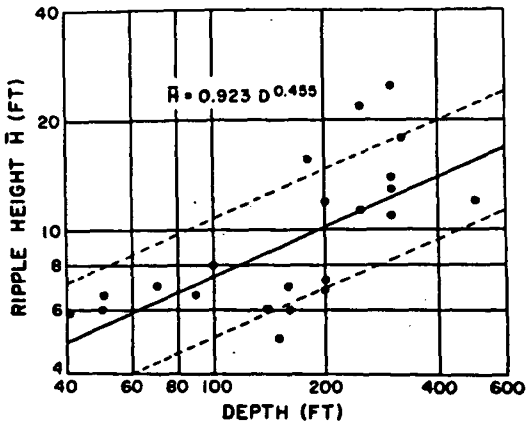

Figure 5.44. Logarithmic regression of ripple height as a function of depth. The dashed lines represent one standard error.

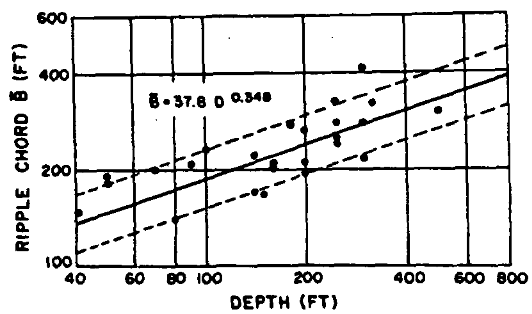

Figure 5.45. Logarithmic regression of ripple chord as a function of depth. The dashed lines represent one standard error.

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

$$\bar{B} = 393.5 \ (DS)^{0.66} \qquad (5\text{-}11)$$

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.

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)

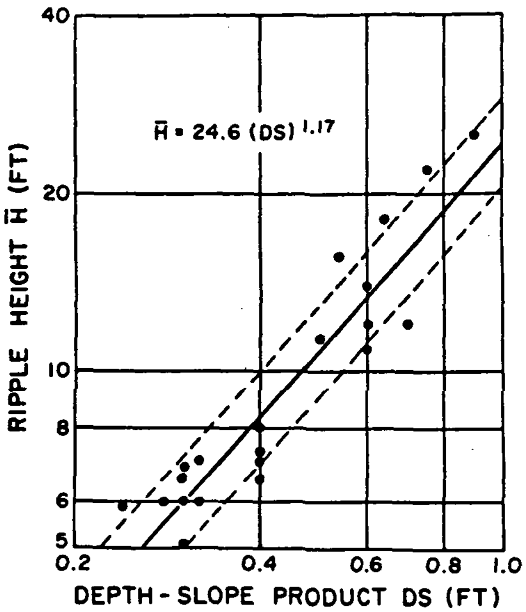

Figure 5.46. Logarithmic regression of mean ripple height as a function of the product of depth times slope. The dashed lines represent one standard error.

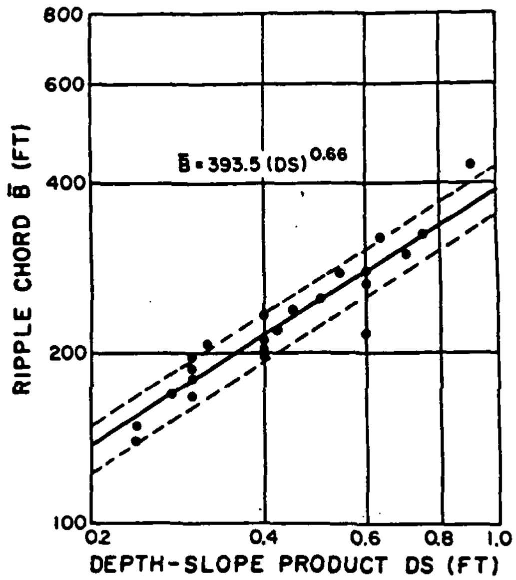

Figure 5.47. Logarithmic regression of mean ripple chord as a function of the product of depth times slope. The dashed lines represent one standard error.

through that subsection according to the formula

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

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:

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

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:

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

in which the units are as shown in Figure 5.50.

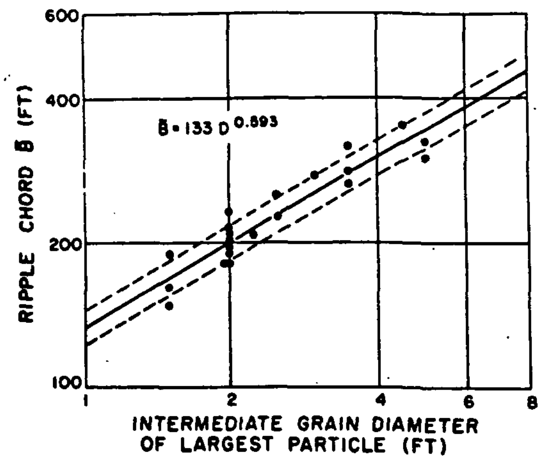

Figure 5.48. Mean ripple chord as a function of the maximum grain size found on the surface armor of the ripples. The dashed lines represent one standard error.

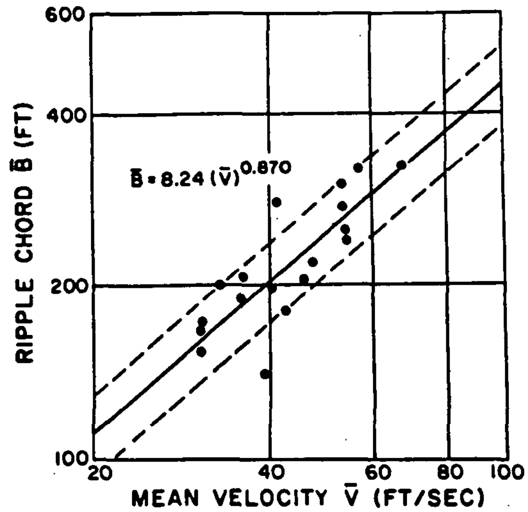

Figure 5.49. Ripple chord as a function of mean flow velocity (discharge velocity) as calculated by the slope-area method. The dashed lines represent one standard error.

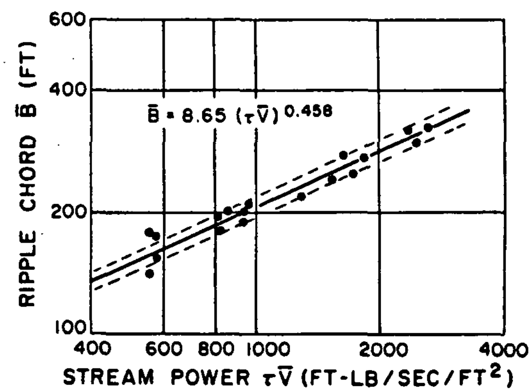

Figure 5.50. Mean ripple chord as a function of stream power. The dashed lines represent one standard error.

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."

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Chapter 6

Origin of The Cheney-Palouse Scabland Tract

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ABSTRACT

The Cheney-Palouse tract of the Channeled Scabland is the largest continuous tract of scabland in eastern Washington. The tract is composed of a varied assortment of bedrock erosional forms, loess islands and gravel bars. Prominent bedrock longitudinal grooves and inner channels formed by macroturbulent plucking erosion of the jointed rock. Loess island forms vary as a function of their position within the flow. The three major types (submerged, partially submerged, and subaerially exposed) created sedimentologic conditions and resulting bar forms distinct from one another. Other bar forms, notably expansion bars, account for most of the sedimentation in the tract.

In form and process, much of the Cheney-Palouse is analogous to a braided stream. First, the geometry of even the most complex reaches of the tract can be classified in the same manner as braided streams are classified. Second, the degree of development of smaller and topographically higher elements of the system is similar to observations made on the hierarchy of form and process in modern braided streams. Finally, the loess islands, although erosional, appear to have acted in the same manner as longitudinal bars in braided streams during the passage of a flood wave. They diverted the flood flows toward the banks and created zones of deposition in their lee.

INTRODUCTION

This paper is being published 55 years after the first interpretation of the Cheney-Palouse scabland tract by J Harlen Bretz (1923a, 1923b) and 40 years after a paper of the same title (Flint, 1938b) rejected Bretz' catastrophic flood origin of the region's landforms. The purpose of the present paper is twofold: (1) to demonstrate the general verity of Bretz' original interpretation, and (2) to show that the details of landform genesis in the tract are consistent with the dynamics of catastrophic flooding (Baker, 1973a, 1973b).

The Cheney-Palouse scabland tract is the easternmost system of flood-scarred channel ways in the Channeled Scabland (Fig. 6.1). The tract was first denoted as a single element of the Channeled Scabland by Bretz (1923b). It begins where Missoula Flood water spilled over the upturned northeastern margin of the Columbia Plateau near Spokane, Washington. The tract terminates 135 km to the southwest where the floods crossed the pre-glacial divide between the Snake and Palouse Rivers. This longest continuous scabland tract (Bretz, 1923b) is bordered on the east by undisturbed "Palouse hills" topography and on the west by loess topography interrupted locally by the major westward-flowing distributary coulees which carried Missoula Flood waters into the depositional basins of the western scablands.

Viewed from a LANDSAT image, the Cheney-Palouse appears to be an enormous braided stream with broad anastomosing channels separated by mid-channel islands. However, the anastomosing channels are scoured into the loess and underlying basalt bedrock. The midchannel islands are not depositional bars as one finds in braided streams. Rather, they are the eroded remnants of the once continuous loess cover. Nevertheless, we will show that the hierarchical arrangement of the channel elements is analogous to the hierarchy of forms described for braided streams by Williams and Rust (1969).

The Cheney-Palouse scabland tract is a rare terrestrial example of large-scale bedrock erosion by floods whose flow was confined neither by resistant channel walls nor by major geologic structures. We have previously cited this region as a major terrestrial analog to the Martian outflow channels (Baker and Patton, 1976).

PREVIOUS INVESTIGATIONS

Bretz (1923a) described the following distinctive features of the Cheney-Palouse tract: (1) the scarped loess islands with sharp upstream prows separated by scabland channels, (2) the elongation of the loess islands parallel to the overall trend of the tract, (3) gravel terraces which were almost everywhere associated with the lee side of the loess islands, and (4) large enclosed elongated rock basins which were created subfluvially by hydraulic plucking of the basalt bedrock. He also noted that 30 - 60 m of basalt must have been eroded from the level of the pre-glacial topography of the Cheney-Palouse. He concluded that only an enormous deluge could have created these landforms.

In the controversy that arose over the origin of the Channeled Scabland (see Baker, Ch. 1, this volume), the Cheney-Palouse tract played a significant role. The probable reason is that much of the tract exhibits relatively small-scale erosional and depositional features compared to those of the western scabland. Therefore, those investigators who felt that Bretz was teetering on the brink of catastrophism with his flood hypothesis sought to explain at least this segment of the scablands in terms of normal stream erosion.

Allison (1933) and Flint (1938b) produced the most detailed hypotheses to explain the origin of the Cheney-Palouse in terms of normal stream processes. Allison (1933) proposed that an enormous dam of floating ice bergs in the Columbia River Gorge was responsible for ponding of water supplied by moderate flooding of pro-glacial streams. This ice dam supposedly ponded water to over 300 m above sea level forming glacial Lake Lewis. As the ice dam grew headward, it locally diverted streams to create the unusual scabland relationships. Allison (1933, p. 683) also hinted that aggradation of proglacial streams eventually caused divide crossings and scabland erosion as far east as the Cheney-Palouse tract. Although Allison believed that there was a "Spokane Flood," he felt that it was of moderate size and that his combined mechanism of ice damming and flooding removed the interpretation of the Channeled Scabland from the realm of Bretz' "impossible" catastrophic flood.

Flint (1938b) using Allison's ice dam hypothesis as an impetus for more detailed study,

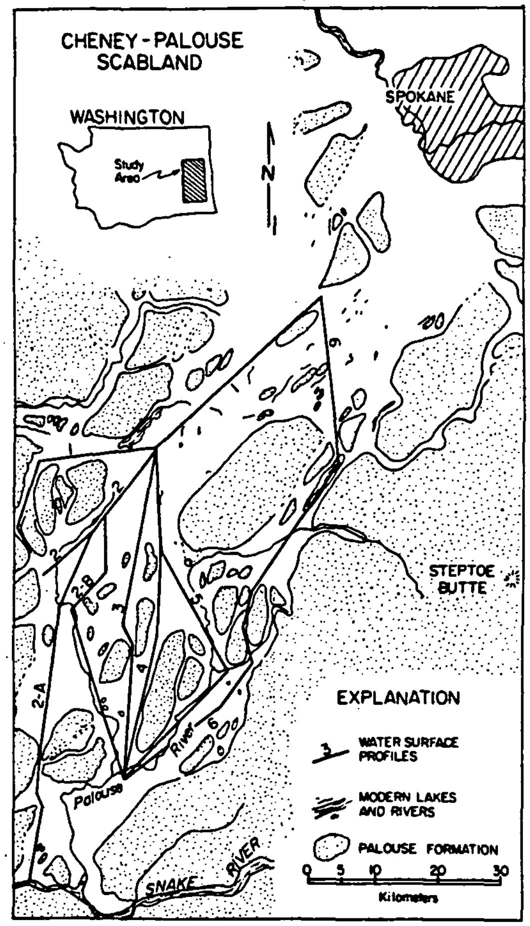

Figure 6.1. Highly generalized sketch map of the Cheney-Palouse scabland tract showing the location of various water-surface profiles that have been determined for the last major Pleistocene flood. Many of the small residual hills of Palouse Formation ("loess islands") have been omitted for simplicity.

thoroughly examined the Cheney-Palouse. Flint's hypothesis was that proglacial streams of normal discharge draining into Lake Lewis gradually aggraded their valleys as the lake level rose. Raised on their valley fills, the streams eroded the loess scarps and even spilled over preglacial divides. When the ice dam in the Columbia was finally breached, the relatively small proglacial streams re-excavated the fill leaving loess islands and various unpaired terraces. The streams then eroded into the underlying basalt, producing the scabland topography visible today.

Flint made detailed descriptions of the scarped loess islands which he interpreted as erosional remnants left by laterally planating streams. He was particularly intrigued by those loess islands that had several terraces attached to their upstream prows and downstream tails. The terraces do not occur along the loess island flanks. Flint also noted that other gravel "terraces" were only preserved at those localities where they were protected by upstream basalt knobs. Finally, several terraces had long gravel ridges on their surfaces oriented transverse to the flow which he could not explain.

Flint's (1938b) major addition to Allison's ice dam hypothesis was the creation of Lake Riparia. Flint thought that the aggradation of Washtucna Coulee caused the Palouse River to top its divide and establish a new course through Palouse Canyon to the Snake River. The sediment eroded during this process created a large fan delta and dammed the Snake River forming Lake Riparia upstream. The remnants of this fill were Mid-Canyon and Shoulder Bars. In short, Flint did not see any need to invoke a catastrophic flood because he believed that the bedrock erosion and sediment transport required to create the Cheney-Palouse tract was not extraordinary for proglacial streams operating for long periods of time.

The reinvestigation by Bretz and others (1956) refuted these alternative theories. They demonstrated that the erosional terraces were in fact constructional bars, many with enormous foreset beds that could not be rationalized in terms of slowly aggrading streams of normal discharge. More significant was the recognition that the gravel ridges originally mentioned by Flint were giant current ripples similar to those described by Pardee (1942) in the basin of Lake Missoula itself. Bretz and others (1956) reiterated the original conclusion that the erosional bedrock forms, loess islands, depositional bars, and divide crossings of the Cheney-Palouse tract were created by catastrophic flooding.

BEDROCK EROSIONAL FORMS

A variety of erosional forms characterize the scabland topography on the exposed Yakima Basalt. These include the numerous potholes, large elongate scour holes, longitudinal grooves, cataracts, and deep narrow inner channels winding the entire length of the Cheney-Palouse.

The longitudinal grooves are particularly well developed in the converging channel south of Sprague Lake between the Karakul Hills loess islands and the town of Lamont (Fig. 6.2, top, center). The widest of these grooves, now occupied by Palm Lake, is just under 300 m across. The grooves can be up to 15 to 25 m deep. The Palm Lake grooves occur as two bands which extend nearly across the channel. Both the upstream and the downstream bands are about 1.5 to 2.5 km long parallel to the flow direction. The upstream grooves are cut into the entablature surface of a basalt flow about 10 to 15 m above the level of the underlying basalt flow in which the downstream grooves are incised. The upstream grooves are spaced approximately 500 to 700 m apart, and many have been eroded headward through the basalt flow which forms the intervening plateau surface. These grooves now appear as dry canyons separated by isolated basalt mesas. The downstream grooves are spaced about twice as far apart and usually terminate upstream at a small cataract which may be up to 20 m high. These grooves do not line up with the upstream set and probably developed independently. Although some of the grooves can be traced downstream, the majority of these longitudinal forms disintegrate into a maze of butte-and-basin scabland topography.

As pointed out by Baker (1973b), the scabland grooves are analogous to grooves found in other bedrock stream channels and in experimental flume studies of simulated bedrock erosion (Shepherd, 1972; Shepherd and Schumm, 1974). Shepherd (1972) noted that in an essentially straight

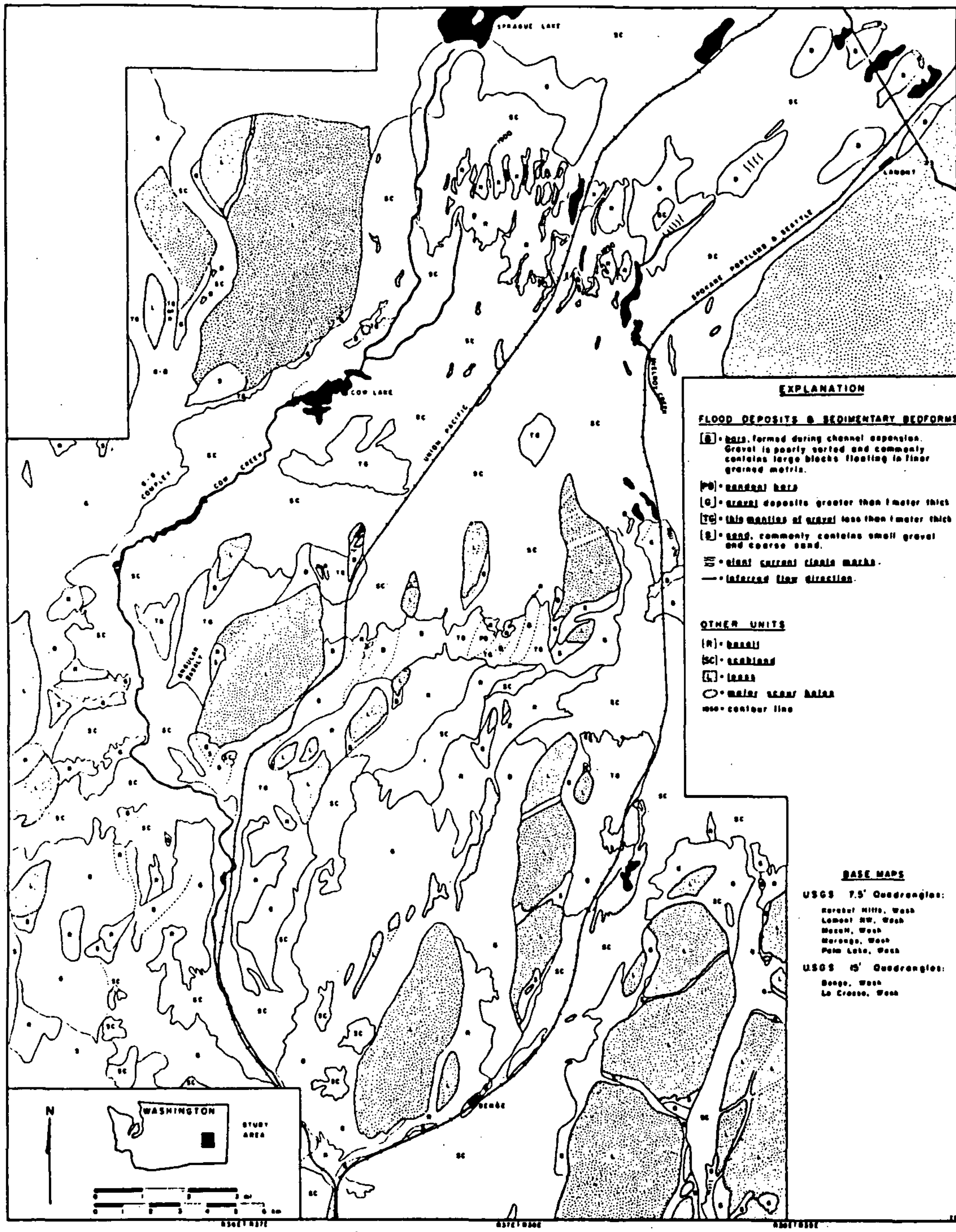

Figure 6.2. Geomorphic map of the central portion of the Cheney-Palouse scabland tract between Sprague Lake (north) and Benge (south). The map illustrates typical relationships between the various channel elements.

channel, a sequence of erosional bed forms developed on simulated bedrock. This sequence began with longitudinal lineations which became enlarged into prominent longitudinal grooves. Although potholes and erosional ripples developed with the lineations, with time, the grooves became the dominant bed form. Shepherd (1972) hypothesized that secondary circulation cells in the flow were responsible for groove erosion. Furrows were produced where the vortices attached to the bottom, and ridges were left where separation zones occurred. Shepherd (1972) varied the flow characteristics in the flume to determine which factors were most important in groove formation. He found that increases in slope and water discharge had little effect on accelerating the erosion process, but that the grooves rapidly grew when the sediment discharge rate was increased. This is because the experimental bedrock was a dense clay-sand mixture which was extremely cohesive (Shepherd, 1972). Therefore, abraison, not plucking, was the most important mechanism for eroding the experimental grooves.

Several differences should be noted between Shepherd's experimental grooves and the Palm Lake grooves. First, it is probable that the scabland floods, given their enormous discharges and their hypothesized macroturbulent nature, were underloaded with sediment. Second, the jointed basalt bedrock lends itself to plucking and quarrying, a fact that has been noted by all scabland investigators since Bretz (1924). Third and last, the experimental grooves evolved more or less simultaneously along their length while there is ample evidence that the scabland grooves evolved by cataract recession.

Shepherd and Schumm (1974) noted that with time the experimental bedrock grooves coalesced, and a dominant bedrock inner channel was formed leaving remnant paired bedrock benches as evidence of the old channel floor. Cow Creek (Fig. 6.2, west side of map) presently occupies such an inner channel. It is the deepest channel way in the Cheney-Palouse and is bordered by erosional scabland to its juncture with the Palouse River at Hooper. Other prominent inner channels in the Cheney-Palouse tract include Bonnie and Rock Lakes and the Palouse Canyon where the Palouse River crosses the former Palouse-Snake divide. These inner channels are significant in localizing the bedrock scour and causing the greatest degree of scabland development.

The long profiles of these inner channels are highly irregular. The Rock Lake system (profile 6, Fig. 6.1) has several lakes along its profile, indicating the presence of several enclosed rock basins. The reconstructed high-water surface profile (Fig. 6.3) shows that many of the larger basins coincide with steep energy gradients. Pronounced constriction of the flow induces greater erosion, as discussed by Baker (1973a, p. 15-16).

One of the most abrupt constrictions in the Cheney-Palouse tract occurred at Staircase Rapids, just north of Washtucna (see Bretz and others, 1956, p. 1000-1003). The water-surface profile (Profile 2A, Fig. 6.1) shows the pronounced ponding of water in the Rattlesnake Flats area (Fig. 6.4). The relatively subdued topography of Rattlesnake Flats contrasts sharply with the scabland and cataracts of Staircase Rapids. The flood water-surface gradient through the rapids averaged about 12 m/km (Fig. 6.4).

The influence of bedrock structure on scabland erosional forms is especially evident on the Palouse-Snake divide crossing. The major drainage lines in the unaltered Palouse Hills trend northeast to southwest in this area of the Cheney-Palouse (Lewis, 1960), and it appears that initially the flood followed the major valleys as it cut across the divide. There are several lines of evidence to support this. First, the gross orientation of the divide crossing is from the northeast to the southwest, an orientation that is also reflected in the smaller ancillary divide crossings. For example, a divide crossing northeast of Nunamaker farm (Sec. 24, T.14N., R.35E.) perfectly parallels the unaltered preflood drainage. Second, all of the remnant loess islands within the divide crossing are oriented in the same direction. This includes not only those loess islands along the far eastern margin of the crossing, but also two small loess islands directly east of the H U Ranch cataract (Sec. 27, 28, 33, 34, T.14N., R.36E.). There is, however, only one major bedrock cataract oriented in this direction and that is the cataract containing Devil's Lake (Sec. 9, 16, 17, T.14N., R.37E.). The major cataracts including H U cataract, Palouse Canyon, and the cataracts containing Wind Lake and Deep Lake all trend approximately east-southeast. A second set of smaller cataracts trends southeast.

We suggest that the initial flood flows across the divide followed the loessial topography which, in turn, was oriented according to the prevailing wind direction (Lewis, 1960). Subsequent bedrock scour was then localized by the fracture set oriented nearly perpendicular to the initial flow direction (Trimble, 1950). The weaker rock in the fracture zones were preferentially quarried during the flood. A feedback mechanism can be envisioned in which these zones of bed relief perpendicular to the flow added to the turbulence and accelerated the erosive processes.

LOESS REMNANTS

The most conspicuous macroforms of the Cheney-Palouse scabland tract are the erosional remnant loess islands. Originally described by Bretz (1923b), these streamlined hills generally have sharp upstream prows, steep faceted flanks, and long tapering tails. In the Cheney-Palouse, there are three distinct varieties of loess islands. The detailed reconstruction of flood high-water surface profiles allows us to distinguish forms that were (1) submerged beneath the flow, (2) partially submerged by several major divide crossings, or (3) unsubmerged and exposed above the

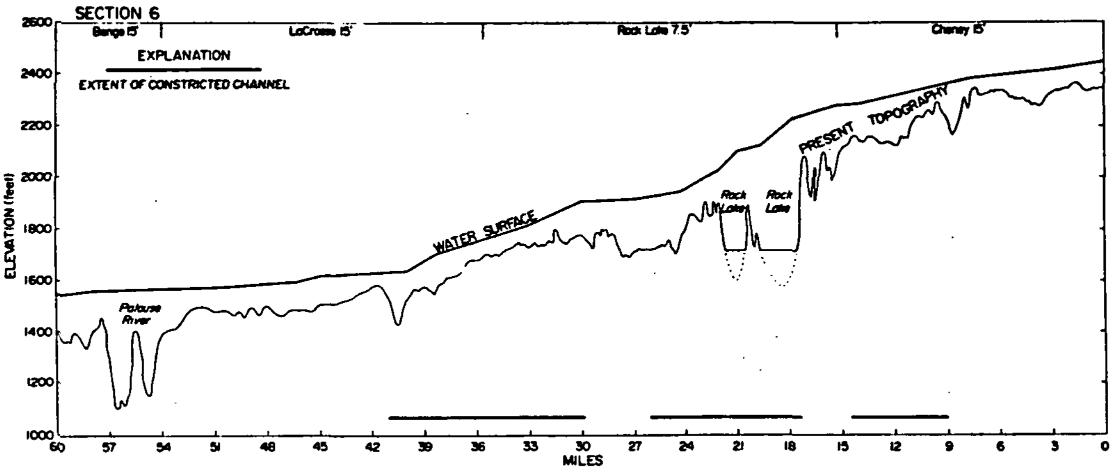

Figure 6.3. Profile 6 through the eastern part of the Cheney-Palouse tract (see Fig. 6.1 for location). Note the steepening of the water-surface gradient through constricted reaches.

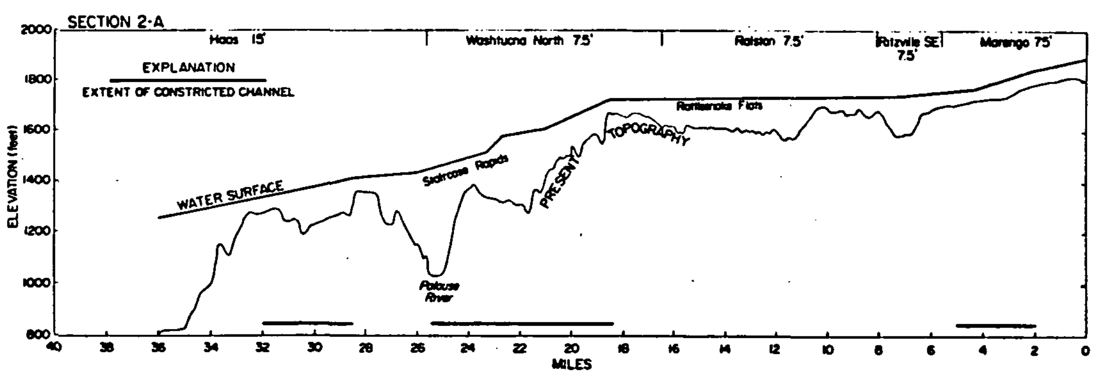

Figure 6.4. Profile 2-A through the western part of the Cheney-Palouse tract (see Fig. 6.1 for location). Note the pronounced ponding of water at Rattlesnake Flats and the steepening through Staircase Rapids.

flow. These major differences in position of the loess islands within the flow also caused distinct sedimentologic variations in the style of deposition and position of the gravel bars attached to the islands.

Submerged Remnants

Subfluvially eroded loess islands have streamlined shapes similar to airfoils (Baker, 1973b). They are the smallest of the loess forms present in the Cheney-Palouse (Fig. 6.5A). Because these forms were completely covered by the flood, the flanking scarps are not as well developed, and all preflood drainage topography on the tops of the islands was obliterated. These islands have gravel bars attached to their tails in much the same manner that wind-shadow dunes form in the lee of flow obstructions. In this case, the gravel bars drape over the tail of the eroded loess form, and the shape of the resulting streamlined hill is partly influenced by the loess island and partly by the gravel bar. This relationship can be seen in several roadcuts. The gravel thinly veneers the

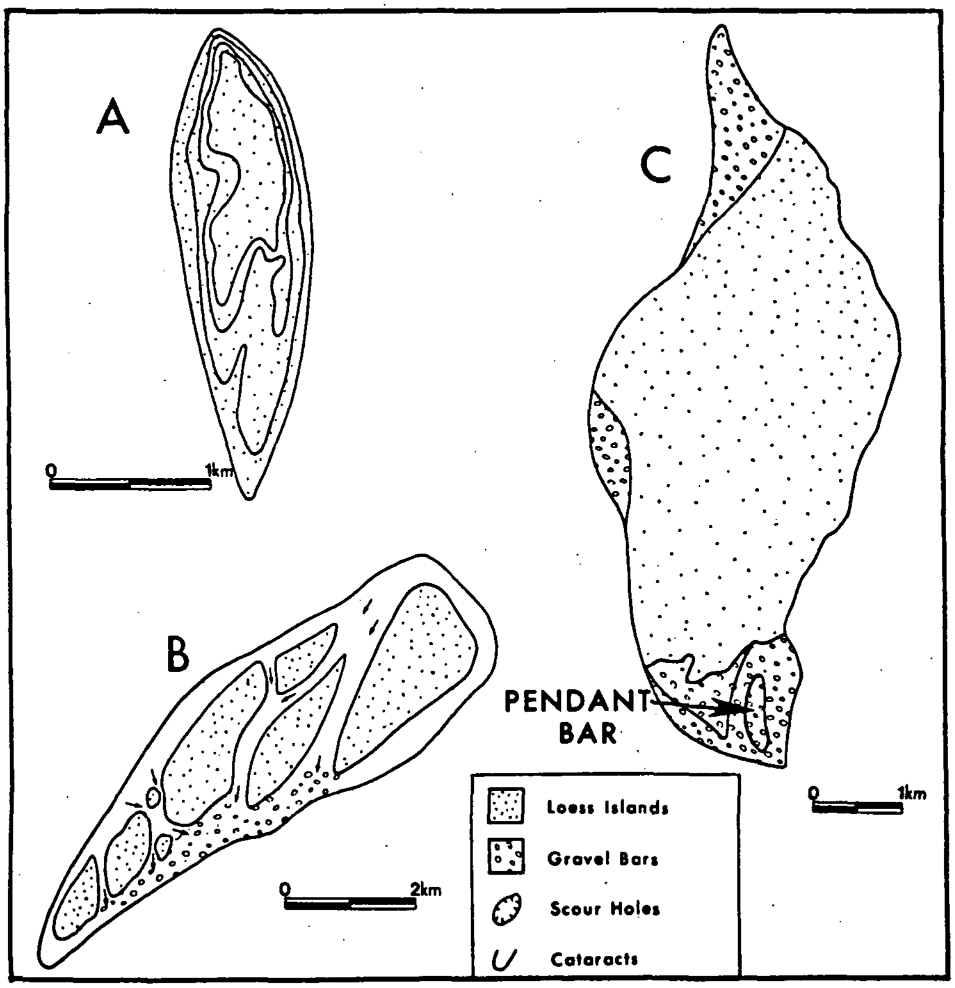

Figure 6.5. Loess island forms in the Cheney-Palouse scabland tract illustrating the three major morphologic types. A. Subfluvially eroded loess island located in the Karakul Hills (Secs. 2, 11, and 14; T. 19N., R. 36E.). B. Partially submerged loess island near Amber Lake (Secs. 1, 2, 11, 12, 15; T. 21N., R. 40E.). Arrows denote the major divide crossings and the gravel pattern indicates a bar. C. unsubmerged loess island immediately northwest of Marengo (left center of Fig. 6.2).

upstream portions of the residual hill and thickens downstream toward the "tail" of the streamlined form which is composed entirely of transported sediment. Because of the high degree of streamlining, it is often difficult to determine where the bar attaches to the loess island. The problem is made more difficult by the cover of late Pleistocene-Holocene loess draped over the entire feature.

The gravel "tails" on these loess islands are composed predominantly of cobble gravel in a matrix of granule gravel and coarse sand. The deposits are generally finer-grained than most main-channel facies in the Channeled Scabland. We suggest that deposition probably took place during waning flood stages after the loess island had become a significant obstruction inducing a zone of flow separation in which sediment could accumulate.

Partly Submerged Remnants

A second variety of loess island includes those transected by one or more major channels still shallow enough not to have eroded into the underlying basalt (Fig. 6.5B). Many such channels eroded large re-entrants on the downstream margins of the islands which were later partially filled with sediment. Consequently, these loess islands have the thickest accumulation of flood sediment. Their deposits exhibit large-scale avalanche cross-bedding, cut-and-fill sequences, soft sediment deformation structures, and a rapid vertical variation in grain size. It is not uncommon for a layer of open-work gravel to be succeeded by a layer of coarse sand followed by another layer of gravel. Light colored beds made up of reworked loess are also common. This mixture of fine and coarse sediment indicates that an extremely wide range of grain sizes were in transport and that perhaps pulses in the flow velocity were responsible for the fluctuations in grain size through time. The mode of deposition at these loess islands was probably similar to the process of deposition at the eddy bars which developed at the mouths of pre-flood tributaries to the Channeled Scabland (Baker, 1973a). However, water and sediment entered the loess island re-entrants from divide crossings that were at different angles to one another and from the main channel. The combination of these several flow directions thus created large eddies which accumulated flood debris.

Unsubmerged Remnants

Loess islands whose crests were not topped during the Missoula Flood are characterized by the well-developed steep flanking scarps which truncate and behead the pre-flood drainage systems still evident on their crests. Many of these islands were crossed by small distinct channels during the last flood, but these channels had little effect on the overall morphology and sedimentation. The islands have well-developed quadrilateral shapes; many very similar to the rhombic or diamond shape of longitudinal bars typical of braided streams.

Although the loess islands are dominantly erosional forms, some of their streamlined character is caused by deposition of gravel bars both at the prow and downstream in the lee of the islands (Fig. 6.5C). The bars are easily distinguished from the residual loess by their lower elevation and by their smooth flat surfaces, devoid of any stream network development. An example is the prominent bar which is attached to the prow of the Marengo loess island, perhaps because of the upstream flow stagnation plane. The bar is composed of angular boulders up to .75 m in intermediate diameter. This is a much larger grain size than that of the bar at the lee side of the island which is composed of cobble and granule gravel. No stratification was evident in exposures of either bar.

Discussion

The bars associated with these last two types of loess remnants were originally interpreted as terraces by Flint (1938b). Although Bretz and others (1956) did not make a detailed study of these surfaces, they did hypothesize that the smooth "terrace" surfaces might be the result of gravel bars, buried resistant soil horizons, bedrock ledges or incomplete flood incision. They further noted that eight of Flint's (1938b) examples of terraced loess islands were remnants which had smaller secondary channels cut through their divides. Our study supports the interpretation of Bretz and others (1956). Many of Flint's "terraces" were gravel bars deposited in the re-entrants described for the second loess island type. Investigation of several other terraced loess islands showed that bedrock ledges do occasionally form conspicuous terrace-like forms on some loess islands. Finally, the prominent loess island just east of Macall siding (Fig. 6.2, right center) has a prominent flat surface on its downstream end caused by an exhumed resistant petrocalcic horizon. The petrocalcic horizon (caliche) caps a pre-Wisconsin flood deposit (See Baker, Ch. 2, this volume). It is quite probable that here the Palouse Formation was deposited over the gravel bar of a pre-Wisconsin flood.

GRAVEL BARS

Gravel bars in the Cheney-Palouse include pendant bars, expansion bars, and the previously described bars associated with loess islands. In general, the gravel bars in the eastern scablands are smaller than those to the west. The major reason for this is the lateral spreading of Cheney-Palouse flood water, resulting in lower flow depths than those attained in the great coulees of the western Columbia Plateau. The exceptions to this trend are Staircase Rapids Bar at Washtucna and Shoulder and Mid-Canyon Bars on the Snake River. These bars all formed downstream of local flow constrictions.

Pendant Bars

Pendant bars are not restricted to any particular geomorphic setting within the Cheney-Palouse. They occur in channels of all sizes, although they are rare in the deepest scoured channels. They occur most commonly along the margins of the flow and in smaller channels where resistant basalt knobs created the necessary flow obstructions from which the bars could accrete. One of the largest pendant bars in the Cheney-Palouse is adjacent to a scour hole southeast of Rock Lake. An excellent exposure at the toe of the bar demonstrates the large-scale avalanche beds that typically occur within these bars.

Locally, a single resistant basalt flow provided points of flow separation and reattachment, allowing a string of bars to form across the channel. Examples of this can be seen along the Ritzville Macall Road east of the Marengo loess island where numerous bars are attached to the downstream step created by a resistant basalt flow (Fig. 6.2, center). Bar deposition at this location was also favored by the transition at this point from upstream constrictions between loess islands to a major channel expansion that probably reduced the flood velocity. A similar situation occurred northwest of the town of Lamont (Fig. 6.2). Gravel pits show that the pendant bars are made up of large basalt blocks, many of which still have a polygonal columnar structure. Thus, the boulders were probably transported only short distances by the macroturbulent suspension mechanism described by Baker (1973a, 1973b). On the other hand, the bars also contain large granitic boulders that could have been transported only from the Medical Lake area 40 km to the northeast.

The pendant bars in the Cheney-Palouse tract are relatively small when compared to bars in the westward flowing distributaries such as upper Crab Creek and Wilson Creek. Although Cheney-Palouse pendant bars may extend downstream 1.5 km or more from their points of attachment to flow obstructions, the gravel is usually 10 m or less in thickness. This contrasts with the 30 m thick pendant bars reported by Baker (1973a) in Upper Crab Creek. Again, this reflects the relatively shallow flow depths of the Cheney-Palouse scabland tract.

Expansion Bars

Expansion bars are widespread gravel deposits immediately downstream from channel expansions. They often are found where several small channels exit from an assemblage of loess islands. The flow expansion as well as the shadowing effect of the loess islands created a low-velocity zone in the flood flow where deposition was favored. Other expansion bars formed downstream from cataracts or simply where the channel was unusually wide. An example lies south of the Karakul Hills loess islands where several small channels flow out onto a wide plain (Fig. 6.2, upper left). The thin character of this deposit can be seen downstream where basalt crops out only 3 m below the surface of the bar. This bar can be traced downstream to where it is truncated by scabland on the north side of Cow Creek. An example of an expansion bar downstream from a cataract is the gravel accumulation on the east side of the channel which terminates immediately northeast of Benge (Fig. 6.2, bottom). The bar parallels the loess island which forms the eastern side of the channel. At the upstream end of the bar, near the cataract, large slabby basalt boulders up to 3.5 m in intermediate size litter the surface. Three kilometers downstream, the intermediate grain size of the largest boulders has decreased to about one meter, but the roundness of the boulders has only slightly increased. The bar is extremely thin, and, immediately to the west, in the center of the channel, basalt crops out a few meters below the surface of the bar.

Although these bars are thin, they can cover significant areas, up to 50 km². The overall extent of these bars is obviously controlled by the channel geometry. Where abrupt constrictions occur, the bars are terminated, such as south of Sprague Lake where a channel converges between two loess islands. Also, where the bars extend into the major channels such as Cow Creek, they are abruptly terminated.

Giant current ripples are fairly common on pendant and expansion bars. In addition, such ripples may form at the downstream ends of scour holes. The ripples in these locations tend to migrate up the adverse slopes of the scour holes. A prominent example is the Macall ripple field (Sec. 18, T.18N., R.38E.) located in a scour hole immediately east of the Marengo loess island. These ripples occur on extremely thin gravel fills. They are really starved ripples, since the troughs of the ripples may be only a meter above bedrock.

Sedimentary Characteristics

A study was made of the distribution of the largest boulders in bars and on bedrock surfaces in the Cheney-Palouse. Unlike the predominant basalt boulders, boulders of granitic composition can be attributed to a known source area at the northern end of the tract, and these boulders were measured wherever they were found. The results indicate no systematic variation in grain size when the Cheney-Palouse is considered as a single unit. This is not surprising because most of the sediment was probably locally derived, and, therefore, the maximum size in any deposit is less a function of distance of transport than it is of local current velocity, turbulence, and joint spacing in the basalt. The largest boulder, 4 x 3 x 3 m in size, was found at midlength in the Cheney-Palouse approximately 12 km north of Benge. The boulder is one of several large basalt blocks deposited on a scarified basalt surface.

Within a single flood bar, there may be a downstream decrease in grain size. Gravel on the previously described expansion bar in the channel north of Benge abruptly decreases in grain size downstream. On the other hand, a large pendant bar in the channel just east of the Marengo loess island has boulders in its downstream tail which are only slightly smaller than those immediately downstream from the basalt flow to which the bar is attached. Expansion bars may be expected to show evidence of hydraulic sorting, since the flow conditions varied along the reach in which they were deposited. Pendant bars, on the other hand, are generally smaller in areal extent, and longitudinal variation in flow conditions is not required for their formation. Therefore, current sorting and large grain size variation would not be evident along their length.

One might hypothesize that granitic boulders derived from the Medical Lake area to the north would decrease in size down the tract as a function of selective sorting and perhaps breakdown of the larger sizes. Our reconnaissance data indicate that the grain size of the granite boulders does not change radically downstream. The largest granite boulder found in a Cheney-Palouse scabland deposit has a long axis of 270 cm and lies south of Rock Lake near Ewan (Bretz and others, 1956). Near Marengo, there is a granite boulder 190 x 120 x 70 cm in size, and Bretz and others (1956) report granite boulders having long axes of 165 cm in the Cow Creek scabland just north of Hooper and in Shoulder Bar in the Snake River. Therefore, from all the available data, it does not appear that there is a rapid downstream decrease in size, in sharp contrast to that reported for the Ephrata Fan in the western scablands (Baker, 1973a). Our hypothesis is that along the deepest channels, the bedrock inner channels, the velocities and competence of the flow were undiminished from one end of the

Cheney-Palouse to the other. This is supported by the nearly uniform water surface profile for the main channels, like Cow Creek (Fig. 6.6). In the western Channeled Scabland, abrupt expansions of major channels, such as the Lower Grand Coulee flowing into the Quincy Basin, caused rapid reduction in stream competence and rapid sediment deposition.

DISCUSSION

When viewed in detail, the Cheney-Palouse appears to be a complex landscape. This is the result of the grouping and superposition of the three primary forms: (1) loess islands, (2) gravel bars, and (3) erosional bedrock scabland (Fig. 6.7). The organization of these primary forms in the Cheney-Palouse can be classified in the same manner as other complex fluvial landscapes such as sandur plains (Church, 1972) and braided streams (Williams and Rust, 1969). The classification demonstrates that the landforms in the Cheney-Palouse form predictable geomorphic assemblages in many aspects similar to modern fluvial systems.

The Karakul Hills loess island assemblage (Fig. 6.8) is analogous to a single spool or diamond bar on a sandur plain (Church, 1972). Isolated from other loess island groups by major scabland channels, the Karakul Hills assemblage is dissected by a sequence of channels, each at higher elevations, smaller and less well developed. The primary zone of deposition is downstream from the loess island group similar to the common pattern for longitudinal bars in braided streams.

By arranging several of these loess island-bar sequences together, more complex geometries can be created such as the assemblage of loess islands northeast of Sprague (Fig. 6.9) or the loess islands and bars at Willow Creek near La Crosse originally described by Bretz (1928b, p. 648).

The above examples are for simple geometries and for fairly limited areas within the Cheney-Palouse. A more complex assemblage comprises the entire center of the map in Fig. 6.2. This particular loess island grouping is in the center of the Cheney-Palouse tract and is bordered on the east by the Cow Creek channel and on the west by the unnamed channel that includes Twelve Mile Lake. The two channels converge and form the southern boundary at Benge. Thus, the loess islands are preserved in a reach where the dominant channels have diverged. The reach covers an area 24 km long and 16 km wide. Directly south of Benge, the convergence of these channels has obliterated all of the loessial topography, and the tract consists entirely of scabland. We consider the two channels to be first-order channel elements (Williams and Rust, 1969). They are almost entirely scabland with only minor sedimentation along their flanks.

Within the area surrounded by the two first-order channels are three distinct levels of smaller

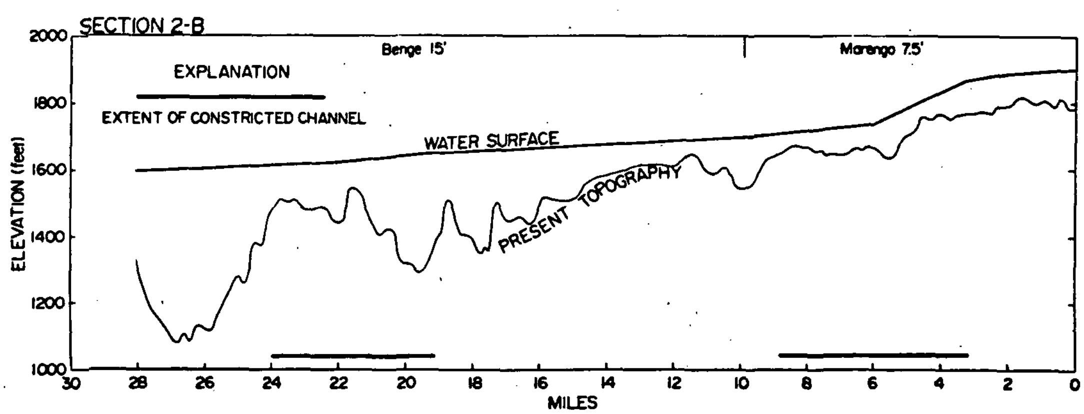

Figure 6.6. Profile 2-B along Cow Creek, a prominent inner channel of the Cheney-Palouse tract (see Fig. 6.1 for location).

Figure 6.7. Geomorphic map of the Macall area in the central Cheney-Palouse tract.

PLATE 1. Sand dune field at the Potholes Reservoir. North is to the left.

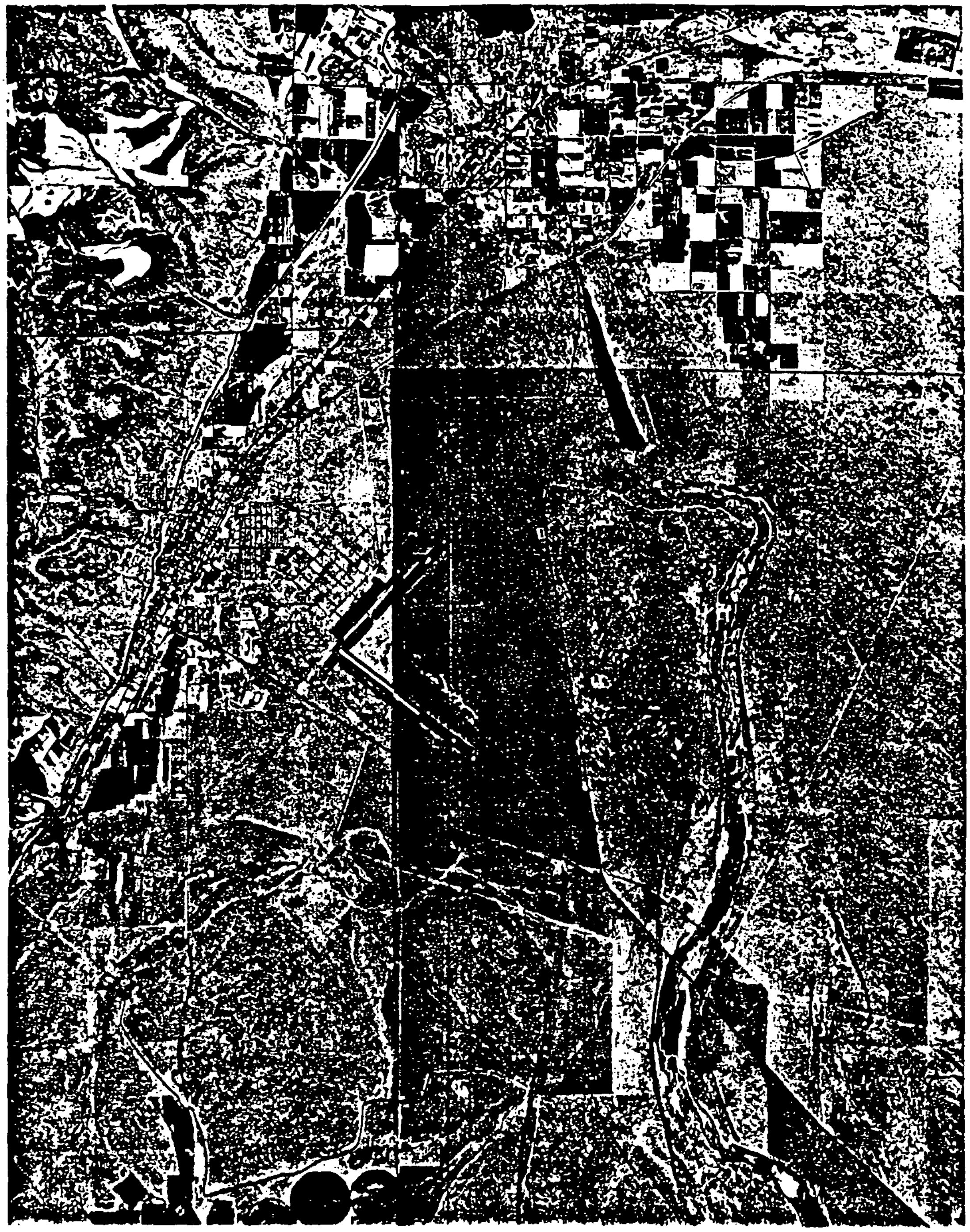

PLATE 2. Upper Ephrata fan. North is up.

PLATE 3. Lower Grand Coulee. North is up.

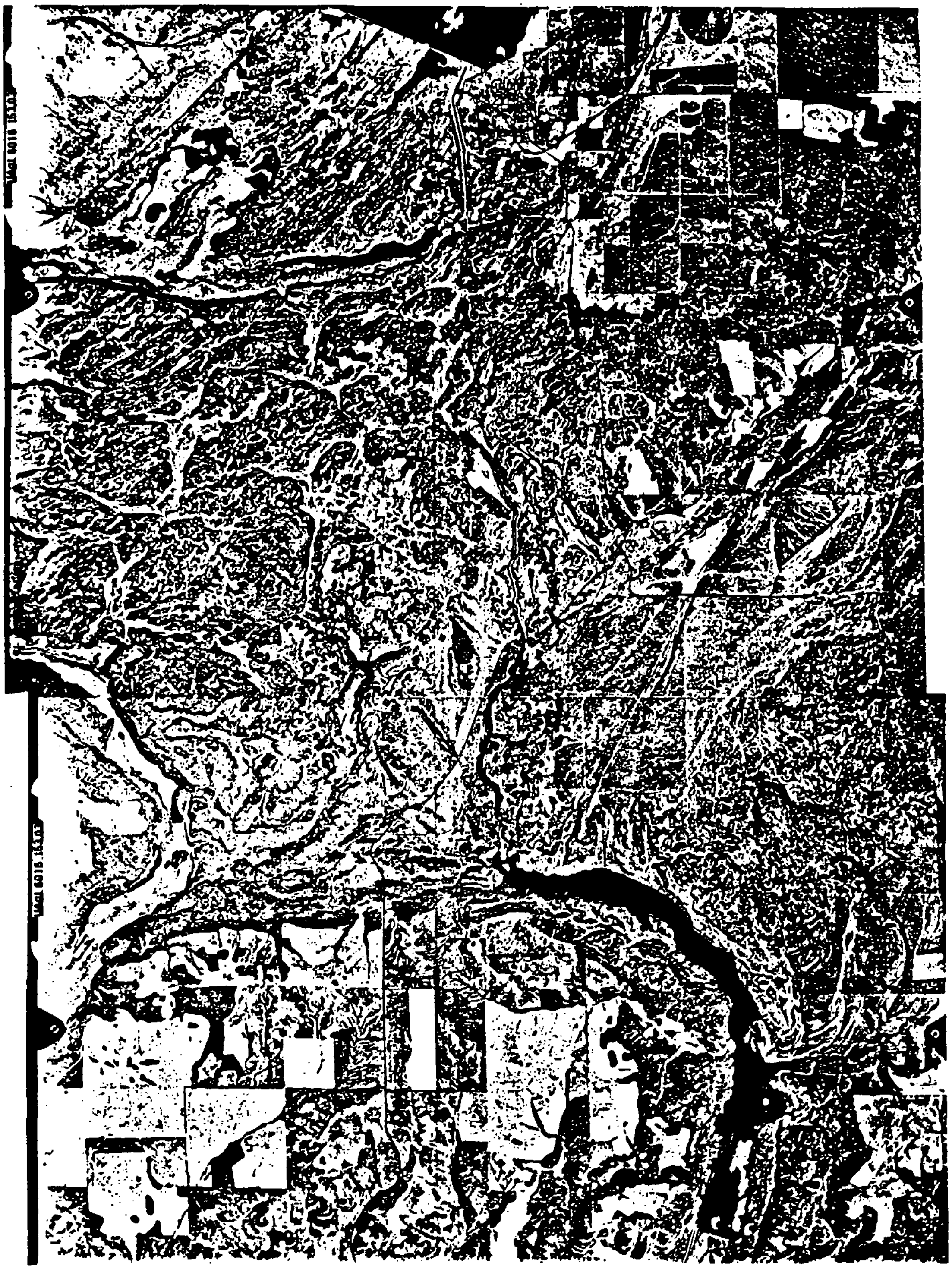

PLATE 4. Hartline Basin with Pinto Ridge. North is up.

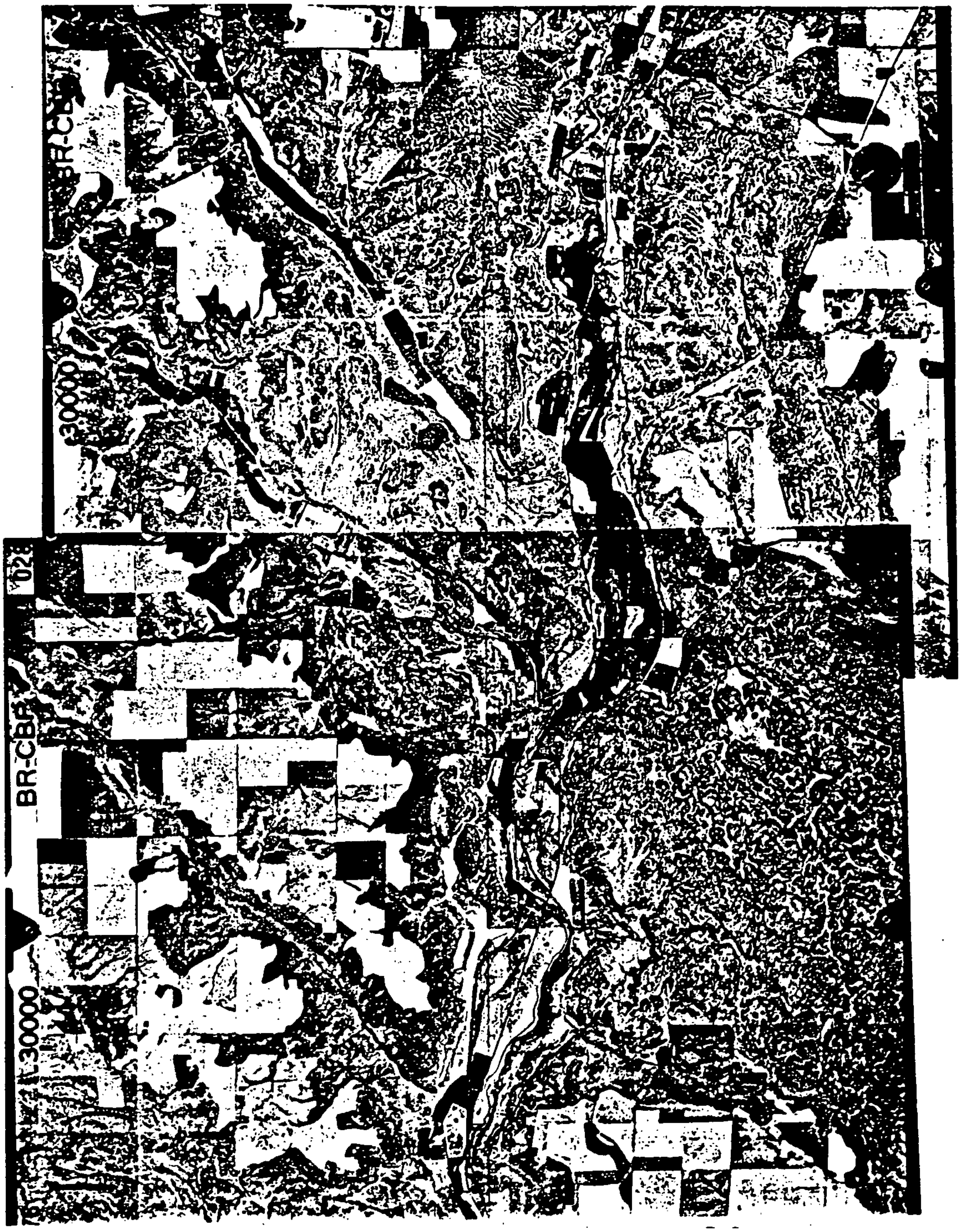

PLATE 5. Upper Crab Creek and Wilson Creek. North is to the left.

PLATE 6. Potholes Coulee and Babcock Bench. North is up.

PLATE 7. West Bar, Crescent Bar and Crater Coulee. North is up.

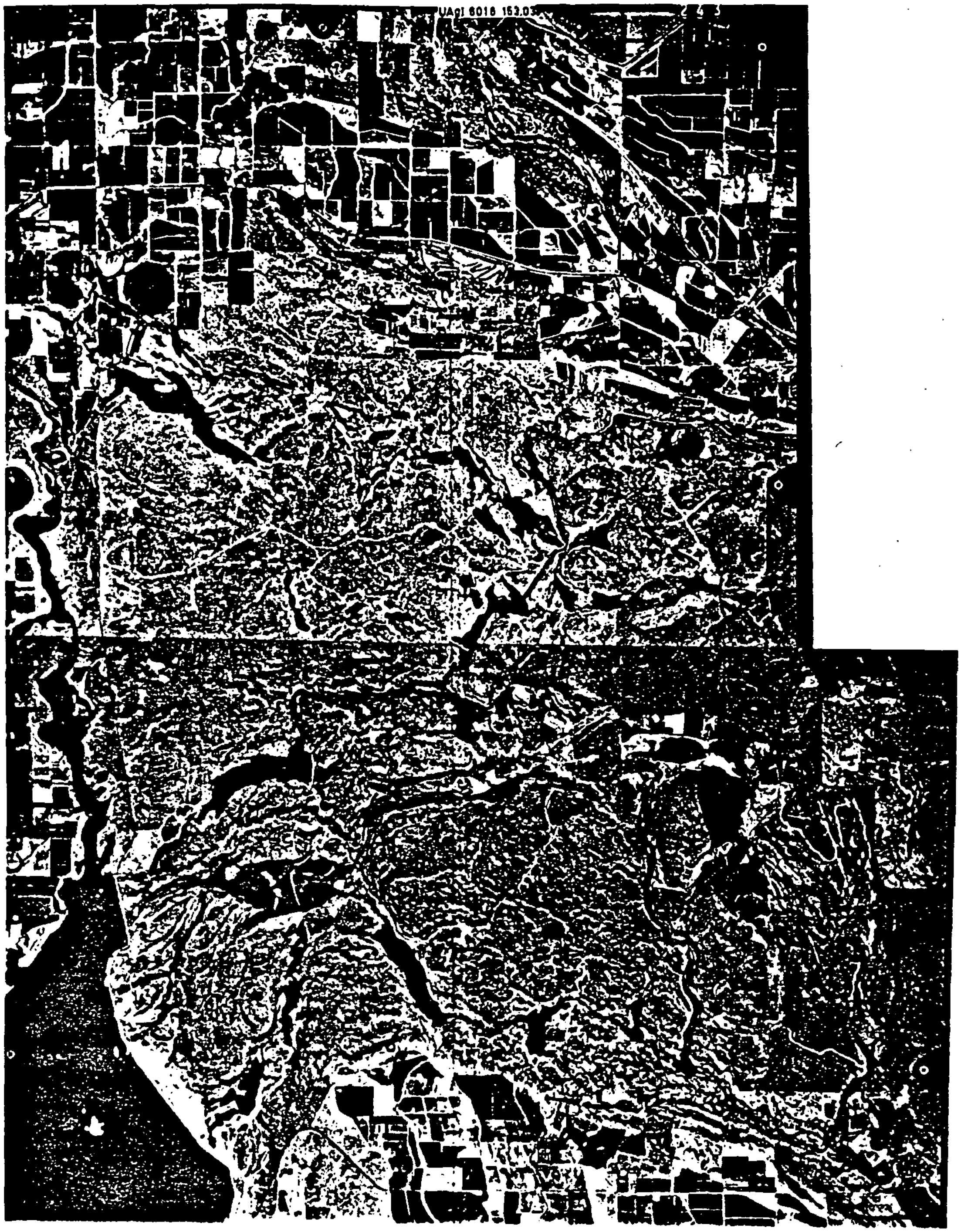

PLATE 8. Drumheller Channels. North is to the left.

channels which cut through the basic form. The largest channels have cataracts and weakly developed butte-and-basin topography. Most of the pendant bars and expansion bars downstream from cataracts are associated with these channels (Fig. 6.2). Still smaller channels at higher elevations have been eroded through the loess cover but have only sligthtly scarified the underlying basalt. Finally, the smallest channels are the eroded divide crossings, some of which are filled with slackwater deposits.

The Cheney-Palouse tract contains groupings of erosional residuals (loess islands) that have been modified by three or four levels of stream erosion as indicated by channel size and degree of scabland topographic development. The characteristic arrangement of erosional elements probably represents variations in flow velocities, rates of erosion, and rates of deposition for various elevations in the flood channel way. In modern braided stream environments, Williams and Rust (1969) have noted a decrease in the flow regime, water discharge, rate and mode of sediment transport, and period of activity with increasing elevation or ranking of the channel. We speculate that the same conditions existed during the creation of the Cheney-Palouse. The lowermost channels probably carried the greatest discharges for the longest durations generally with the greatest velocity and turbulence. These channels would logically be expected to exhibit the highest degree of scabland development and the least amount of associated sediment deposition. Sedimentation was greatest in the secondary channels probably because sediment concentration was still high, although velocities were somewhat reduced. These factors, when combined with the numerous channel expansions and large flow obstructions, created numerous zones conducive to deposition.

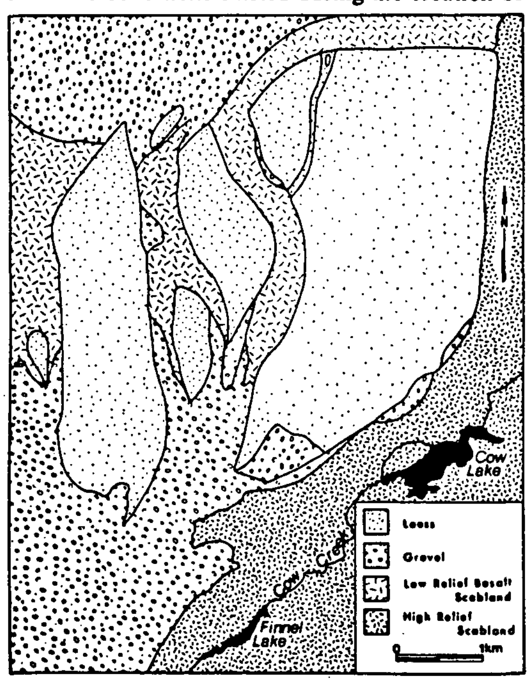

Figure 6.8. Map of the Karakul Hills loess islands.

The analogy between the subaerially exposed loess island complexes and the longitudinal bars of braided streams has already been noted. As in braid bars, the greatest potential for gravel deposition was downstream from the largest channels dissecting individual loess islands. The loess islands created low velocity zones in their lees which localized the deposition of secondary gravel bars. The loess islands also behaved as major elements within a braided stream channel as they forced the main flow against the channel margins. Therefore, the loess islands must be, at least in part, responsible for the width of the Cheney-Palouse tract.

A major question which remains unanswered is exactly what mechanism was responsible for initially allowing the formation of the loess islands. In alluvial braided streams, the formation of a midchannel bar is usually started by deposi-

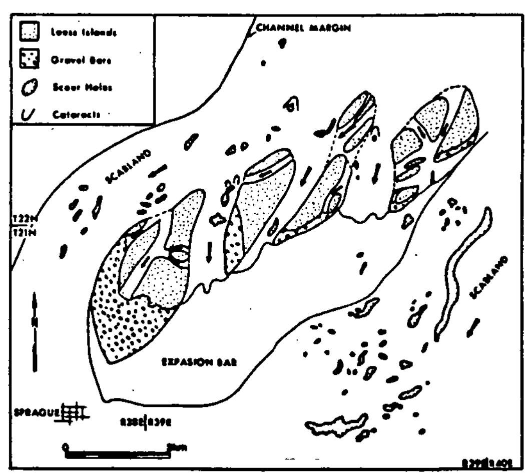

Figure 6.9. Map of the Sprague loess island assemblage.

tion of bedload because of a local incompetence in the flow (Leopold and others, 1964). The braided channels formed in this manner may eventually increase in depth and slope by erosion and cause a temporary increase in the competence of the flow (Leopold and others, 1964).

The Cheney-Palouse loess islands could not have formed in this manner because they are erosional and formed during downcutting. Nevertheless, the end result of the process was the same: increases in relative depth, velocity, and flow competency. Perhaps, as Church (1972, p. 74) suggests, the braiding was partly caused by the increasing boundary resistance that occurred as the channel widened by bank erosion. In order to maintain a great enough velocity for sediment transport, the channel divided, and incision created relatively narrow and deep secondary channels. Therefore, in the Cheney-Palouse, the erodible loess hills which formed the channel margin were probably the ultimate cause of the anastomosing pattern.

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Chapter 7

Field Trip Stop Descriptions

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INTRODUCTION

Fifteen sites within the Channeled Scabland have been selected as stops for the ground portion of this field conference. The stops were selected with the dual aim of visiting locations critical to the arguments for a catastrophic flood origin of the scablands as well as permitting an examination of the variability in both erosional and depositional features. The stop locations are plotted on a generalized geologic map of the Channeled Scabland (Fig. 7.1). Their coordinates are given in Table 1 (at the end of the chapter).

STOP 1: WILLIAMS LAKE GRAVEL PIT

The region southwest of Cheney is characterized by streamlined hills of Palouse loess separated by wide scabland channels eroded into Yakima basalt. As shown by Patton and Baker (Ch. 6, this volume), bars of flood gravel commonly extend both along the flanks and at the downstream end of these erosional remnants. The bars probably formed as a result of reduced turbulent action and sediment deposition in the wake-zone immediately downstream from the loess hill. Figure 7.2 shows the inferred flow pattern around the loess hill southeast of Williams Lake and demonstrates that individual gravel bars might have developed on both downstream flanks of the loess remnant and migrated toward the common center of the downstream wake-zone. The composite photo of the pit face (Fig. 7.3) shows that the majority of foresets dip toward the southwest (left in figure). This is consistent with the hypothesis that the bar originated as a pendant to the remnant one mile to the northeast of this location. The upper gravel unit is poorly stratified but appears to indicate a more southerly or southeasterly dip, suggesting that individual bar slipfaces migrated in alternate directions as the pendant accreted.

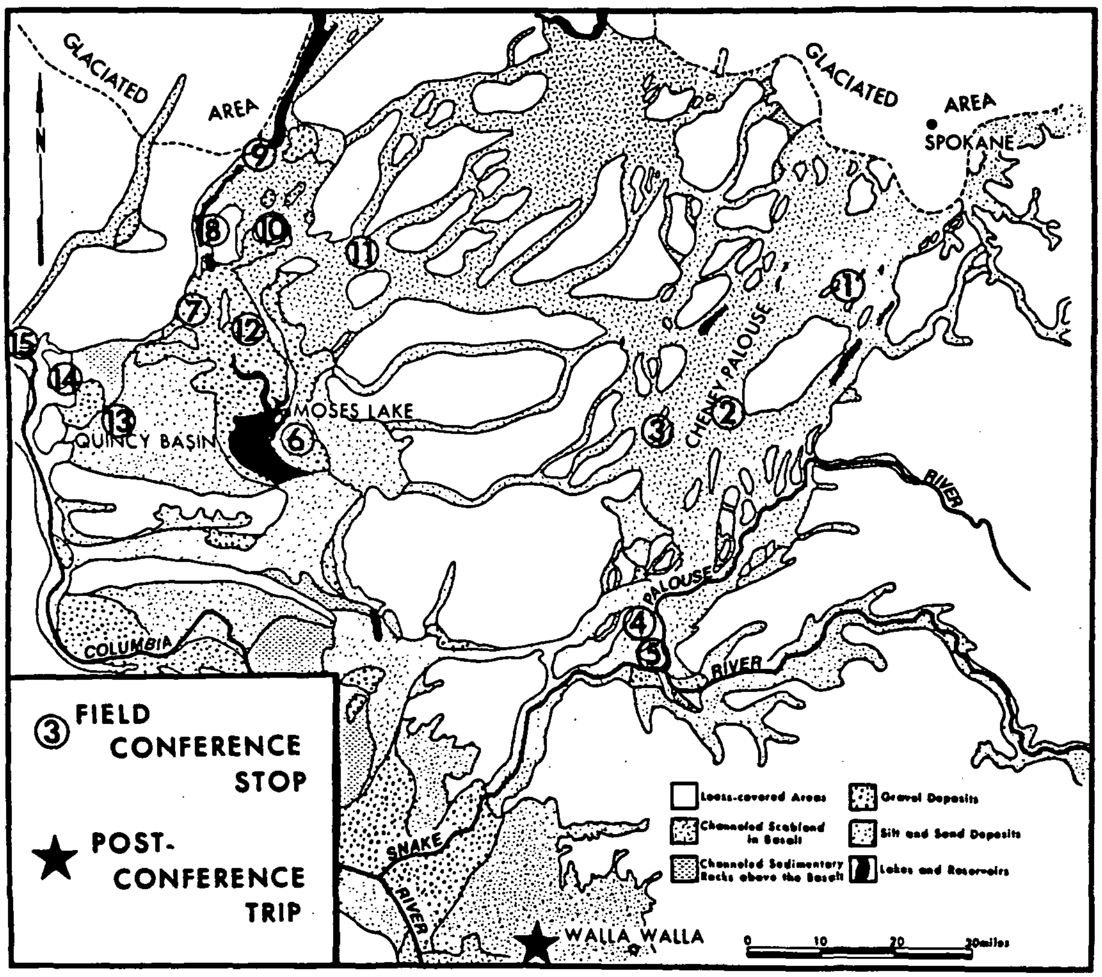
Throughout the pit, one finds individual cobbles and boulders scattered among much finer gravel in the cross-beds (Fig. 7.4). This is common in flood gravel throughout the scablands (see, for example, Fig. 7.32). It is assumed that bed form migration on the bar surface combined with dispersive shear during the process of sediment avalanching down the slipface produced the regular size-sorting in the foresets. Therefore, the large clasts must have arrived by processes other than traction transport. One possible mechanism would be macroturbulent suspension transport. The presence of loess clasts in the deposit (Fig. 7.5) is also suggestive of suspension transport as tractive movement of such clasts would quickly lead to their disintegration.

STOP 2. MACALL-RITZVILLE ROAD

The inferred similarities between braided stream flow and certain characteristics of scour and deposition within the Cheney-Palouse scabland tract (Patton and Baker, Ch. 6, this volume) can best be discussed with reference to the loess remnants and gravel bars along the Ritzville-Macall road. Although these morphologic relationships are best observed from the air, the ground view provides a better perspective of scale.

Over a distance of about 3 km, the road traverses the following features in a westward succession: a major pendant bar (Figs. 7.6 and 7.8), a scour hole within a secondary channel, a second pendant bar and two basalt knobs with attached, downstream aligned, pendant gravel bars (Fig. 7.7). To the west of this last bedrock outcrop, a train of giant current ripples can be observed climbing up the adverse downstream slope of a scour hole. According to Patton and Baker (Ch. 6, this volume), these ripples are developed on gravel less than a meter thick in the ripple troughs. Relatively thin gravel is typical of bars in the area. The ones accessible from this road may average 10 to 15 meters in thickness. These thicknesses are substantially less than the reported 40 meters of fill in the Quincy Basin (Stop 7) and may relate to the shallower water depths and generally more uniform flow conditions than those responsible for the Ephrata Fan expansion bar.

Figure 7.1. Generalized geologic map of the Channeled Scabland (after Bretz, 1959, his Plate 1) with field trip stop locations superimposed.

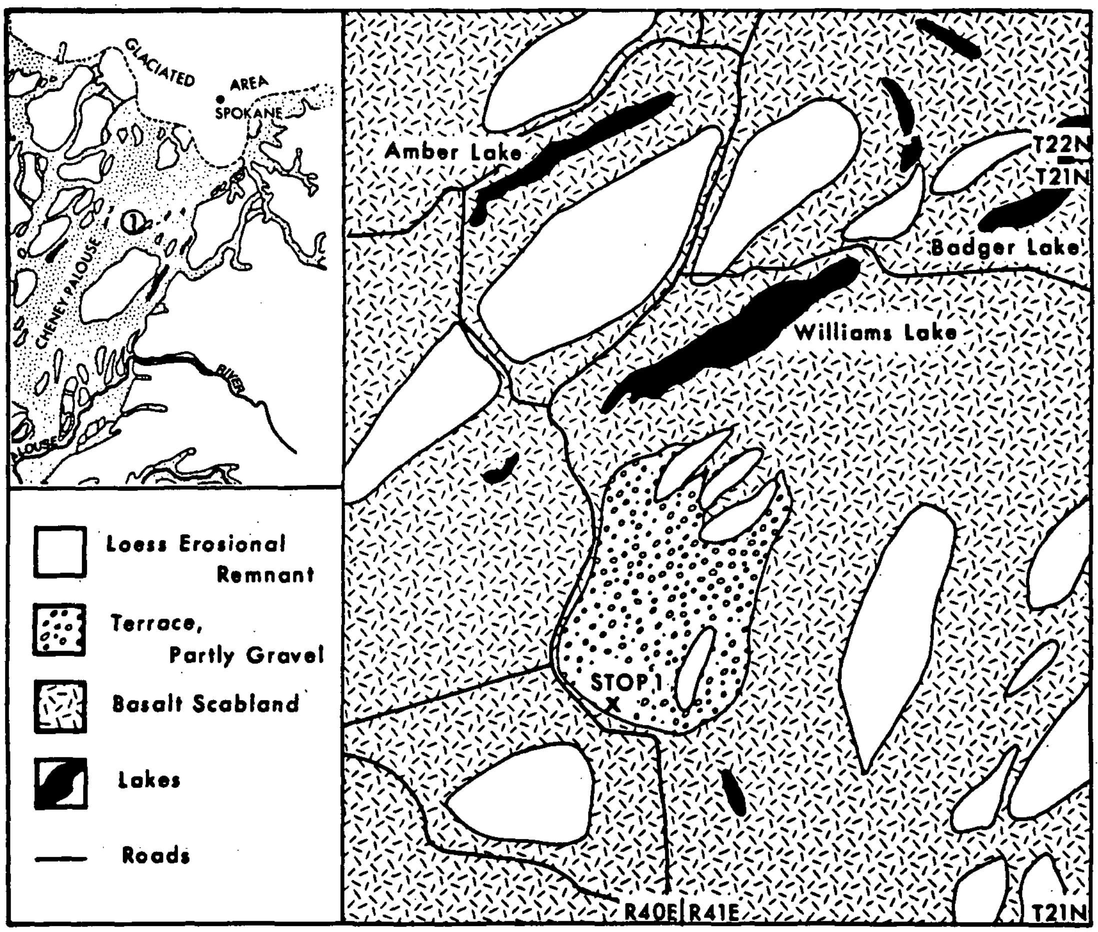


The numerous bars along this stretch of road all are attached to a downstream bedrock step created by a resistant basalt flow (Patton and Baker, Ch. 6, this volume). Their precise location may also have been aided by the transition from upstream constrictions between two loess islands to a major channel expansion (Figs. 7.6).

Further south (downstream), this channel maintained a more uniform width (Fig. 6.2) and its floor consists largely of stripped basalt with large areas of scabland morphology (Fig. 7.8).

In terms of channel bifurcation and sediment deposition, patterns of similarity exist between the Cheney-Palouse tract and modern braided streams. For example, the loci of sediment deposition correspond to sites of reduced current velocity or turbulence. In Cheney-Palouse (Fig. 7.6), as in modern braided streams (Boothroyd and Ashley, 1975; Nummedal and others, 1974), such zones generally occur where channels of higher elevation and less discharge rejoin a larger channel skirting the bar margins. The sand-wedge slipfaces of Rust (1972) and Boothroyd and Ashley (1975) appear to have their catastrophic equivalents in some of the pendant gravel bars (and expansion bars) of the Cheney-Palouse. Nevertheless, it is important to remember that the hills of the Cheney-Palouse, which by-and-large are responsible for the braiding (Patton and Baker, Ch. 6, this volume) are erosional rem-

Figure 7.2. Simplified morphologic map of a region southeast of Williams Lake (Part of Fig. 7.1 inset for location). Mapped from U.S.G.S. 15' topographic map, Cheney quad.



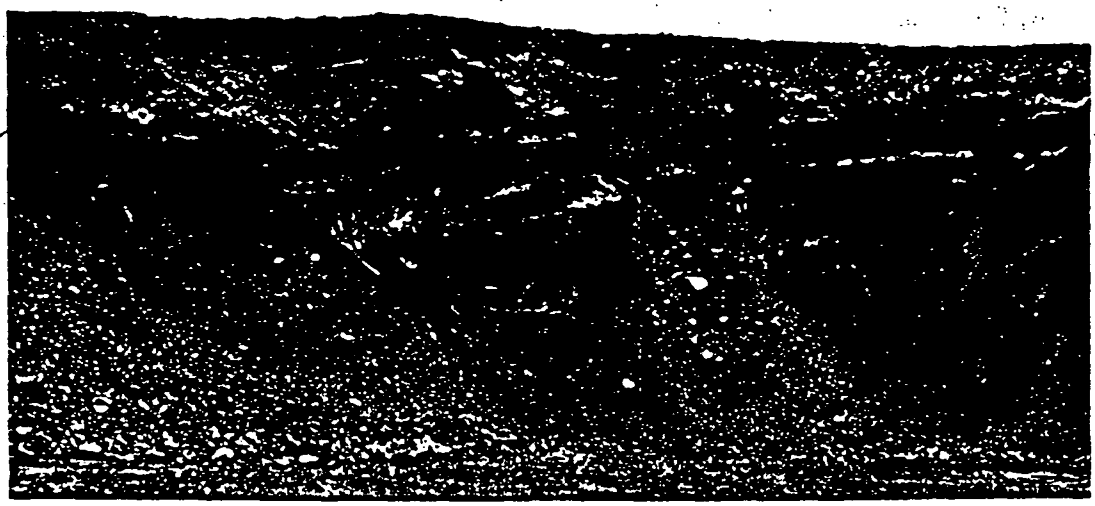


Figure 7.3. Composite photograph of the face of the gravel pit southeast of Williams Lake. The majority of the foresets dip toward the southwest (left).

Figure 7.4. Close-up of bar cross-stratification. Large cobbles (about 10 cm long) are dispersed throughout the finer cross-bedded gravel.

nants of a pre-flood geological formation and as such genetically totally unrelated to braided stream bars.

Figure 7.5. Loess clasts are common throughout the gravel bar. This one measures about 15 x 15 cm.

STOP 3. MARENGO RAILROAD CUT

As shown in Figures 7.6 and 7.9, the tracks of the Chicago, St. Paul, Milwaukee and Pacific Railroad cut through the pendant bar at the south end of a large loess remnant about ½ kilometer west of the Marengo siding. The origin and in-

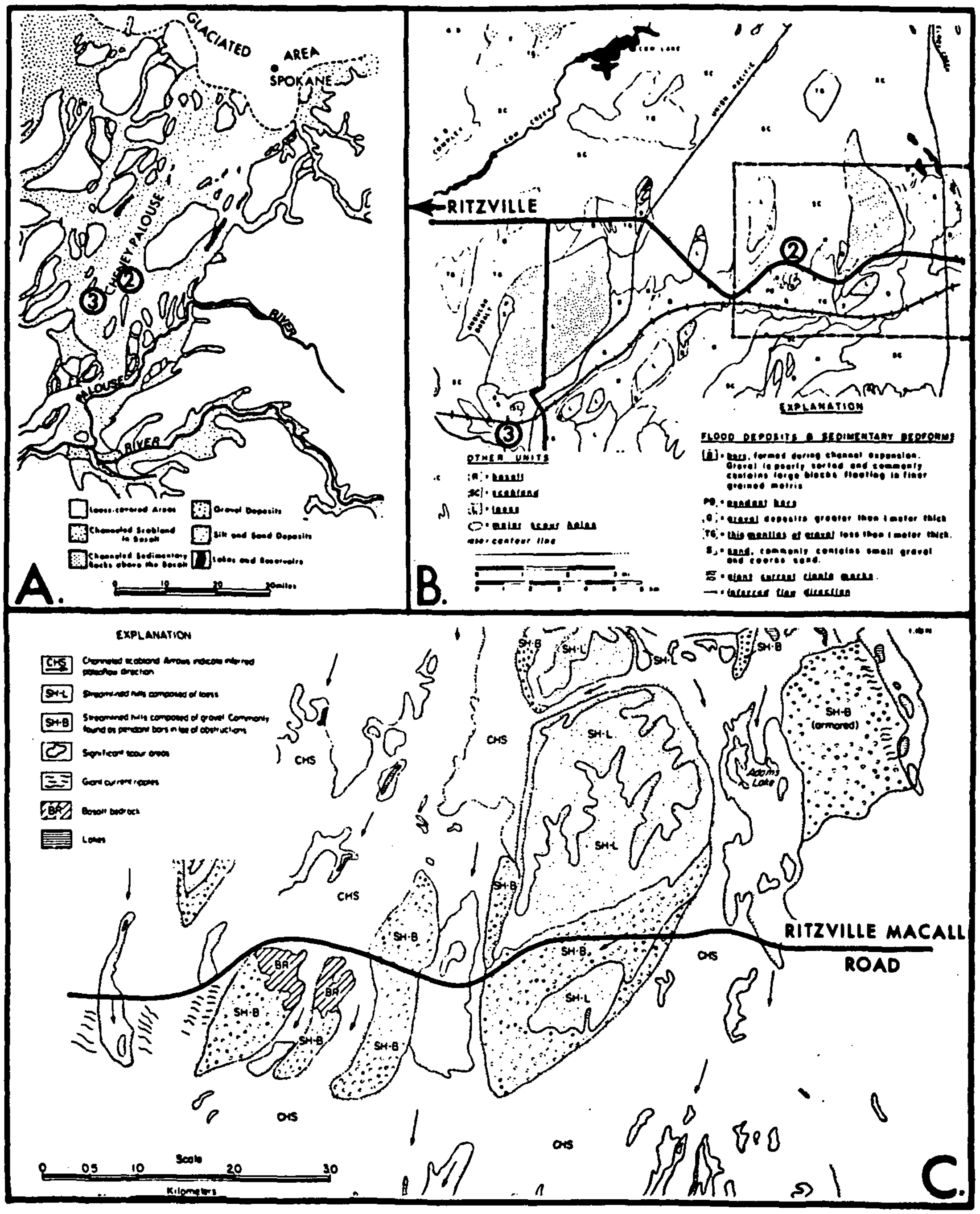


Figure 7.6. Morphological maps of the Ritzville-Macall road traverse across the Cheney-Palouse scabland tract. A is a part of Fig. 7.1 for location. B identifies the setting of Stops 2 and 3 relative to bars and loess hills of the Cheney-Palouse tract (mod. from Fig. 6.2). C shows morphologic detail in the vicinity of Stop 2 (modified from Fig. 6.7).

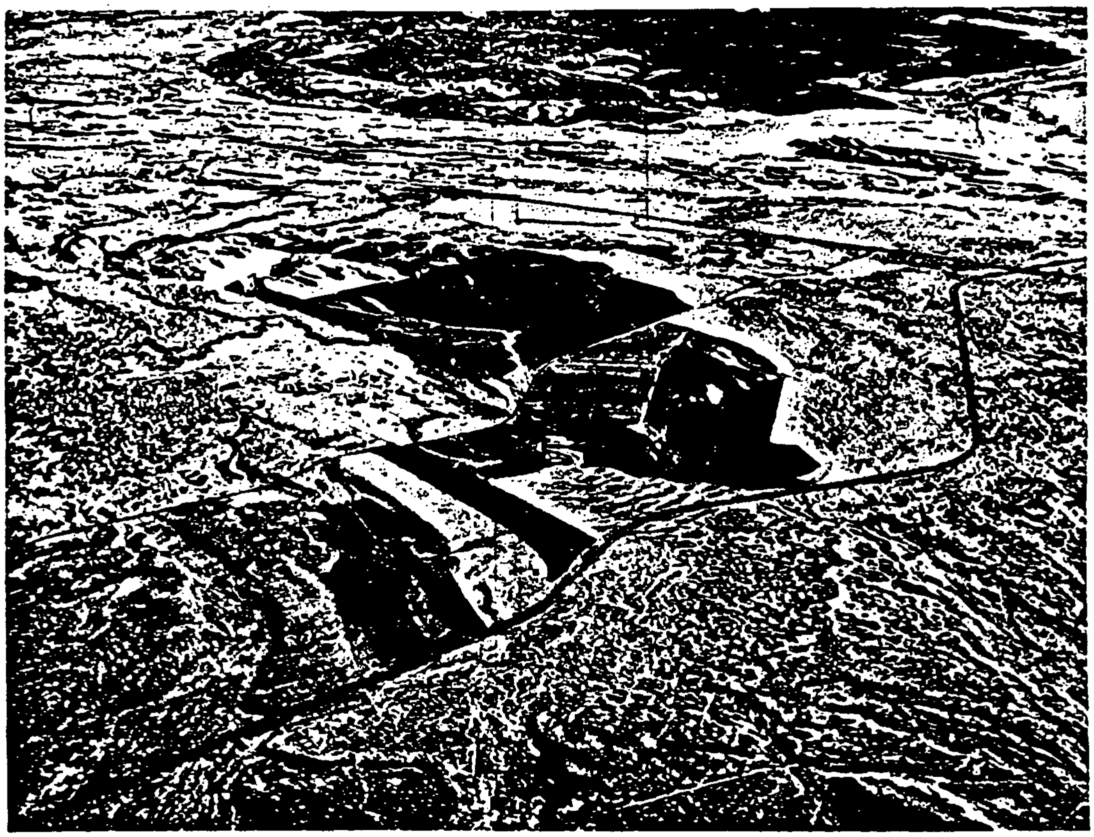


Figure 7.7. Oblique aerial view toward the northeast of the Macall area. Compare maps in Fig. 7.6 for identification of the individual features. Do not confuse the Macall-Ritzville road (across the center of the photo) with the railroad tracks (lower right) (photograph 6239 by John S. Shelton).

ternal stratification of this bar are probably similar to what was observed at the Williams Lake gravel pit. Sedimentary structures, however, are not that well displayed in this section.

The railroad-cut provides excellent insight into the Pleistocene stratigraphy of the Cheney-Palouse tract. The open-work, coarse flood gravel interbedded between loess units containing well-developed soil horizons constitute evidence for two floods, one perhaps pre-Bull Lake, the younger probably early Pinedale (see Table 2.1). The stratigraphic section measured from the top of the cut in Figure 7.10 by Patton and Baker is presented in Figure 7.11 (See also Baker, Ch. 2, Fig. 2.5, this volume). The lower gravel has all the fabric characteristics of flood gravel elsewhere in the scablands. It is distinguished by a thick weathering rind and is overlain by a dark yellowish brown loess capped by a petrocalcic horizon (K-horizon) strongly suggestive of its pre-Bull Lake age (Fig. 7.12). Overlying the petrocalcic horizon is more loess with other soil profiles. These younger soils also have mature profiles, textural B horizons and calcareous Cca horizons (Fig. 7.13). These caliches, however, are less thoroughly cemented and do not qualify as K-horizons (Baker, 1977, p. 408). They comprise the Palouse Formation which is interpreted as Bull Lake in age. Overlying this sequence is a second layer of flood gravel, significantly less

Figure 7.8. Oblique aerial view toward the south of the Macall area. See maps in Fig. 7.6 for location (photograph 6235 by John S. Shelton).

weathered than the lower unit. The section is capped by the modern soil profile developed in Late Pinedale (and Holocene) loess. Elsewhere in the Cheney-Palouse scabland tract, Patton and Baker (Ch. 6, this volume) have found additional evidence of multiple flood events.

Figure 7.9. Oblique aerial view toward the southwest of the Marengo loess hill. Stop 3 is in the railroad-cut through the pendant bar visible in the upper right (arrow).

STOP 4. PALOUSE-SNAKE DIVIDE CROSSING

The region around the Palouse-Snake River junction contains, next to Grand Coulee, perhaps the most impressive morphologic evidence of

Figure 7.10. The Chicago, St. Paul, Milwaukee and Pacific Railroad cut through the pendant bar at the downstream end of the Marengo loess hill.

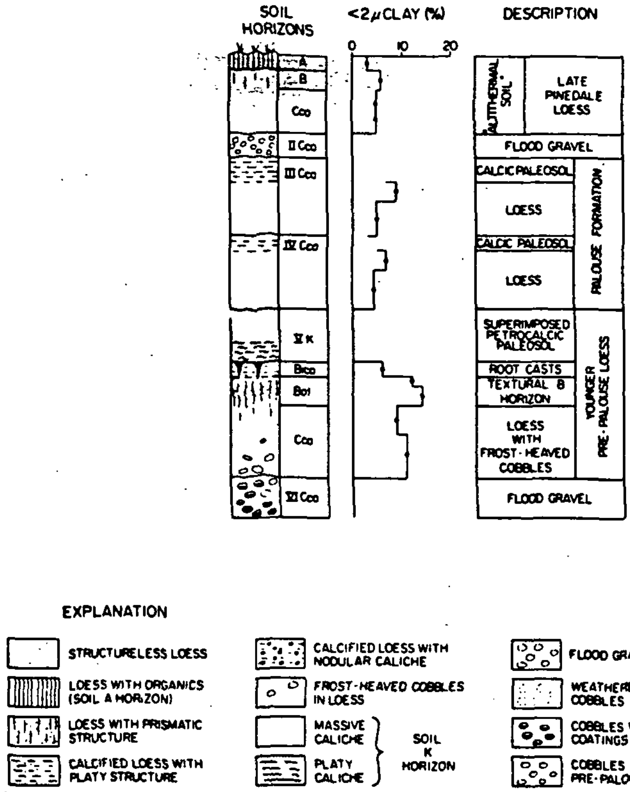


Figure 7.11. Measured stratigraphic section at the Marengo railroad-cut (modified from Fig. 2.5).

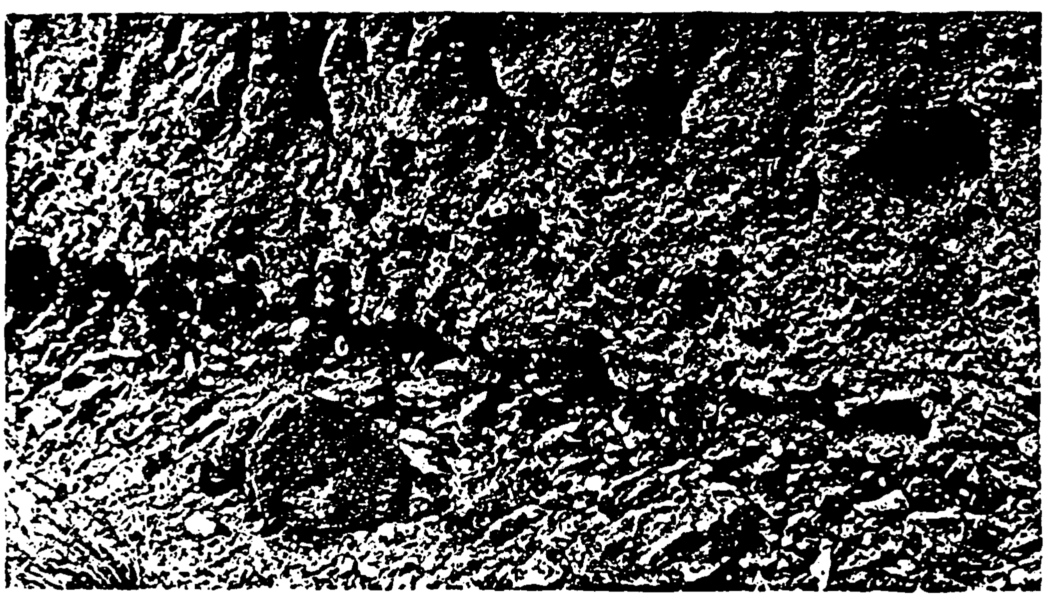
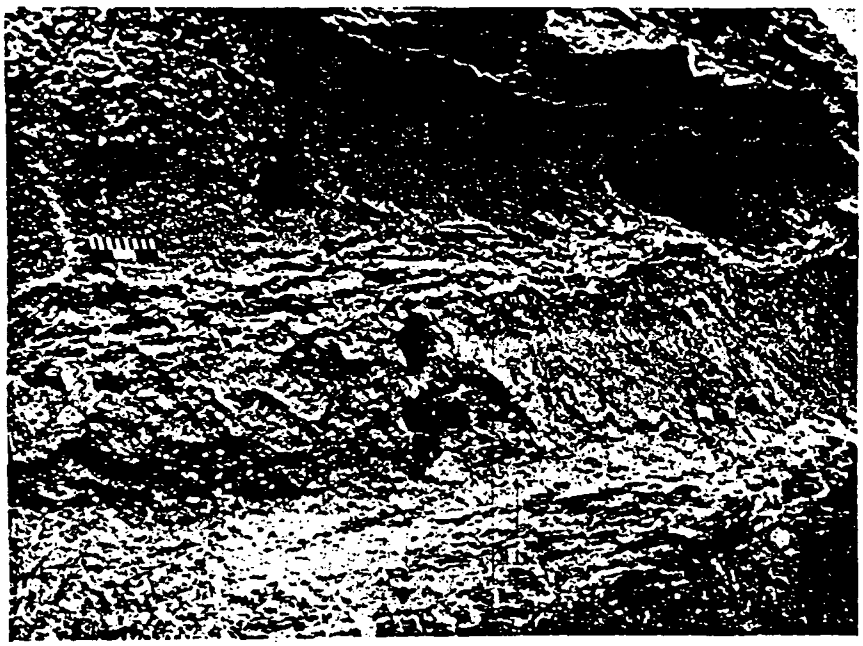


Figure 7.12. Close-up view of the lower flood gravel at the Marengo section. The basalt cobbles have thick weathering rinds and are overlain by a pre-Bull Lake loess.

catastrophic flooding. In the pre-flood drainage pattern, the Palouse River flowed toward the southwest from Washtucna, down the lower Washtucna Coulee (Fig. 7.14). With the late Pleistocene Missoula Flood rushing down the Cheney-Palouse tract, most of the water took a shortcut directly across the Palouse-Snake divide south of Hooper and Washtucna. Bretz (1959, p. 40) declares: "There is no more significant area than this divide for testing rival theories." The bedrock divide is 16 kilometers across and with its loess cap at least 100 meters higher than the floor of the preflood Palouse Valley. The flood overflowed this valley and overtopped the divide. As a result of the overtopping, it stripped nearly all loess off the divide, leaving only a few narrow, streamlined hills near the summit (Figs. 7.15, 7.16 and 7.17). These 50-meter high hills are probably subfluvial in origin (Baker, 1973a). Flanking these hills are the 125-meter deep Palouse Canyon to the east (Fig. 7.17) and the dry cataract next to the H U Ranch to the west. Two large rock basins are found near the very summit of the divide crossing (Fig. 7.14). As a result of the flood, the lower Washtucna Coulee was left without any through-drainage, and today contains only small alkaline lakes. The Palouse River was diverted to the new gorge carved by the flood into the highly jointed bedrock south of Washtucna. The river itself occupies what appears to be an inner channel (Shepherd and Schumm, 1974) developed subfluvially during the erosion of the canyon. Headward recession of the cataract initiated on the north wall of the Snake River Canyon produced Palouse Falls, where the river today drops through 60 meters of free fall from the upper to the lower canyon (Fig. 7.18).

This is part of the erosional landscape which Flint (1938b) attributed to the action of "leisurely meandering glacial streams, no larger than the Snake River of today," an idea which Bretz and others (1956, pp. 1015-1020) dismantle by a meticulously detailed analysis of the flow patterns across the Palouse-Snake divide. Bretz (1959, p. 41) concludes: "The scabland features just enumerated are utterly impossible erosional forms for such streams [i.e. leisurely meandering] to have made."

Figure 7.13. Section of the Palouse Formation at the Marengo railroad cut.



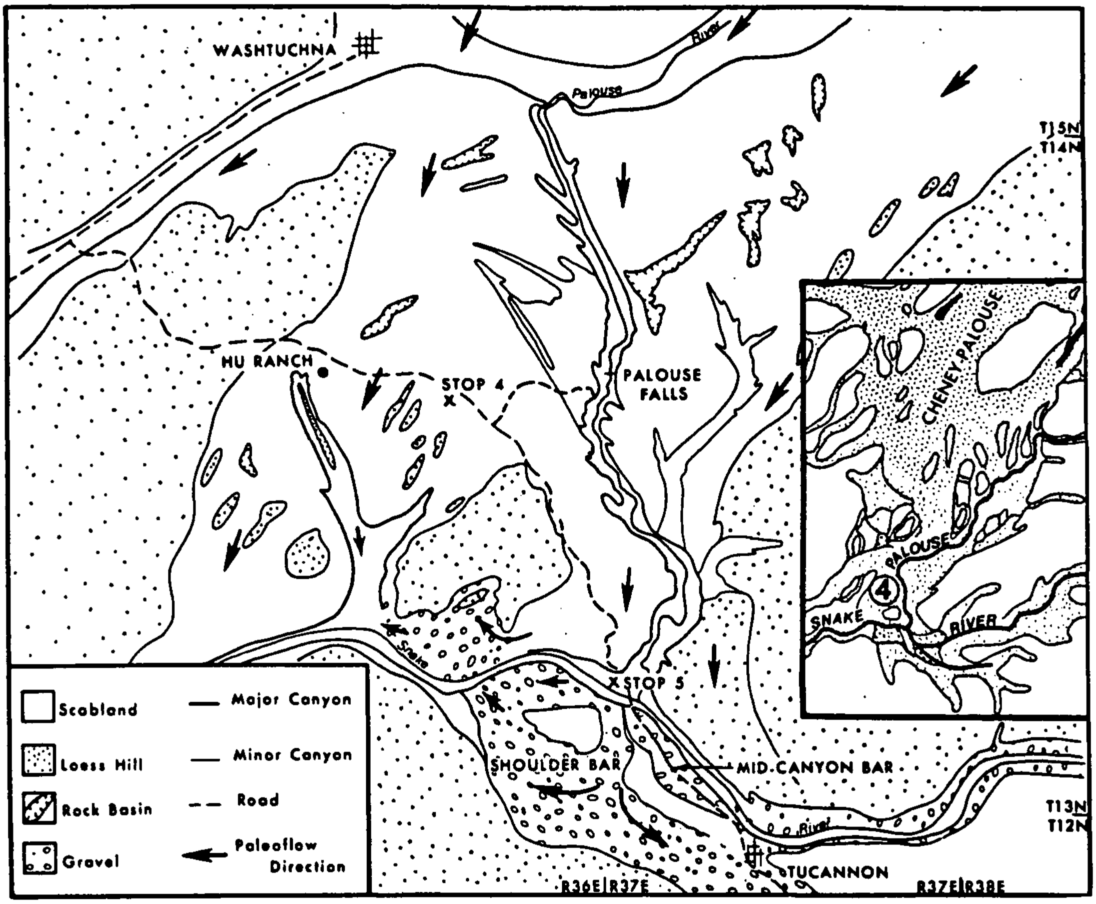


Figure 7.14. Generalized morphological map of the region surrounding the Snake-Palouse River junction. Mapped from U.S.G.S. 15' topographic maps, Starbuck and Haas quads. Arrows indicate the inferred paleoflow directions.

STOP 5. PALOUSE-SNAKE JUNCTION

From this vantage point underneath the Union Pacific Railroad trestle, one can reconstruct the pattern of flow as the flood currents swept across the Snake River Canyon from the north. South of the Snake River at this point are the only two high-elevation flood gravel deposits found anywhere along the canyon of the lower Snake River. These are Shoulder Bar and Mid-Canyon Bar (Fig. 7.14). Shoulder Bar (Bretz, 1928b, p. 657; Bretz and others, 1956, p. 1020) represents a square mile of flood gravel with a surface elevation 150 meters above the valley floor. Flow across the bar was northward as

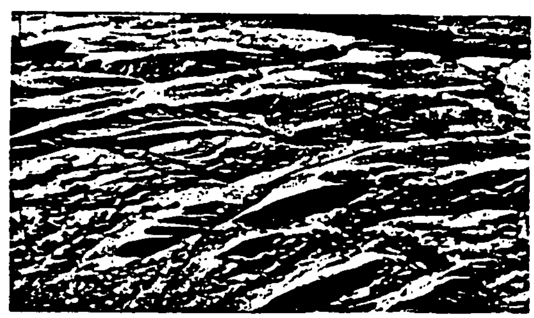


Figure 7.15. Oblique aerial view toward the southeast of streamlined Palouse loess hills just east of the H U Ranch. Flood flow direction was from lower left to upper right.

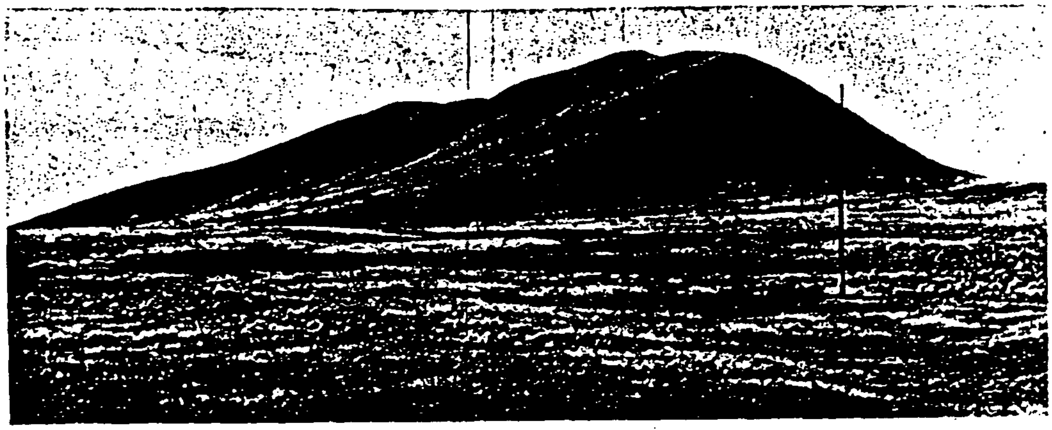


Figure 7.16. Ground view of the prominent loess hill in the lower right of Fig. 7.15.

Figure 7.17. Oblique aerial view toward the northwest across the western part of the Palouse-Snake divide. In the foreground is the lower part of the Palouse River Canyon with the falls off the edge of the picture to the right (photograph 2042 cr by John S. Shelton).

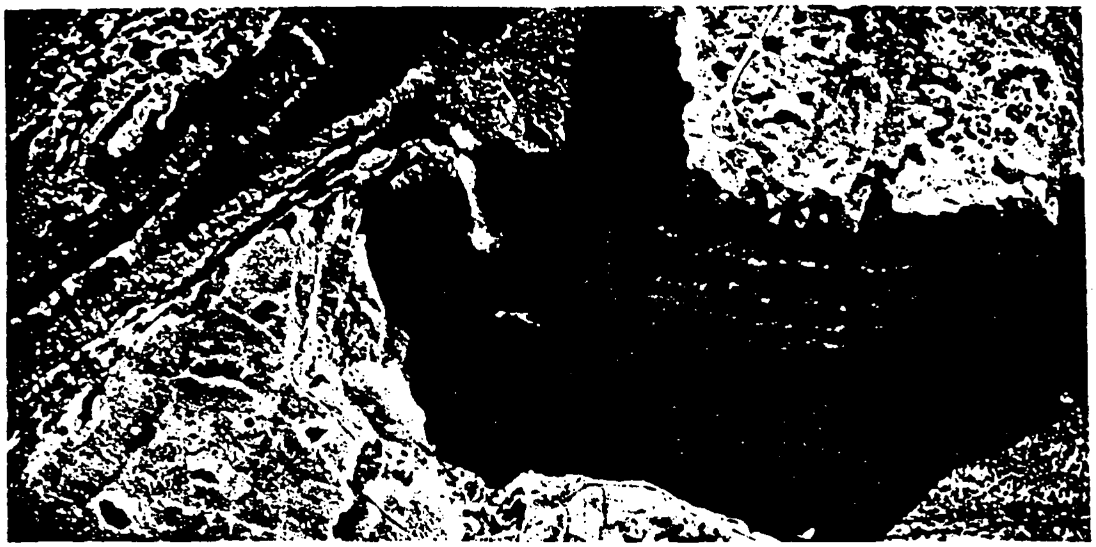


Figure 7.18. Aerial view of Palouse Falls. Note the numerous joints in the bedrock to the right of the falls. The Palouse River Canyon upstream of the falls was eroded along one such joint set.

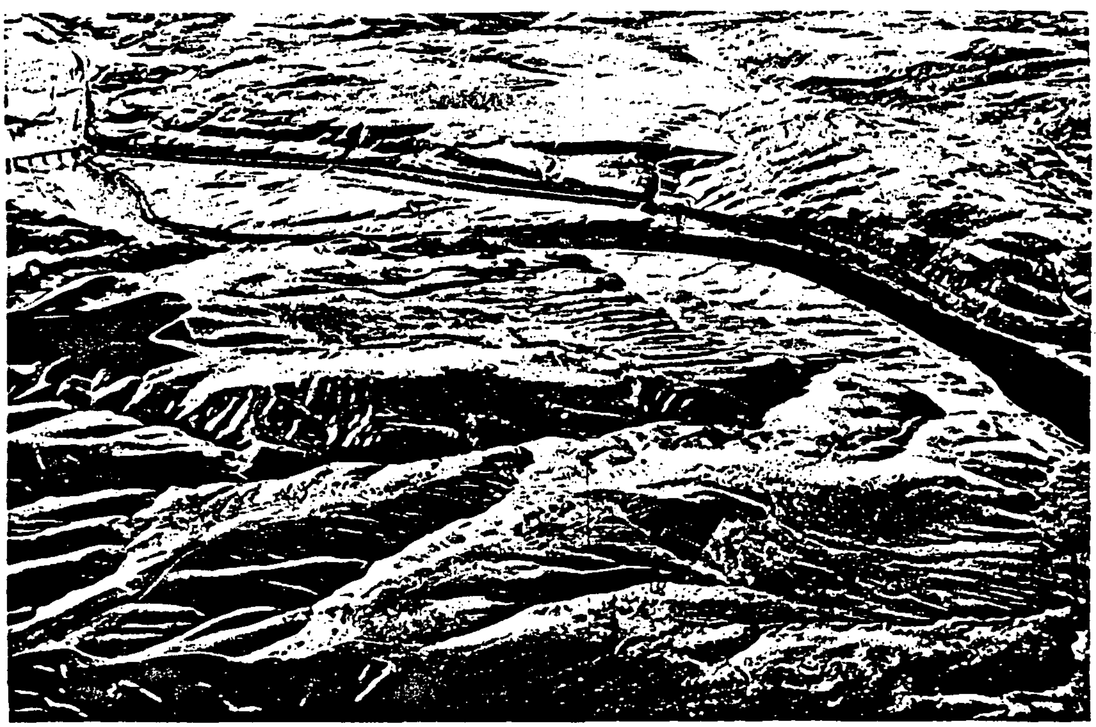


Figure 7.19. Oblique aerial view toward the southeast across the Snake River immediately downstream of the Palouse confluence. Refer to map in Figure 7.14 for locations. Shoulder Bar has steeply dipping foresets and gravel ripples both demonstrating flow northward into the Snake River Canyon. Mid-Canyon Bar is visible in the upper left corner of the photograph (photograph 6245 by John S. Shelton).

demonstrated both by large foresets dipping *northward into* the Snake Canyon and giant current ripples covering the surface (Fig. 7.19). The gravel consists of a mixture of scabland basalt and well-rounded Snake River cobbles and pebbles. The only explanation for this feature is that strong flood currents from the mouth of the Palouse Canyon carried gravel across the Snake River valley *up* the south flank to an elevation of 150 meters, then were deflected around the basalt hill north of Shoulder Bar and returned to the Snake River further west. From there, flood waters followed the present river further downstream (Fig. 7.14). As a consequence of flow divergence and reduced current velocity, a large amount of the sediment load was dropped to build Shoulder Bar.

Some of the flow across the Snake River Canyon was deflected upstream to build the Mid-Canyon Bar adjacent to the modern stream course and Trickle Bar behind the saddle in the bedrock across from the Palouse River mouth (Fig. 7.14). The morphology of this saddle itself is suggestive of an origin by scour during the flood. The trough (fosse) on the landward side of Mid-Canyon Bar (Fig. 7.19) appears to reflect the location of a thread of high turbulence along the side of the steep bedrock flank. Ripples and internal cross-stratification of the point bar between the Union Pacific tracks and the Snake River demonstrate the flood origin of this feature as well (Fig. 7.20). Thus a significant amount of flow was discharged directly down the Snake River Canyon without being detoured across Shoulder Bar. The amplitude of these giant ripples is demonstrated in Figure 7.21. Large basalt columns, probably derived from near-by outcrops attest to the competency of the flow (Fig. 7.22).

The back-rush up the Snake River, as evinced by the topography and stratification of the Mid-Canyon Bar, was capable of eroding Snake River

Figure 7.20. Oblique aerial view toward the east of an unnamed flood bar between the present course of the Snake River and the Union Pacific Railroad. Giant current ripples demonstrate flow down the Snake River Canyon (toward the bottom of the photo).

Figure 7.21. Ground view of the surface of the bar shown in Fig. 7.20. Note man for scale (arrow).

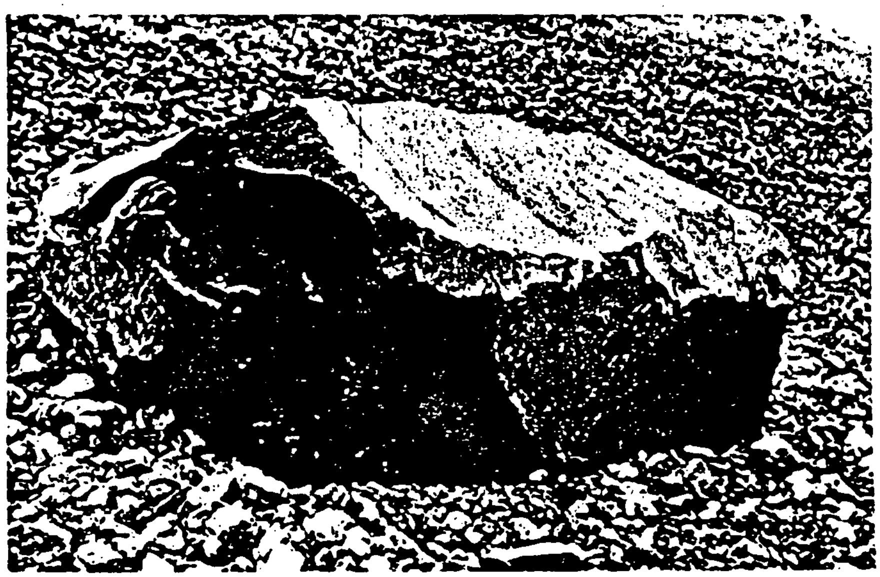


Figure 7.22. Basalt column excavated in the gravel pit on the bar shown in Fig. 7.20. The column is nearly 2 m in diameter.

gravel and redepositing it with foresets dipping upstream all the way to Lewiston, Idaho, 110 km distant (Bretz, 1969, p. 527).

As an interesting archaeological sidelight on this part of the country, the following is repeated from Webster and others (1976, p. 18):

> "Upstream along the Palouse River, approximately 2 miles from this point, was the Marmes Rockshelter. This archaeological site received worldwide attention in the spring of 1968 when human remains were discovered *in situ* 14 feet beneath the surface of the modern flood plain. These remains were established reliably as being at least 10,000 years old—the oldest well-documented human remains in the New World. Numerous artifacts, cultural features and animal bones were associated directly with the human remains. Because the site was to be flooded by impoundment behind Lower Monumental Dam in less than a year, emergency salvage excavations were begun in May 1968 and continued through February 1969 when the reservoir and site were flooded. Marmes Rockshelter contains an unparalleled stratigraphic and cultural record spanning more than 10,000 years."

No formal stops are planned between the Palouse-Snake River junction and Moses Lake. If time permits, however, a quick stop will be made at Roxboro in Lind Coulee to view some of the best-developed giant current ripples in the Channeled Scabland (Figs. 7.23 and 7.24). The ripples display a distinct increase in spacing, lateral crest continuity and height toward the south side of the channel where, supposedly, the current velocity was the highest. Measurements yield a typical height of 3.5 meters, a spacing of 78 meters and a water depth (from high-water marks on the coulee flanks) of 80 meters at this ripple train (Baker, 1973a, p. 60, Table 2). From these data and a measured regional slope of 0.002, he calculated a current velocity of 17 m/sec, a dis-

Figure 7.23. Part of U.S. Bureau of Reclamation topographic map of the Columbia Basin Project. The 2-foot contour interval provides an excellent display of the morphology of the Roxboro ripple field. Section lines provide scale. North is toward top.

charge of about 3 x 10^6 m^3/sec and a Froude number of about 0.7. Due to the relatively uniform flood flow through this part of Lind Coulee, these are probably some of the most precise paleohydraulic calculations that can be made. The results are judged to be representative values for many of the east-west trending coulees connecting the Cheney-Palouse scabland tract with the western regions of the plateau.

STOP 6. MOSES LAKE DUNE FIELD

The fine sand deposited by the Missoula Flood at the distal margin of the Ephrata Fan throughout the central Quincy Basin has been subject to extensive aeolian reworking. Small sand dune fields are common throughout the scablands, but nowhere is the dune field as extensive as to the south and west of Moses Lake.

The western edge of the field, at Winchester Wasteway about 30 km west of Moses Lake, is characterized by a series of semi-parallel east-west trending sand ridges (Fig. 7.25). The ridges today are all stabilized by sagebrush and grass. Actively migrating slipfaces and recognizable parabolic dune forms increase in frequency toward the east (Plate 1 and Fig. 7.26). That portion of the dune field which extends into the agricultural land on the east side of Potholes Reservoir is largely active. The parabolic dune form is readily apparent both here and on the west side of the reservoir (Plate 1).

The orientation of the parabolic dunes demonstrates wind-driven sand transport from west to east, a direction consistent with the prevailing as well as dominant westerly winds recorded at the Grant Co. airport at Moses Lake. It appears that as the parabolic dune form migrates, it leaves behind a series of linear parallel to subparallel ridges, some of which may become buried and re-excavated as a new active dune passes over an older partly stabilized sand ridge (Fig. 7.27).

The development of a more or less contiguous field of parabolic dunes appears consistent with the rather abundant sand supply and unidirectional winds of the area. A somewhat more restricted sand supply seems to characterize adjacent scabland areas. For example, the Wahluke Slope has beautiful examples of single, large-scale parabolic dunes migrating eastward, inside of which are fields of individual barchanoid dunes.

Figure 7.24. Oblique aerial view to the south across Lind Coulee at the Roxboro ripples. Flow was from left to right. North is toward bottom.

STOP 7. EPHRATA GRAVEL PIT

To reach this gravel pit, one drives east on Washington State route 282 from Ephrata toward Moses Lake. Immediately after ascending a steep, 15-meter high incline south of Ephrata, the gravel pit is entered on the south side of the road. The pit is operated by Columbia Concrete Products. The steep incline represents the western margin of the large Ephrata Fan system, an extensive coarse gravel unit deposited as flood waters from the lower Grand Coulee expanded into the wide Quincy Basin after leaving the last coulee constriction at Soap Lake (Plate 2, Fig. 7.28). Drainage into the Quincy Basin also came from Dry Coulee and Long Lake Coulee which, together with the lower Grand Coulee, all drained the large Hartline Basin to the north. Flow also came as a broad front across the divide south of upper Crab Creek. These drainage lines complicate somewhat the paleoflow pattern along the eastern flank of the fan. The dominant shape of the deposit, however, is clearly that of a fan with its apex (but not its highest elevation) at Soap Lake, demonstrating that the source of the bulk of the sediment was the Grand Coulee.

The deposit must have been built as a subfluvial bar, not as a common alluvial fan on which the active channels constantly change location but never submerge the entire fan at one time. The highest portions of the fan, near Ephrata, stand 90 meters higher than the scoured rock bottom of Soap Lake (Bretz, 1969, p. 528) about 8 km to the north. This steep adverse slope (0.01) can only be explained by a rapidly diverging flow pattern and shoaling to the south of the Soap Lake efflux section. Baker (1973a, p. 15) reconstructs a ponded water surface at 425 meters elevation throughout the Quincy Basin based on high-water marks. If this level is coincident with maximum scour at Soap Lake, the flow depth would have changed from about 100 meters at Soap Lake to a mere 10 meters east of Ephrata. Further hydraulic calculations by Baker (1973a, his Plate 1) yield a peak discharge for the Soap

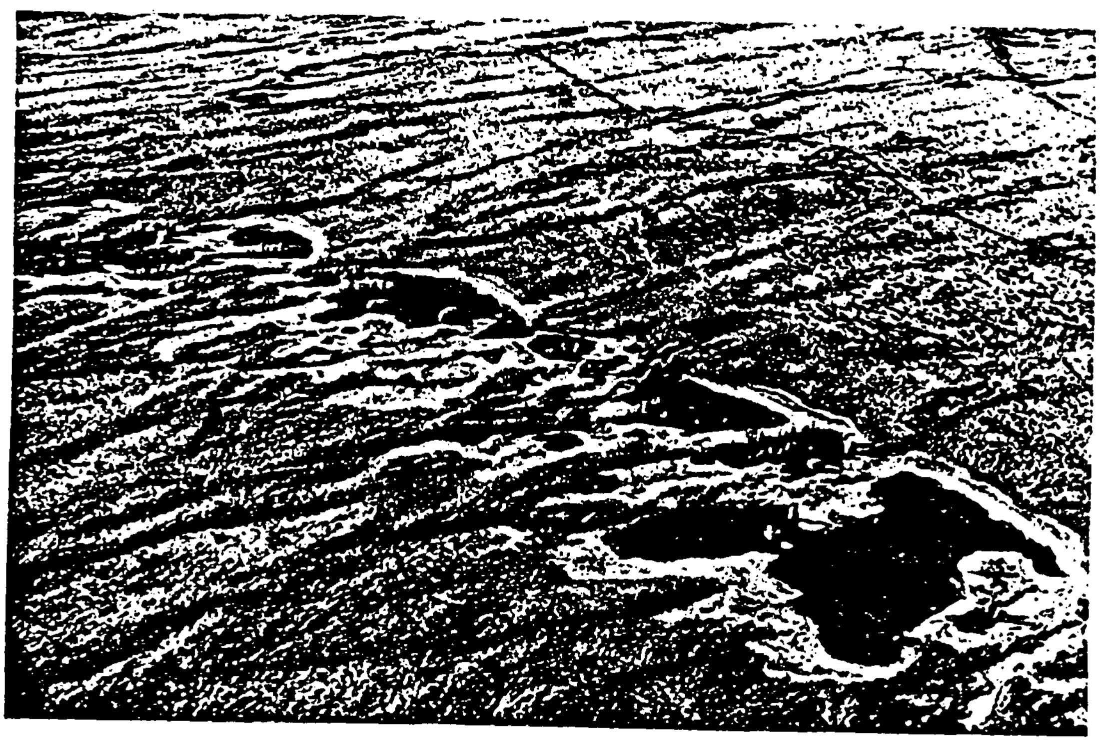


Figure 7.25. The western part of the Moses Lake sand dune field is characterized by a series of stabilized linear sand ridges, probably the remnants of migrating parabolic dunes.

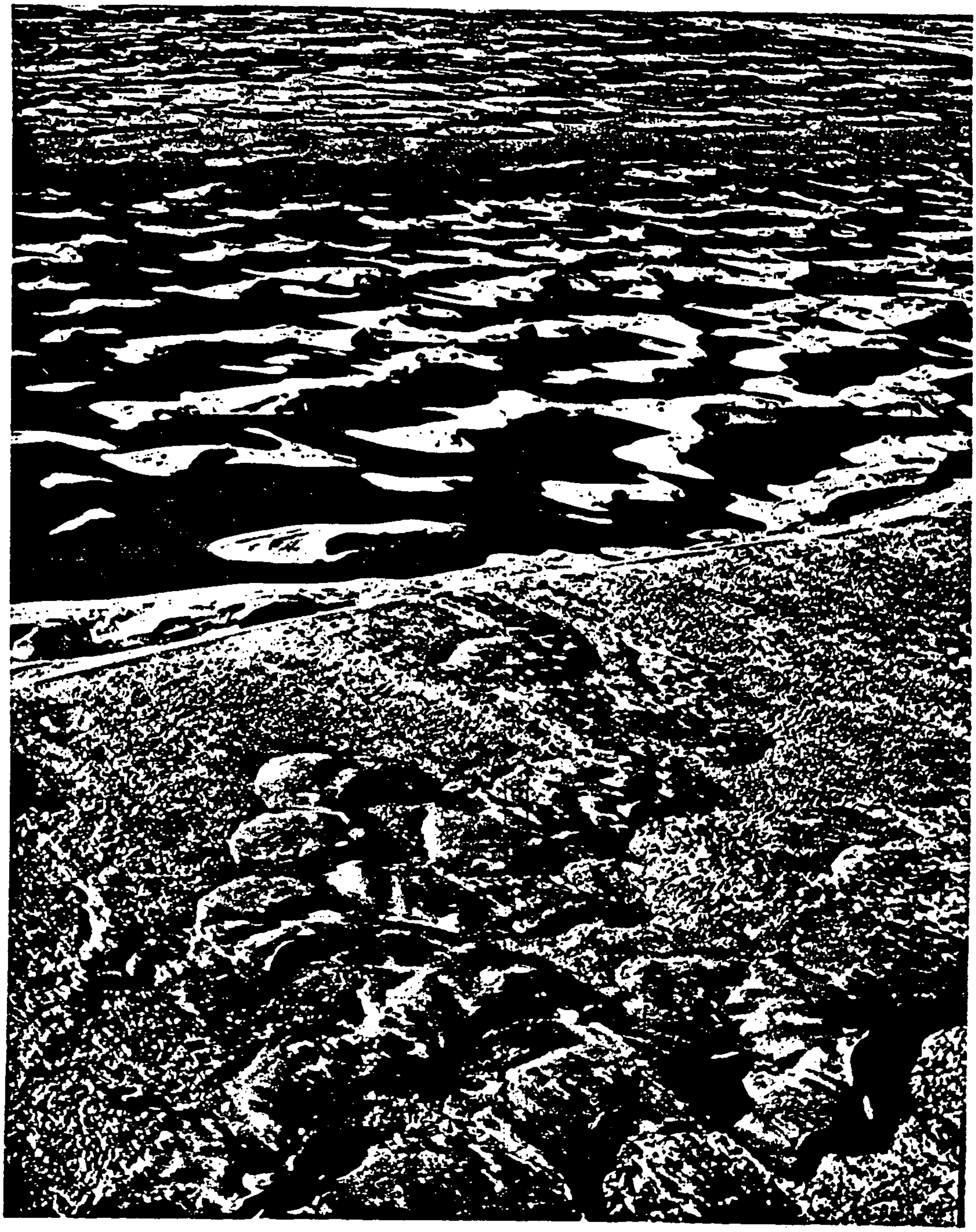


Figure 7.26. Oblique aerial view to the west of the central part of the sand dune field at the Potholes Reservoir. (photograph 2029 cr by John S. Shelton).

Lake section of about 5 x 10^6 m^3/sec. The maximum known thickness of sedimentary fill in the Quincy Basin is 40 meters (Bretz and others, 1956, p. 969).

The surface of the Ephrata Fan displays a series of lobate depositional units commonly bounded on their downstream margin by slipfaces (now inactive and somewhat modified) (Figs. 7.29 and 7.30). This surface morphology indi-

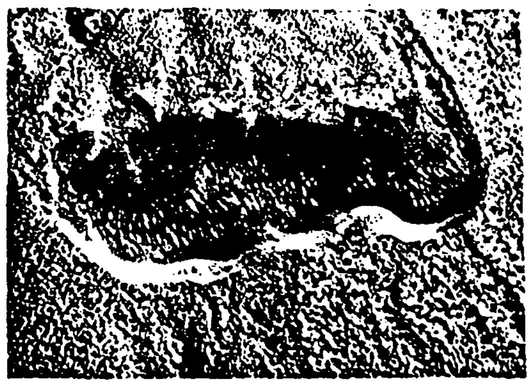


Figure 7.27. Vertical aerial view of a parabolic dune to the east of the Potholes Reservoir. Note the burial and re-excavation of linear sand ridges.

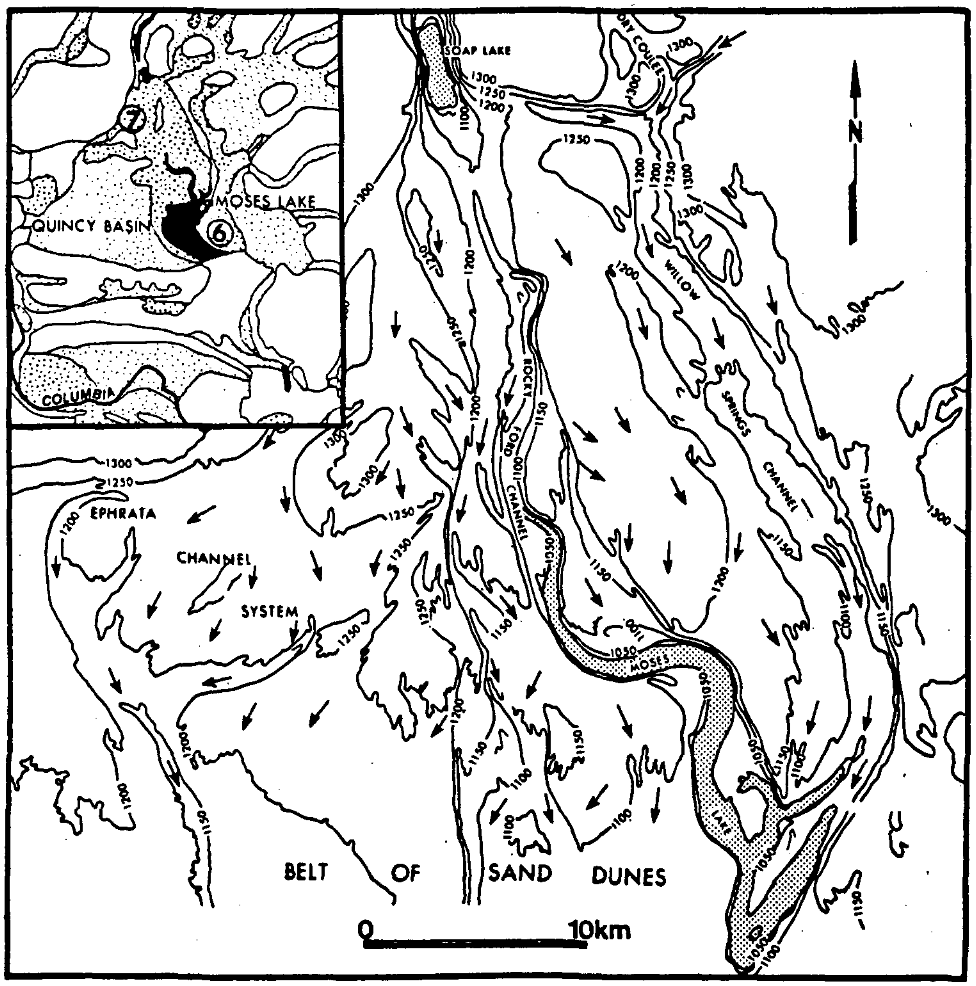


Figure 7.28. Topographic map of the Ephrata Fan complex in the upper Quincy Basin. Arrows indicate inferred surface flow directions. Modified from Bretz (1959, his Fig. 20, p. 33). Contours in feet above m.s.l.

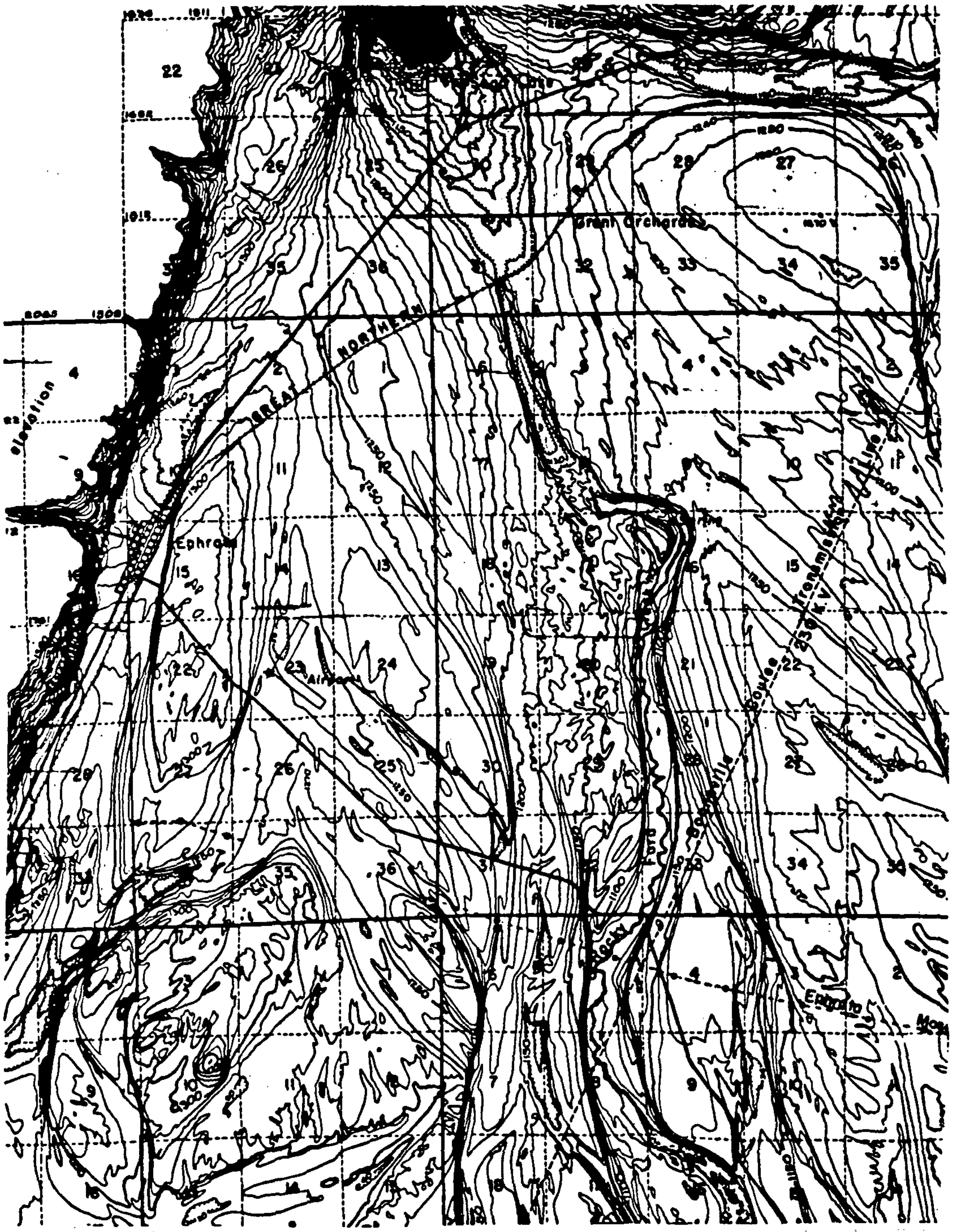


Figure 7.29. Part of U.S. Bureau of Reclamation topographic map of the Columbia Basin Project. Contour interval is 10 feet. Note the large linear bar to the east of Ephrata. The large diamond-shaped bar in the lower left corner is more typical of most of the gravel bars on the Ephrata Fan.

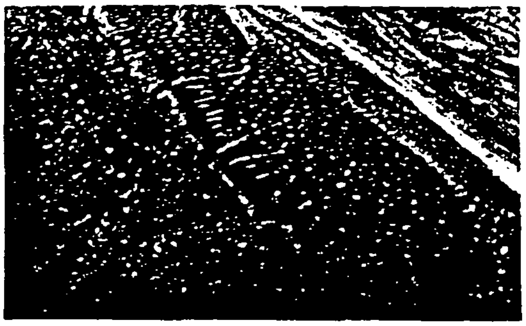


Figure 7.30. Oblique aerial view of a part of a linear gravel bar near the center of the Ephrata Fan. Flow across the bar came from the upper right to the lower left.

Figure 7.31. North-facing wall in the Columbia Concrete Products gravel pit near Ephrata. Cross-stratification dips toward the southwest. Note the concentration of large boulders at the bounding surfaces of the sets. Note man in the lower left for scale.

Figure 7.32. South-facing wall in the Columbia Concrete Products gravel pit near Ephrata. Large boulders are scattered throughout the deposit. Part of a slightly rounded basalt column is visible near the center of the photo.

cates that the Ephrata Fan was built by bar-slipface migration. The internal stratification observed in the gravel pit is consistent with this interpretation. The pit walls are characterized by thick cross-stratified sets with a dip toward the southwest (Fig. 7.31), reflecting deposition on bar slipfaces by currents flowing in a general direction from the center of the fan system across its western margin and into the Ephrata channels. Large boulders are found throughout the unit (Fig. 7.32. Most commonly, however, they are concentrated along the bounding surfaces of individual cross-stratified sets. This indicates that some scour occurred between the deposition of the individual sets, thus forming a surface armor prior to burial by a new bar slipface. The flood gravel is typically poorly sorted, with a number of broken rounds. Mud coating on the individual cobbles is ubiquitous (Fig. 7.33). The deposit is dominated by basalt fragments probably derived by erosion of the Grand Coulee. One can also find granitic rocks brought, at the very least, from the exhumed granite outcrops at the head of Grand Coulee near the present Grand Coulee Dam on the Columbia River, about 50 km to the north.

STOP 8. LENORE LAKE

The role of geologic structure in the Quaternary evolution of the Columbia Basin has been very significant (Baker, Ch. 2; Swanson and Wright, Ch. 3, this volume). Specifically, the

Figure 7.33. Close-up view of the open-work flood gravel in the Columbia Concrete Products pit at Ephrata. Note the mud coating.

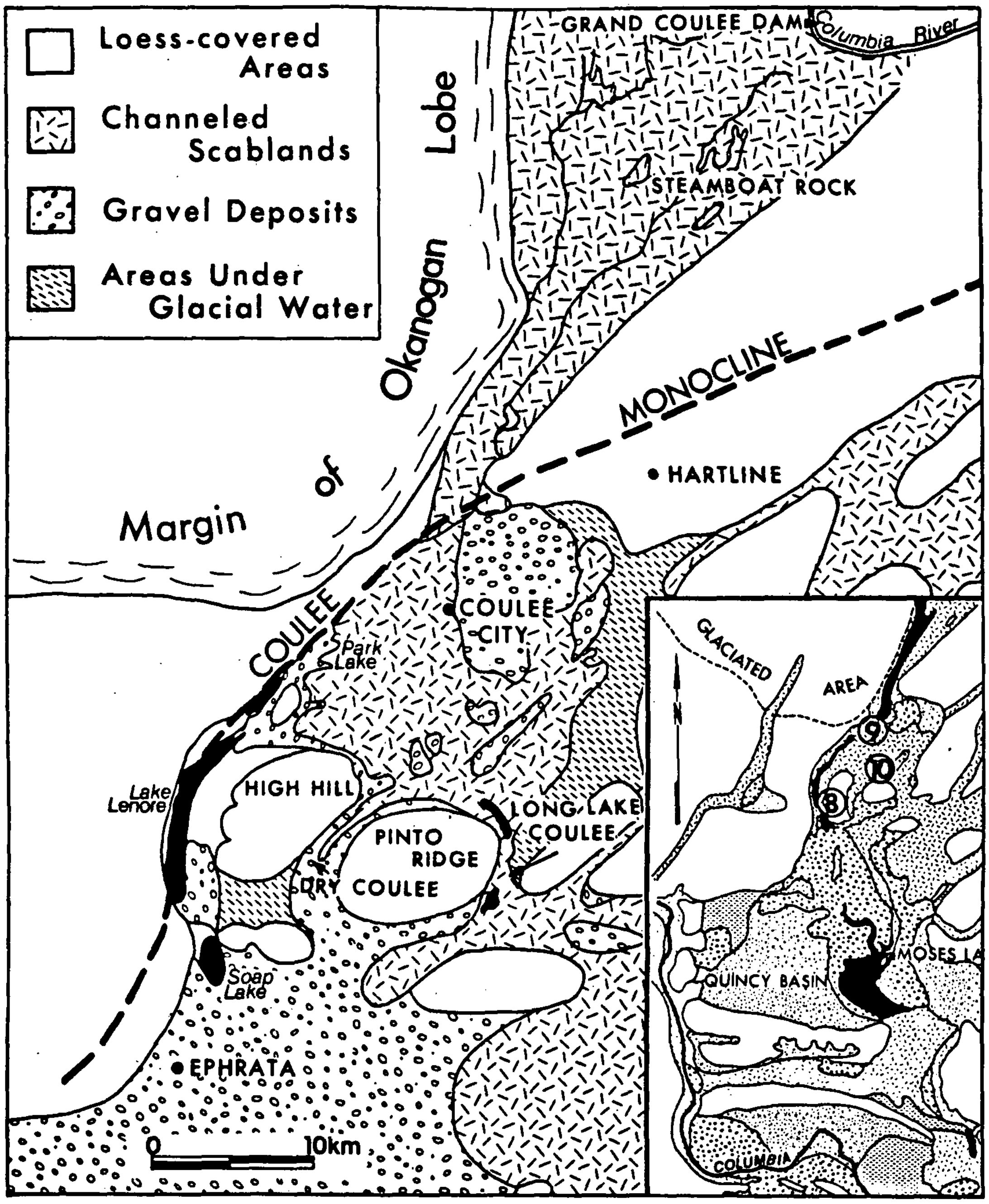


Figure 7.34. Generalized geologic map of the Grand Coulee region. Insert shows locations of stops 8, 9 and 10. Modified from Bretz (1959, his Plate 3).

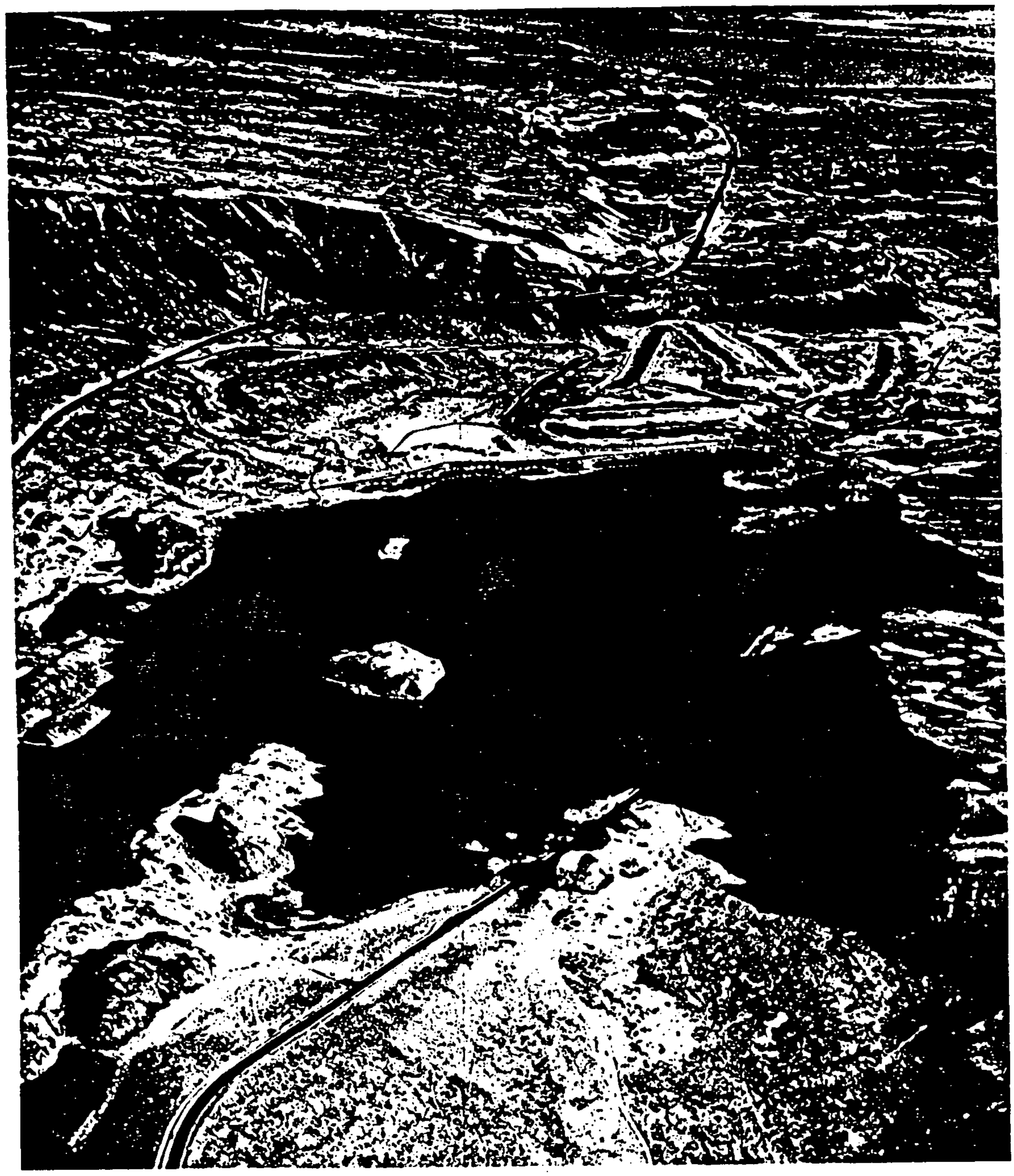


Figure 7.35. The Coulee Monocline to the north of Park Lake (see Fig. 7.37 for location). (Section of photograph 2001 by John S. Shelton.)

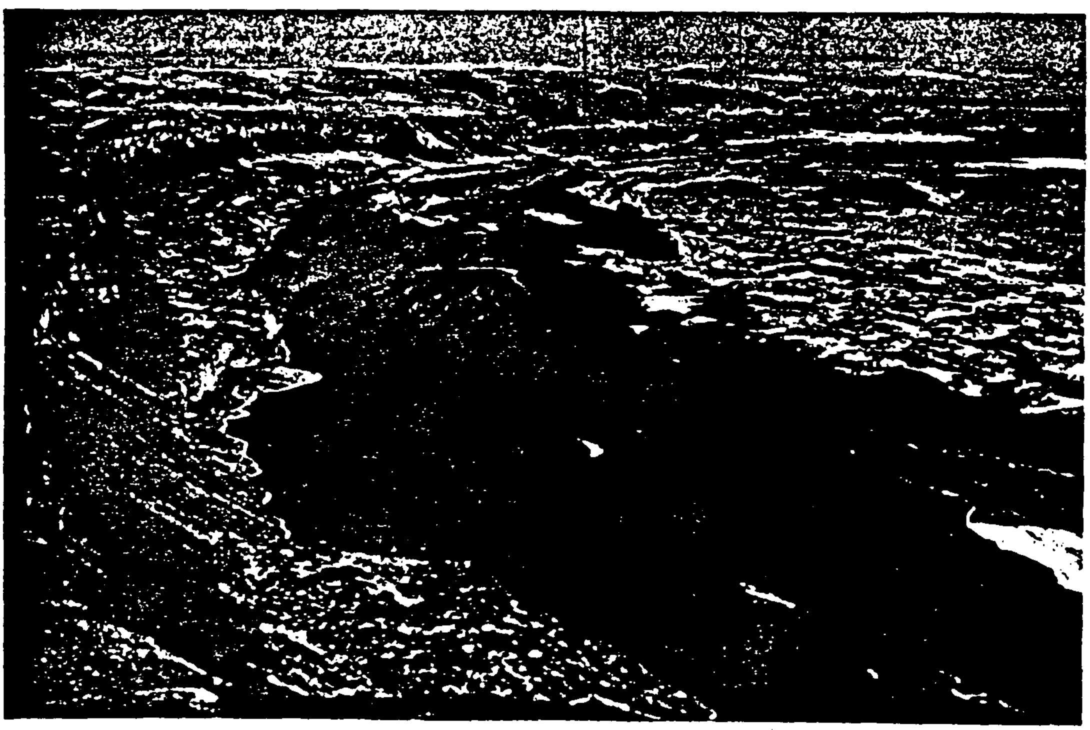
shape, location and, in part, size of lower Grand Coulee is related to the existence of the Coulee Monocline, a large flexure in the Yakima Basalt trending generally southwest-northeast through the western Columbia Basin (Fig. 7.34). For about 25 km, the course of lower Grand Coulee is superimposed directly on the Coulee Monocline. The weakened rocks on the steep (45° to 60° tilt) eastward-dipping limb of the monocline yielded more readily to the erosive action of the flood than the flat-lying beds further east. Thus, the zone of maximum flood erosion moved into the weakened belt, leaving a string of lakes as a result of deep sub-fluvial plucking action (Plate 3). Deep erosion of the monoclinal limb left the lower Grand Coulee with a western wall nearly 300 meters high. Pre-flood drainage ravines which descended the monoclinal limb were left as hanging valleys high above the present valley floor giving the skyline a gabled appearance. Steeply-dipping beds of the monoclinal limb are preserved in the hogback islands of Lake Lenore and sections of the coulee wall near Park Lake (Fig. 7.35).

Through the Lenore Lake section of the Grand Coulee, one can see a particularly well-developed inner channel (Fig. 7.36). Flanking the deep, mostly lake-filled, inner channel are benches with an irregular surface topography made of basaltic knobs and potholes, a characteristic result of sub-fluvial plucking in layered volcanic rock. Baker (Chs. 4 & 5, this volume) discusses how the turbulence pattern of high-velocity flow interacts with the structure of the basalt flow to produce this diagnostic scabland morphology.

STOP 9. DRY FALLS OVERLOOK

Dry Falls constitute the head cataract of lower Grand Coulee (Plate 3) carved by northward recession of the Lenore Canyon system. Extending from the overlook and east to Castle Lake (Fig. 7.37), this compound cataract reached a width

Figure 7.36. Oblique aerial view toward the north of Lenore Lake. Note the lake-filled inner channel and the hogback islands in the foreground.



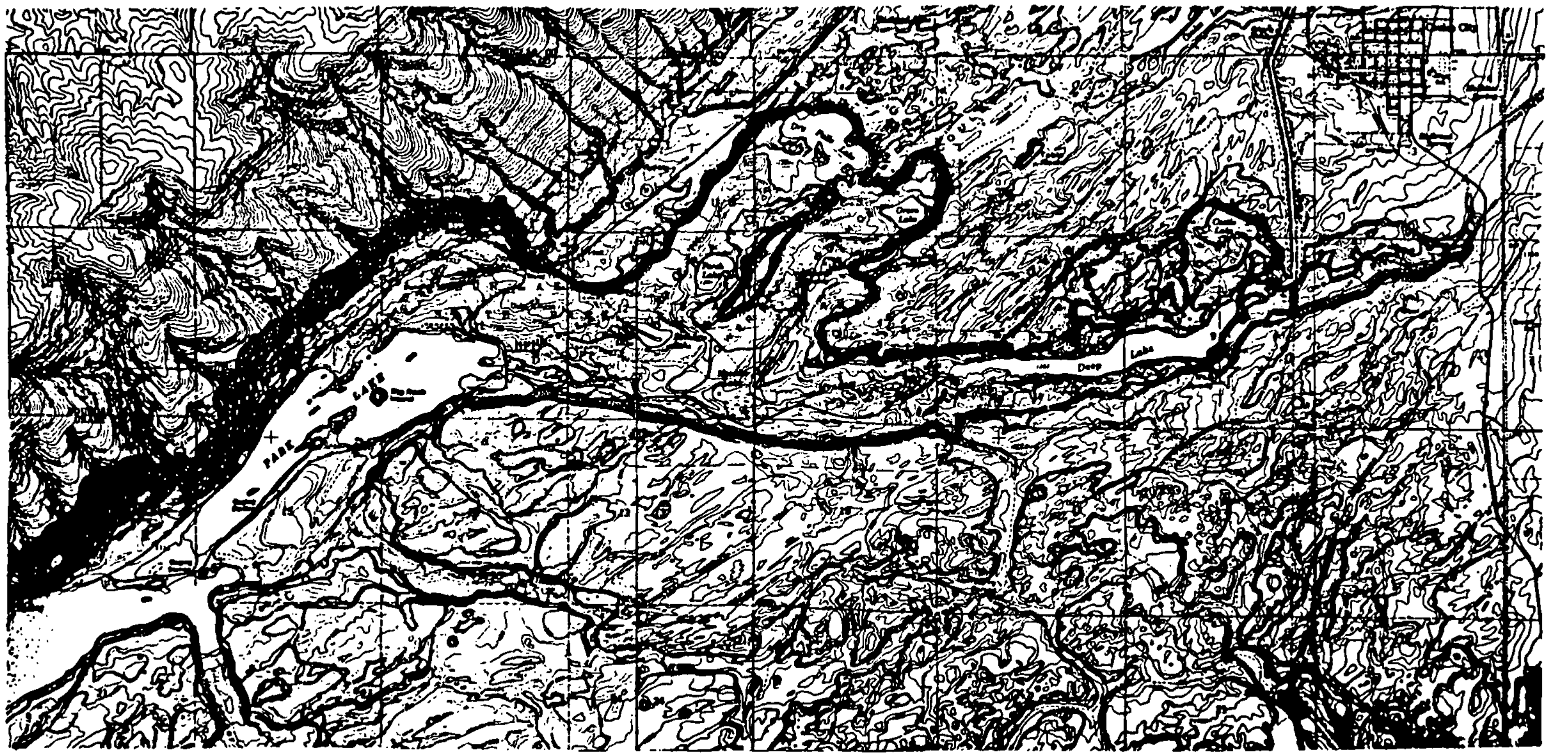


Figure 7.37. Sections of U.S. Geol. Survey topographic maps of Park Lake and Coulee City 7.5-minute quadrangles. Contour interval is 10 feet.

of nearly 6 km and a maximum height of 125 meters, substantially larger than the current Niagara Falls gorge. Bretz (1969, p. 524) points out that the cataract developed as the main thread of flood incision moved into the zone of weakened rocks along the Coulee Monocline and then was maintained as more and more of the discharge began flowing through this rapidly expanding canyon. Plunge pool lakes in alcoves at the base of the cliffs mark what appears to be the locations of discharge concentrations (Fig. 7.38). Large boulder piles can be seen immediately downstream of many of the plunge pools.

Isolated plateau remnants (e.g., Umatilla Rock, Fig. 7.39) have been left standing between more rapidly retreating recessional gorges. High-water marks along the margin of the Hartline Basin seem to suggest that many of the cataracts in the area might have been formed subfluvially rather than by the classical plunge pool undercutting (Baker, 1977, p. 401). This might apply to Dry Falls as well, although it is unclear whether the water depth remained adequate to maintain subfluvial recession at the time the Leonore Canyon system had been eroded to the present depth.

Immediately upstream of the cataract, as well as east of Park Lake, the scabland plateau is covered with a series of parallel longitudinal grooves ranging from 30 to 60 meters in spacing and measuring up to 3 meters in height (Figs. 7.40 and 7.41). The grooves generally parallel the flow direction and cut across any structural trends. They are formed on the resistant basalt

Figure 7.38. Oblique aerial view to the northeast of Dry Falls cateract. Coulee City is visible in the background. Large alcoves, plunge pools and boulder piles characterize the falls. Umatilla Rock, an isolated plateau remnant is visible on the right (photograph 5602 by John S. Shelton).

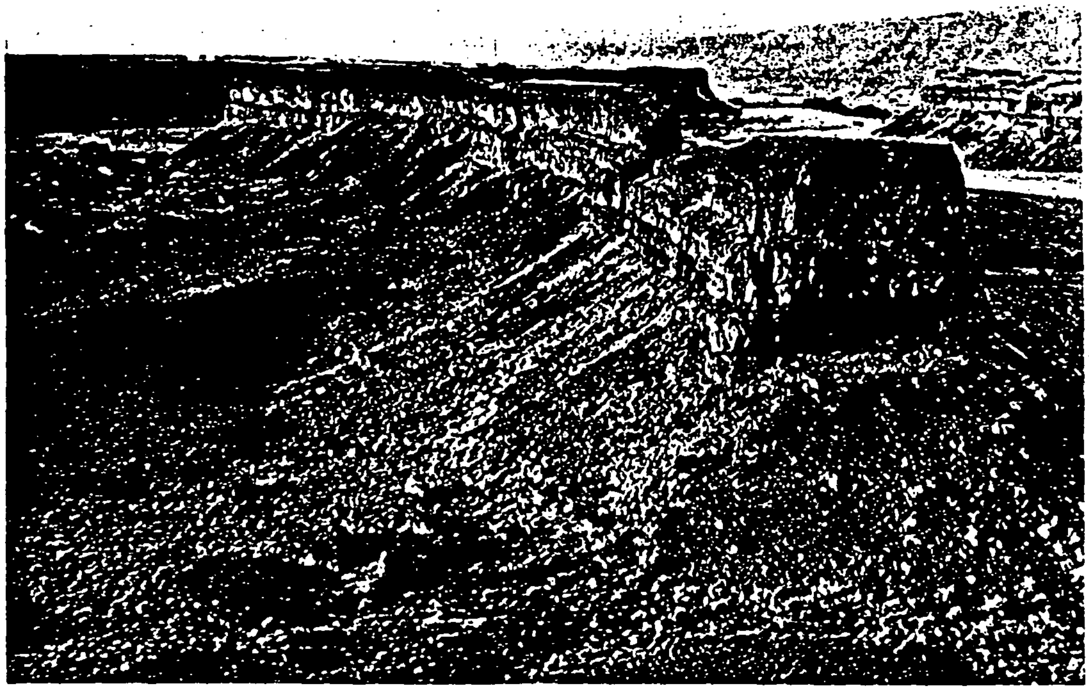


Figure 7.39. Umatilla Rock, an isolated plateau remnant in the middle of the major Dry Falls cataract (for location see Fig. 7.37).

Figure 7.40. Oblique aerial view of longitudinal grooves in basalt north of Dry Falls. The grooves are spaced about 50 meters apart.

Figure 7.41. Ground view of the grooves north of Dry Falls. Note man for scale.

entablature (Baker, 1977, p. 401) and may owe their origin to longitudinal roller vortices which are known to have formed similar grooves in laboratory experiments (Shepherd, 1972; Allen, 1971a) (Fig. 7.42). Similar grooves often are observed in swift-flowing bedrock streams under upper flow regime conditions.

Although this field trip will not visit the upper Grand Coulee (Banks Lake), its relationship to the lower Coulee should be briefly reviewed. As first proposed by Bretz (1932a), upper Grand Coulee was eroded across a previously intact divide. In contrast, lower Grand Coulee was excavated along the axis of the Coulee Monocline. The canyon system crosses the trace of the monocline a few miles north of Dry Falls (Fig. 7.34). As flood waters detoured out of the Columbia River Valley at the east margin of the Okanogan Lobe (Baker, Ch. 2, this volume), crossed the Waterville Plateau, and cascaded some 250 m down the southeast slope of the Coulee Monocline, a recessional cataract was developed which gradually migrated more than 30 kilometers across the Waterville Plateau to the south wall of the Columbia valley near the present Grand Coulee Dam (Fig. 7.34). Although the cataract ceased to exist at this point, the upper Grand Coulee continued to act as a major channel leading flood water into the Hartline Basin and lower Grand Coulee. Flow through upper Grand Coulee eroded nearly 300 meters into the pre-flood divide. Scabland surfaces on the plateau adjacent to the coulee indicate that the initial flow covered an area nearly 15 km wide at the northern side of the divide. As the cataract receded and deepened, the channel was narrowed to little more than 2 km immediately north of the Coulee Monocline, widening to about 8 km in the Steamboat Rock-Northrup Canyon area (Fig. 7.34). Between Steamboat Rock and the Columbia River, erosion exposed a series of granite knobs. Upper Grand Coulee today is filled by the Banks Lake reservoir.

STOP 10. PINTO RIDGE

The north flank of the Pinto Ridge anticline provides a good view of the complex erosional and depositional features of the Hartline Basin to the north. This broad topographic basin is

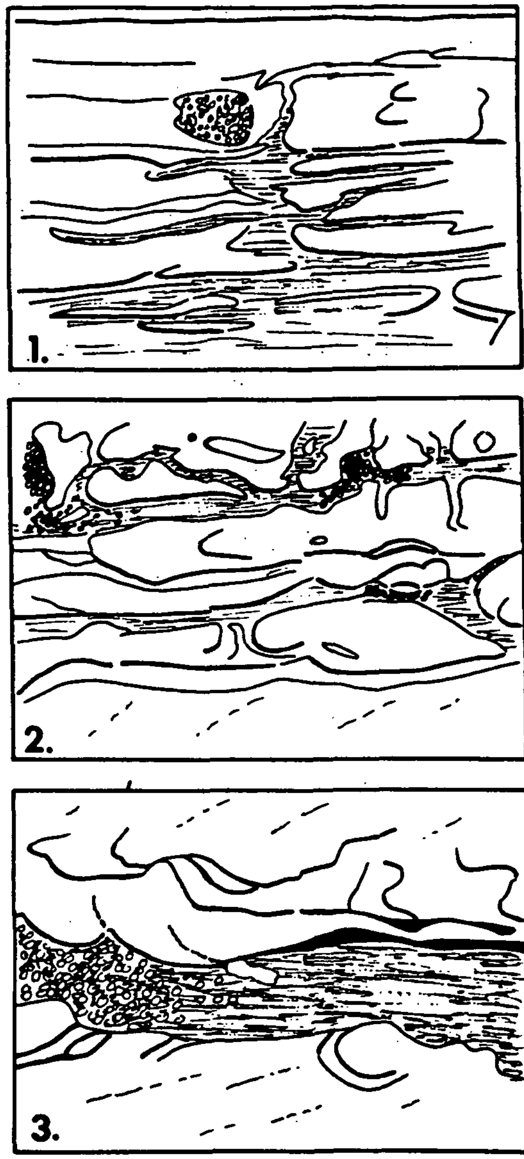


Figure 7.42. Development of experimental scour features in consolidated material through time. 1. Development of linear grooves by longitudinal vortices. 2. Groove enlargement and reduction in number by non-uniform shear stress at the bed. 3. Coalescence of grooves into single channel. Each frame measures about 30 x 40 cm. Redrawn from Shepherd and Schumm (1974, their Fig. 4).

actually a combination of two broad synclines. The Hartline Syncline is bounded by the Coulee Monocline to the north and the High Hill-Trail Lake Anticline to the south. South of the Trail Lake Anticline is the Bacon Syncline (Bretz and others, 1956, p. 968) which is in turn bordered to the south by the Pinto Ridge Anticline (Fig. 7.34).

The Hartline Basin was entered by debris-laden flood water from the upper Grand Coulee and the Almira-Wilbur Coulee to the northeast. With flow divergence into the basin, some of the sediment was dropped to form a large fill between Coulee City and the town of Hartline (Fig. 7.34 and Plate 4). Isolated gravel remnants along the eastern margins of the basin (Plate 4) reach elevations as great as 515 meters where they have buried loess deposits with a calichified upper zone. Bretz and others (1956, p. 968) argue that this elevation of flood gravel demonstrates that it was deposited before the three coulees forming the present outlets of the basin had been carved. Lowering of the general water level in the Hartline Basin subsequent to the formation of Lenore Canyon, Dry Coulee and Long Lake Coulee (Fig. 7.34) caused general erosion of the eastern basin fill leaving major remnants only north of Long Lake.

The pattern of coulee development around the north flanks of High Hill (Plate 3) and Pinto Ridge (Plate 4) suggests that as the flow diverged around these "obstacles," water simultaneously occupied all three outlets of the Hartline Basin. Bottom gradients, the narrowness of channels and the lack of backfill bars support this idea (Bretz and others, 1956, p. 969). Baker and Milton (1974) pursued the idea by suggesting that these two hills affected flow patterns of water draining from the Hartline Basin in a manner similar to that of crater "obstacles" at the mouth of Ares Vallis on Mars.

The scour crescents around the Martian craters, however, are morphologically more similar to those commonly observed around bridge piers and streambed boulders on earth. Such obstacle marks are generally attributed to horseshoe vortex patterns (Shen, 1971). In terms of scale, on the other hand, the Martian scour features correspond to the coulees around High Hill and Pinto Ridge.

The Hartline Basin and lower Grand Coulee offer perhaps the most spectacular erosional topography of the entire Channeled Scabland (Fig. 7.43). The multitude of small and large coulees, cataracts, anastomosing channels and butte-and-basin topography (Plates 3 and 4) provide an assemblage of forms which are an unlikely product of any process but that of a catastrophic flood. Complex vortices, intense turbulence and high pressure gradients are required to pluck and undermine the bedrock of the channel floors. Baker (Ch. 4, this volume) provides details of the postulated hydrodynamics responsible for this scabland topography.

STOP 11. CRAB CREEK

Crab Creek Coulee follows one of the major pre-flood east-west drainage lines across the central Columbia Basin. During the Missoula Flood, it was a major distributary from the Cheney-Palouse tract westward to the Quincy Basin (Fig. 7.1). Its discharge increased toward the west as Lake Creek, Canniwai Creek and Wilson Creek added their flows successively. Baker (1973b) has made a comparison of pre-flood and Missoula discharges of Crab Creek just upstream of the Wilson Creek junction. As seen in Plate 5, the pre-flood valley meanders are preserved in the Yakima basalt walls. The average wavelength is about 2000 meters which indicates a bankfull discharge of about 850 m^3/sec. Baker (1973a) estimated from a series of hydraulic geometry measurements that the peak discharge of the Missoula Flood was about 2.8×10^6 m^3/sec, an increase over pre-flood discharge by a factor of more than 3000!

This central part of Crab Creek displays spectacular examples of two features which provide convincing evidence of a flood origin for the Channeled Scabland: large gravel bars and giant current ripples. Along the section of Crab Creek shown in Plate 5, a series of distinct bars can be observed. Their overall topography suggests subfluvial origin. They are generally situated on the valley floor, streamlined in conformity with the bedrock valley outlines, and show an overall convex cross profile. The giant ripples further corroborate the idea. These are all asymmetric downstream and have steeply dipping internal cross-strata. Where Route 28 cuts longitudinally through a large bar a few kilometers west of Wilson Creek, one can see beautiful ripple cross-sections. Surface undulations are somewhat subdued due to deposition of post-flood loess, although aerial photos still clearly reveal their existence (Fig. 7.44 and 7.45). Topographic maps issued by the U.S. Bureau of Reclamation provide excellent data on the geometry of these and many other ripple fields.

The dominant geometric pattern of the Channeled Scabland seen in LANDSAT images (Ch. 8, this volume) or generalized geologic maps (Fig. 7.1) is that of large-scale channel anastomosis. The flow pattern here at Crab Creek clearly documents the origin of such geometry. The largely east-west trending pre-flood drainage ways on the plateau carried the brunt of the Missoula Flood from the Cheney-Palouse tract into the Quincy Basin. With a discharge increase of a factor of 3000 above normal bankfull flow, however the entire valley had inadequate capacity. Flood water spilled across interfluve devides as a necessary consequence, causing the establishment of new scabland channels along the generally southwest trending tributaries of the main rivers. The coarse pattern of anastomosis, characterizing the central Channeled Scabland between the Columbia River in the north and the Snake River in the south, was generally developed by such accentuation of pre-flood drainage.

STOP 12. ROCKY FORD CREEK

The general processes responsible for the development of the Ephrata Fan were discussed at Stop 7. As indicated there, accumulations of

Figure 7.43. Oblique aerial view of Deep Lake, the easternmost cataract of the Dry Falls system (see Fig. 7.37 for location). Coulee City and the upper Hartline Basin visible in the background (photograph 5600 by John S. Shelton).

gravel at the boundary surfaces of individual cross-stratified sets demonstrate episodes of surface scour even during the general constructional phase of fan development. The large contiguous area of boulder armor along the west side of Rocky Ford Creek (Plate 2) also indicates extensive scour, as does the fact that this surface is, on the average, 15 meters below the finer-grained fan surface to the west (Fig. 7.29). It has been suggested that this extensive scour, as well as the development of Rocky Ford Creek itself, was caused by high-velocity currents, associated with drainage of the Quincy Basin, during the waning stages of the flood (Baker, 1973a). Such a hypothesis is consistent with what is known about channel development on braided stream bars during falling river stage (Smith, 1971; Nummdal and others, 1974). There is reason to believe that the Ephrata channel system to the west and the Willow Springs channels to the east (Fig. 7.28) owe their origin to the same basin drainage.

An indication of the current velocities attained during this phase of basin-fill incision can be gained from the pattern of scour around individual large boulders on this armor surface. A large boulder to the north of the road to the fish hatchery (Fig. 4.15) has a prominent scour crescent on its northwest (upstream) side as well as a large elliptical scour hole to the southeast (Baker, Ch. 4, this volume). This scour pattern can best be accounted for by a dual vortex system generated by this blunt obstacle to the flow. The upstream scour probably owes its origin to the flow separation caused by strong pressure gradients with the consequent development of a horseshoe vortex system. Shen (1971) suggests that this vortex system is largely responsible for

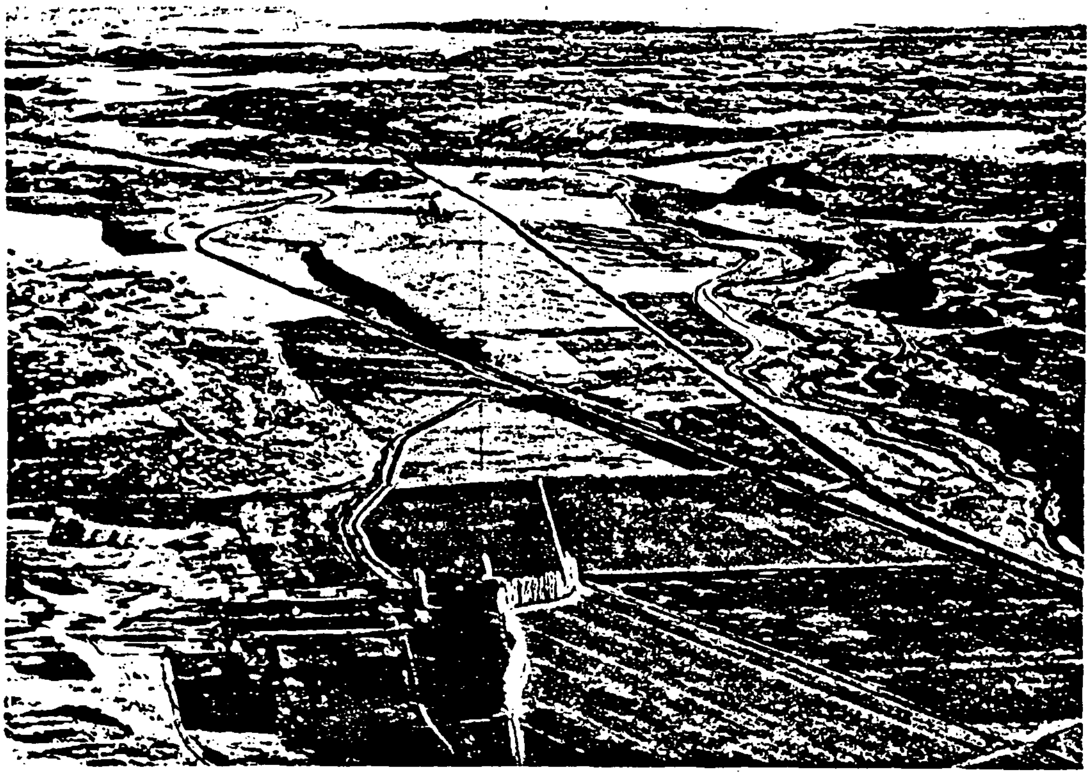


Figure 7.44. Oblique aerial view to the east along Crab Creek. The town of Wilson Creek is visible in the upper left. Washington Route 28 cuts across a large mid-coulee bar in the center of the photo exposing a series of good ripple cross-sections (photo 2052 cr by John S. Shelton. Reproduced from "Geology Illustrated" by John S. Shelton. W. H. Freeman and Company. Copyright © 1966.).

upstream scour around bridge piers and Karcz (1968) attributes common current crescents to the same phenomenon. This author has observed large horseshoe vortex systems around stranded icebergs during a glacier burst (jökullhlaup) in Iceland.

The downstream scour is probably due to a wake-vortex system generated in the lee of the boulder. This vortex system can be either erosional or depositional depending on the flow Reynold's number. At low Reynold's number (low current velocity) deposition would most likely occur behind the obstacle. Flow at high Reynold's numbers might generate a wake-vortex capable of removing even coarse sediments (Shen, 1971). The presence of this downstream elliptical scour behind this and other boulders on the Ephrata Fan, therefore, is suggestive of rather strong currents during the drainage of the Quincy Basin (Baker, 1977, p. 407).

STOP 13. GEORGE GRAVEL PIT

The deduced implications of the stratigraphy in this western Quincy Basin are largely based on unpublished data by George Neff, summarized by Bretz and others (1956, p. 985), Bretz (1969, p. 529) and Webster and others (1976, p. 11). Gravel exposed at George records pre-Bull Lake flooding of the Columbia Basin. At least two episodes of early flooding have been recognized by Neff. This stop exposes the older of the two gravels. One of the deposits may be contemporaneous with the flood responsible for pre-Bull Lake gravels at the Marengo railroad-cut (this chapter, Stop 3), although no certain correlation between old flood gravels in the Cheney-Palouse tract and the western Channeled Scabland has yet been established (Baker, Ch. 2, this volume). The record consists of typical coarse, open-work flood gravels with foresets dipping to the south-

Figure 7.45. Oblique aerial view to the south of giant ripples at the town of Wilson Creek. Flow was from left to right.

east, *away from* the head of the Potholes cataract to the west (Fig. 7.46). Paleoflow directions based on foresets in this rather localized exposure may not be too reliable. However, the implied south-eastward flow gains credence when regional grain size trends and lithologic characteristics of the sediments are taken into account. The exposures contain boulders of a gray-green tuffaceous silt and sand which can only come from the Vantage standstone member of the Yakima Basalt (see Swanson and Wright, Ch. 3, this volume). The nearest outcrops are at Babcock Ridge, 10 km to the west. Also, the grain size of flood gravel decreases in an eastward direction. At Winchester Wasteway, 10 km east of George, flood debris is mostly sand and gravel with the largest cobbles only 10 cm in diameter. East of Winchester

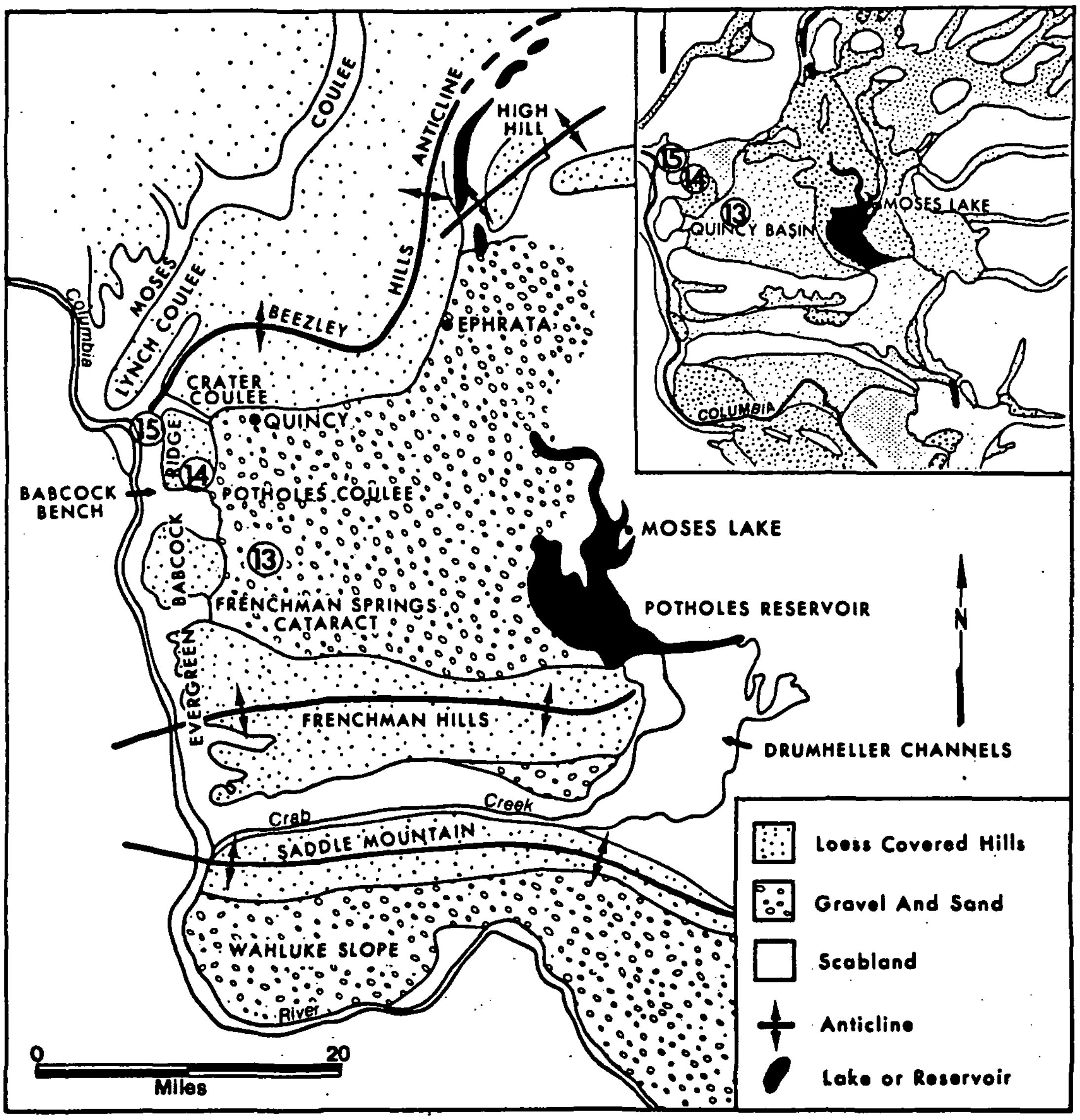


Figure 7.46. Schematic geologic map of the western Quincy Basin and the Babcock Ridge spillways. Field stops 13, 14 and 15 are shown.

Wasteway, older flood gravel is buried by deposits of the last (early Pinedale) Missoula Flood which appear to have filled the bulk of the Quincy Basin.

The uppermost meter of gravel in the deposit at George is capped by horizontally laminated caliche. This is underlain by 30 to 50 cm of carbonate cemented gravel. Local carbonate cementation in the coarse gravel continues to a depth of 2.5 meters. Individual cobbles are carbonate-coated on their underside down to a depth of 3.5 meters. In the upper 2 meters, the weathering rind on the basalt cobbles may exceed 8 cm in thickness. Below 3 meters, however, the gravel shows no evidence of weathering. Based on this soil development, Richmond and others (1965) proposed a pre-Bull Lake age for the deposit.

STOP 14. POTHOLES COULEE

The Quincy Basin is separated from the Columbia River valley to the west by Babcock-Evergreen Ridge, which extends from the Beezley Hills in the north to the Frenchman Hills in the south. The ridge summit attains an altitude of 550 meters, about 100 meters above the average altitude of the sedimentary fill of the Quincy Basin to the east (Fig. 7.46). As the flood water rose in the Quincy Basin, it spilled across saddles in this ridge and cascaded down the east side of the Columbia River valley, the floor of which today is 300 meters below the ridge crest. The spillover created three major recessional cataracts. From the south to the north, these are: the Frenchman Coulee, Potholes Coulee and Crater Coulee (Plates 6 and 7). The latter joined Lynch Coulee before entering the Columbia River.

Bretz has repeatedly used these coulees as one of his main arguments for a flood origin of the Channeled Scabland. In addition to their large size and perpendicular orientation relative to the Columbia River, it is significant to note that each spillway begins at very nearly the same altitude (Bretz, 1959, p. 26). This fact strongly suggests that they were all contemporaneous in origin, requiring that the *entire* Quincy Basin was filled with water up to this elevation at one time. Other authors have suggested alternative ideas to make the spillways sequential in operation and thus avoid such enormous quantities of water. The ideas include tectonic movements such that, after one cataract had formed, another part of the ridge was depressed to draw the discharging water off from one cataract to make a new one. Other ideas include large gravel fills subsequently eroded away, or strategically located ice-jams. None of these ideas has any independent supporting field evidence.

The support for the flood origin of these three coulees becomes even stronger when one considers the fact that a fourth discharge way out of the Quincy Basin, the large Drumheller Channels complex in the southeastern corner of the basin (Plate 8), also has an elevation at its upper limits equivalent to that of the coulees through the Babcock Ridge.

The largest of the Babcock Ridge cataracts, Potholes Coulee (Plate 6) is about 2 km wide, about 150 meters deep and has receded about 3 km into the ridge from the wall of the Columbia Valley above Babcock Bench (Fig. 7.46). A broad region of eroded scabland topography east of the coulee is evidence for the convergence of flood water from the Quincy Basin into the two main cataracts of the coulee. Deep plunge pools and gravel mounds make up the floor of both cataracts (Fig. 5.25). Alignment of the main cataract's head wall appears controlled by the joint pattern in the basalt (Fig. 7.47).

Along the east side of the Columbia River at the elevation of the floor of the two recessional cataracts of Potholes Coulee is Babcock Bench, a basalt scabland surface about one kilometer wide extending nearly 15 km along the river. The channel width and depth expressed by this bench is an indication of the discharge carried by this reach of the Columbia River during the Missoula Flood.

This stop provides excellent exposures for the study of the basalt and associated lithologies of the Yakima Subgroup. A typical basalt flow has a scoracious upper portion grading into the upper colonnade of relatively small and short columns. In many flows, this upper colonnade may be missing altogether. Underlying this is the entablature, consisting of long slender, often twisting and flaring columns. This upper part of the lava flow is generally the one most resistant to erosion. Underlying the entablature is the lower colonnade, made of coarse, 1 to 3 meter diameter, columns (see Fig. 5.37 and 4.14). Depending on whether the lava was emplaced in a lake, the base of the flow is often a pillow-palagonite complex (Fig. 7.48) with gas chimneys.

Between the emplacement of many of the individual basalt flows, enough time elapsed to permit weathering, erosion, the growth of forest cover, and lake formation (Swanson and Wright, Ch. 3, this volume). As a consequence, local sedimentary intercalations in the basalts include conglomerate beds, clay layers and fresh-water diatomite (Fig. 7.49). Pieces of petrified wood from the interbasalt lake beds can be found.

Figure 7.47. East wall of the main cataracts at Potholes Coulee. Its alignment appears to be controlled by a major north-south trending joint. Note the deep water-filled plunge pools on the right (photo by Larry G. Ward).

STOP 15. CRESCENT BAR

Crescent Bar (recreation area) is located where the flood water discharging down Crater and Lynch Coulees joined the Columbia River valley at the northern terminus of the Babcock Bench (Plate 7). Directly across the river is West Bar, a large flood-generated point bar covered with an extensive set of giant current ripples (Fig. 7.50). From Route 28 immediately north of the Crescent Bar exit, one has a good view of the surface of West Bar.

The mouth of Lynch Coulee contains evidence both relating to the (1) relative age of flooding

Figure 7.48. Pillow lava and palagonite overlying altered lacustrine sediments in a minor road-cut near the head of Potholes Coulee.

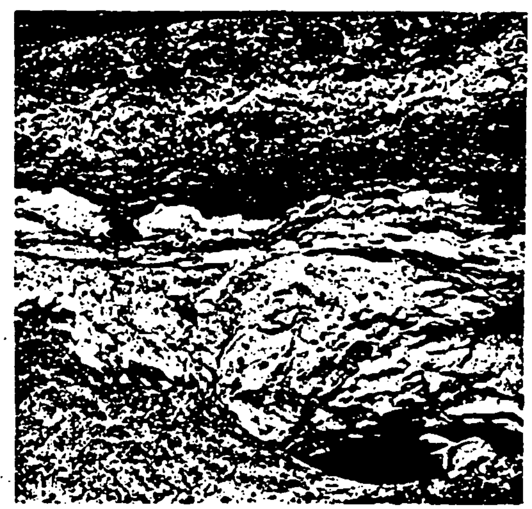


Figure 7.49. Interbasalt diatomite beds exposed near the head of Potholes Coulee.

down the Columbia River Valley and the Quincy Basin and (2) concerning the absolute age of the most recent flood.

According to Waitt (1977b, p. 15), there are two gravel units exposed in Lynch Coulee upstream of the Hyw. 28 culvert. The lower gravel unit has foreset bedding dipping up the coulee. The upper unit has foresets which all dip down the coulee. There is no evidence of weathering on top of the lower gravel, suggesting that both units are of similar age if not contemporaneous. Waitt concludes that both units date from the last major Missoula Flood. The first surge of this flood carried sediment up Lynch Coulee and perhaps into the Quincy Basin; the later part of the flood carried gravel down the coulee into the Columbia Valley.

In the gravel pit east of the Crescent Bar road, there is exposed a fine pebble gravel with northward dipping foresets and an up-Columbia provenance throughout. This is interpreted by Waitt as indicative of a huge counterclockwise eddy that formed in this alcove when the main flood current flowed down-valley over West Bar (across the river).

Overlying the flood gravel are 2 meters of gravel, sand and laminated fine sand and silt diagnostic of temporary slack-water deposition, perhaps in a lake hydraulically ponded behind Wallula Gap. Near the base of the finer unit are three layers of tephra. (Waitt, 1977b, p. 15). This tephra has been interpreted as the Mount St. Helens' Set S ash (Fig. 2.16). As the tephra is equivalent to or only slightly younger than the 13,000 years B.P. for the last major Missoula Flood (Mullineaux and others, 1977). Waitt offers the tantalizing speculation that the contemporaneity of the rather short-lived flood and the Mount St. Helens eruption might imply that the weight of the flood water could have triggered the eruption.

Figure 7.50. Oblique aerial view to the south of West Bar on the Columbia River across from Lynch Coulee. Giant gravel ripples demonstrate flood-flow toward the top of the picture, i.e. down the Columbia Valley. Babcock Bench extends along the river in the upper left of the photo (photograph 5595 by John S. Shelton).

Table 7.1. List of Geographic Coordinates of the Field Trip Stops

Stop	Map. Quad.	Latitude	Longitude	Township	Range	Sect.
1. Williams Lake	Cheney 15'	47°17'N	117°42'W	T21N	R40E	25
2. Macall	Macall 7½'	47°3'30"N	118°05'W	T18N	R38E	18
3. Marengo	Marengo 7½'	47°01'N	118°12'30"W	T18N	R37E	31
4. H U Ranch	Haas 15'	46°40'N	118°17'W	T14N	R36E	27
5. Snake/Palouse Jct.	Starbuck 15'	46°35'N	118°13'W	T13N	R37E	19/30
6. Sand dunes	Sieler 7½'	47°04'N	119°14'W	T18N	R29E	7
7. Gravel pit	Ephrata 7½'	47°18'N	119°33'W	T21N	R26E	22
8. Lenore Lake	Ephrata 15'	47°28'N	119°31'W	T23N	R26E	26
9. Dry Falls	Coulee City 7½'	47°36'N	119°21'W	T24N	R28E	6
10. Pinto Ridge	Stratford 7½'	47°30'N	119°20'W	T23N	R28E	17
11. Crab Creek	Wilson Creek 15'	47°25'N	119°06'W	T22N	R30E	7
12. Rocky Ford Creek	Grant Orchards 7½'	47°19'N	119°28'W	T21N	R27E	8
13. George	George 7½'	47°06'N	119°52'W	T19N	R24E	31
14. Potholes Coulee	Babcock Ridge 7½'	47°09'N	119°56'W	T19N	R23E	9
15. Crescent Bar	Babcock Ridge 7½'	47°12'N	119°59'W	T20N	R23E	19

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Chapter 8

Aerial Field Guide

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There are two overflights planned for the field conference; one for the Cheney-Palouse tract of the eastern Channeled Scabland, the other covering the coulees and basins of the western region. The approximate flight lines are indicated on the accompanying LANDSAT images (Figures 8.1 and 8.2).

The first flight will leave Spokane International Airport; the approximate location of flood spillover from the Spokane River Valley into the Cheney-Palouse tract. The elevation here is about 175 meters above the floor of the present Spokane Valley. The flight will follow the eastern margin of this large scabland tract, passing a series of loess remnants, gravel bars and excavated rock basins. At the junction of the Palouse and Snake Rivers, one can observe the erosion of deep, joint-controlled canyons and the deposition of high-level gravel bars. The return flight will pass Devils Canyon, Washtucna and Staircase Rapids, a multi-level erosional cataract caused by a major flow constriction at the southern end of the Cheney-Palouse tract. Other excavated rock basins and streamlined loess remnants with associated gravel bars will be observed on the last part of the flight past Sprague Lake back to Spokane.

The western scablands overflight will provide a review of the structurally controlled complex pattern of large-scale erosion and deposition characteristic of the region between the upper Grand Coulee (Banks Lake) and the Pasco Basin. The flight will depart Grant County airport, cross the central Ephrata Fan and then head east along Crab Creek to Wilson Creek and Marlin. From here, it will turn west to Long Coulee and enter the Hartline Basin at the north flank of Pinto Ridge. The most dramatic erosional topography of the entire Channeled Scabland will be seen in this part of the flight across the Hartline Basin to Banks Lake and south across Dry Falls down the lower Grand Coulee to its efflux section at Soap Lake. The flight will then skirt the western margin of the Ephrata Fan and cut west, across the Beezly Hills Anticline, to the Three Devils Cataract in Moses Coule. The flight down Moses Coulee will show its well-developed inner channel segments and the complex bar at the junction with the Columbia River. Flying south along the Columbia River, one will see, on the left, the three major western spillways out of the Quincy Basin, Crater, Potholes and Frenchman Coulees. At Sentinel Gap, the river has developed a deep antecedent cut through the Saddle Mountains Anticline. South of this mountain ridge, one finds large flood gravel ripples and modern active sand dunes on the Wahluke Slope. From here, the flight will return to Grant Co. Airport via the large southeastern spillway out of the Quincy Basin (the Drumheller Channel system) and the active sand dune field of the region bordering the Potholes Reservoir.

Figure 8.1. LANDSAT image of the Cheney-Palouse scabland tract with the tentative flight line superimposed. NASA LANDSAT E-5 854-16504-5, 20 August 1977.

Figure 8.2. LANDSAT image of the western parts of the Channeled Scabland with the tentative flight line superimposed. NASA LANDSAT E-2 936-17451-5. 20 June 1976.

Chapter 9

The Touchet Beds of the Walla Walla Valley

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The Touchet beds, named by Flint (1938b) extend along the Columbia River and its tributaries in the region of the Pasco and Umatilla Basins in Washington and Oregon (Fig. 9.1). The type locality for the Touchet beds is the vicinity of Touchet, Washington in the Walla Walla Valley (Flint, 1938b). These deposits were also described by Bretz (1928c, 1929, 1930b), Allison (1933), Lupher (1944) and Baker (1973a). The Touchet beds were deposited in Lake Lewis; the history of the idea of a Pleistocene lake east of the Cascades was summarized by Bretz (1919) and Flint (1938a).

The Touchet beds are as much as 100 m thick (Jones and Deacon, 1966) in the centers of valleys and basins but taper to a thin veneer as they rise to a maximum elevation of 350 m (1150 feet) (Flint, 1938b). The Touchet deposits in the Walla Walla Valley generally grade from coarser to finer both in an upvalley (eastward) direction and from lower to higher elevation (Bretz, 1928c, 1929; Newcomb, 1965). The principal constituents of the Touchet beds are buff to cream-colored clay, slit and sand with gravel ranging in size from small pebbles to large boulders. The mineral composition of the fine-grained portion of the sediments was estimated by P. D. Krynine (in Flint, 1938b, p. 494) to be:

1. Fine-grained fresh rock material (rock flour?)	65-80%
2. Yellow colloidal aggregates and iron-stained mineral grains (reworked Palouse loess?)	20-35%
3. Volcanic ash	1-5%
4. Diatoms	Trace

Much of the coarser sand and the gravel are derived from the Columbia Plateau basalts, whereas granitic and metamorphic erratics may have been ice-rafted from bedrock sources to the north and east (Bretz, 1928c).

The bedding of the Touchet sediments varies from massive to well-stratified. East of Reese (between Touchet and Wallula Gap on Fig. 9.1), the strata are graded and vary from 0.1 to 2 m thick, individual strata being thinner toward the tops of exposures. Individual rhythmites or turbidites (Baker, 1973a) are separated by disconformities and grade from a basal coarse sand upward to silt and clay (Fig. 9.2). Just above some disconformities are erratics (Fig. 9.3) as well as blocks of sediment (Fig. 9.4) which may have been frozen during transport.

The Touchet beds are slack-water sediments associated with the Missoula Flood(s). At least two major floods and perhaps several minor floods are recorded in eastern Washington (Brown, 1973; Baker and Patton, 1976; Baker, Ch. 2, this volume). Mullineaux and others (1977), working with Mount St. Helens tephra, demonstrated that the last major flood occurred about 13,000 years ago. In the Walla Walla Valley above the Touchet beds, there are patches of Mazama ash (Brown, 1971) and wind-blown silt ("post-scabland eolian deposits" of Flint, 1938a).

The origin of the rhythmites within the Touchet beds is problematic. Surely each of the more than 50 graded beds (Fig. 9.5) exposed near Lowden (between Touchet and Walla Walla on Fig. 9.1) cannot represent separate floods. Bretz (1929, p. 539) asked: "Is not some kind

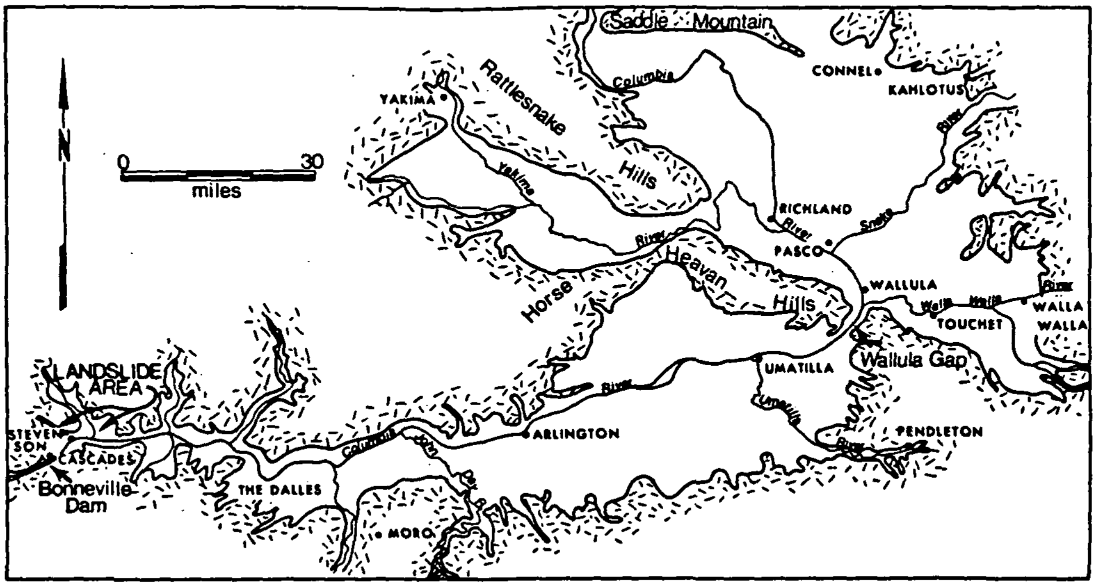

Figure 9.1. Approximate areal extent of Touchet beds and erratics. Areas of modern landslides in the Columbia Gorge are indicated. From Flint (1938b). Touchet-like valley deposits also extend up the Snake River and its tributaries to above Lewiston, Idaho. See Bretz (1929) and Flint (1938b, p. 507-510.)

Figure 9.2. Graded beds separated by disconformities (Burlingame Canyon, 4 km south of Lowden). Each rhythmite here grades from stratified sand at the base to massive silt and clay at the top. Note the thin sinuous clastic dike.

Figure 9.3. Granite erratic within the Touchet beds Burlingame Canyon). Stratified sand is at the base of the picture with massive silt and clay at the top.

of rhythmic pulsation in water supply and sediment supply recorded in the irregular and undulatory strata?" Lupher (1944, p. 1441) wrote, "The common alternation of stratified sand and massive silt layers indicates a periodic advance and retreat of the lake shore." The rhythmites probably record separate pulses of a flood or repeated surges or waves of water up the Walla

Figure 9.4. Sand block above silt block at same horizon as the granitic erratic in Fig. 9.3. The blocks may have beeen frozen during transport.

Walla Valley. Possible mechanisms to explain the graded beds include the following:

1. Variation in discharge at the ice dam as glacial Lake Missoula drained. This could be due to (a) gradual break-up of the ice dam, releasing water at a variable discharge, or (b) periodic floating of a diminishing ice dam by the glacial lake which drains beneath the ice (floating an ice dam is mentioned in Baker, 1973a).
2. Pulsations in the floodwater reaching Lake Lewis, due to (a) different lengths and sizes of the various scabland channels allowing portions of the flood to arrive at the lake at different times, (b) landslides, ice blocks, or gravel bars interrupting the waterflow in some of the scabland channels, or (c) surges of water from sudden channel deepening or the break-through of a cataract upstream from a scabland tributary (Baker, 1973a, p. 47).
3. Pulsations of floodwater between the major part of Lake Lewis in the Pasco Basin and the eastern arm of the lake extending up the Walla Walla Valley. At Reese (between Touchet and Wallula Gap, Fig. 9.1), there is a trench eroded into a basalt barrier (Bretz, 1929). The pulsations could reflect (a) seasonal precipitation which washed silt and sand into the lingering pond in the Walla Walla Valley (Bretz, 1929, p. 534-535), while the outlet of the pond was blocked by silt at Reese, or (b) ice jams, gravel bars, or possibly landslides which may have occurred near Reese could have varied the exchange of water through the pass.
4. Variations in flow rate at the dam sites (Wallula Gap and/or the Columbia River Gorge) of Lake Lewis. Such fluctuations could be envisioned as a result of (a) the break-up or floating of a Cascade glacier forming a dam at the Dalles (Russell, 1893; Flint, 1938b), (b) ice jams in the Columbia River Gorge (Allison, 1933, p. 719-721), or (c) landslides caused by flood erosion. Holocene slides in the Columbia Gorge are well known (Allison, 1933; Flint, 1938b), the evidence of the ones which temporarily dammed Lake Lewis could have been removed. In addition to ice jams, glaciers and landslides, damming of Lake Lewis could also have been controlled by lava flows (Allison, 1933), warping or faulting (Flint, 1938b) and hydraulic damming (Bretz, 1925; 1969; Baker, 1973a).

Another problematic feature of the Touchet beds are the clastic dikes (Figs. 9.2 and 9.5) which have been discussed by Jenkins (1925), Flint (1938b), Lupher (1944), and Newcomb (1962). The clastic dikes form a polygonal network according to Newcomb (1962) and have variable dips. The dikes vary in thickness from 1 mm to a few meters. Most of the large dikes tend toward the vertical (but are not planar). The dike sediment is as variable in grain size as the Touchet beds. In many cases, the dikes consist of vertical laminae of alternating silt and sand (Fig. 9.6). The clastic dikes may be related to the warping, folding and faulting of the Touchet beds. In many cases, the dikes consist of vertical laminae of alternating silt and sand (Fig. 9.6). The clastic dike may be related to the warping, folding and faulting of the Touchet beds (Flint, 1938b).

Below follows a summary of the many proposed origins for the clastic dikes.

1. The fissures which the dikes fill were produced by earthquakes (Jenkins, 1925; Jones and Deacon, 1966). Historic earth-

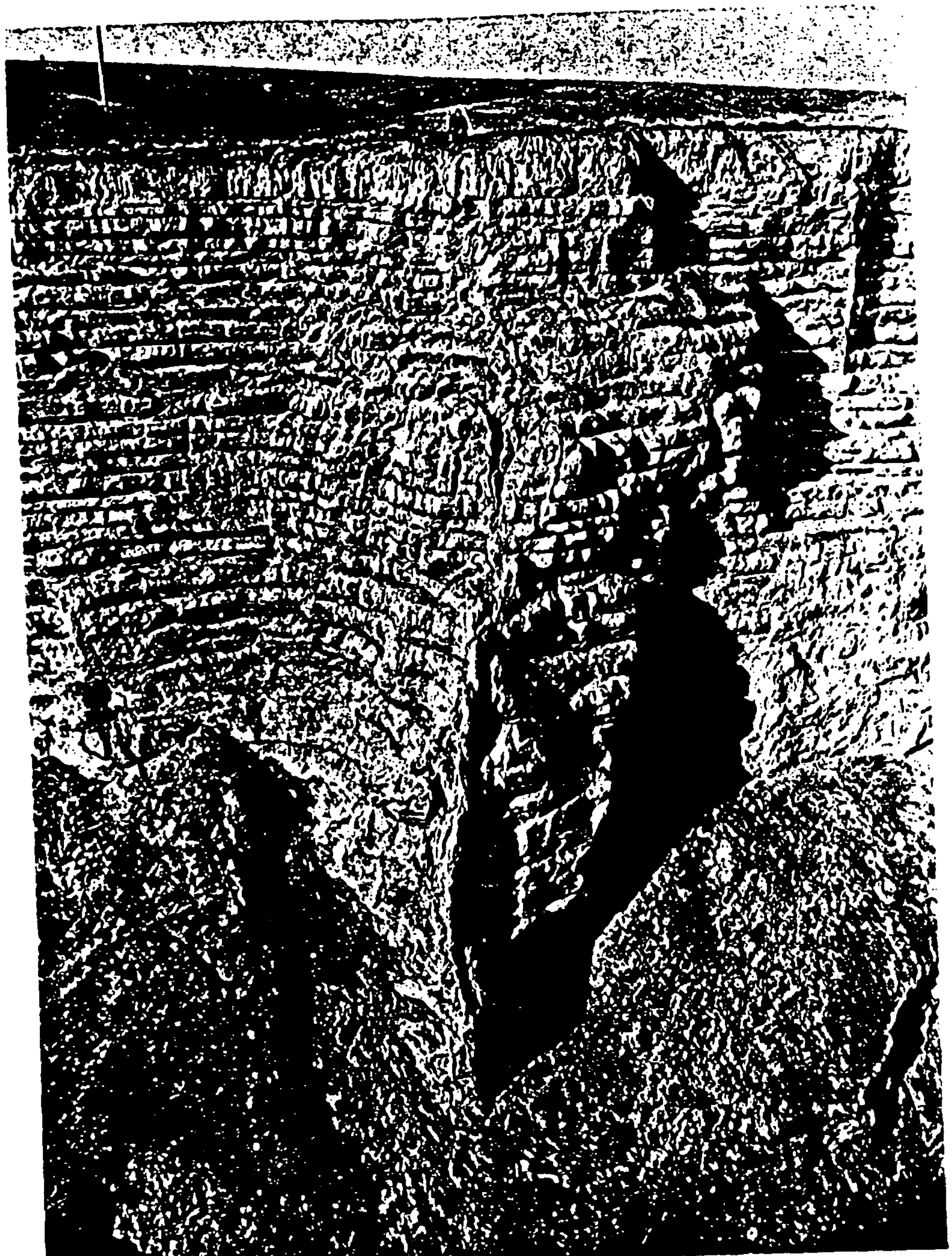

Figure 9.5. Rhythmites and clastic dikes in Burlingame Canyon. The clastic dike in the center of the photograph thins upward, whereas the dike in the upper right thins downward.

quakes in the Walla Walla Valley have not been uncommon (Brown, 1937; Jones and Deacon, 1966). The 1964 Alaskan earthquake produced clastic dikes with a polygonal pattern in saturated alluvium (Jones and Deacon, 1966).

2. The fissure development was caused by "(a) uneven settling and cracking through melting of buried ice, (b) gravity sliding and faulting on inclined zones of subsurface melting, (c) formation of cavities where ice blocks and layers melted" (Lupher, 1944, p. 1432).
3. A process of fissure development recognized in the Lewiston Basin was "erosion by underground streams" (Lupher, 1944, p. 1432).
4. Another dike-producing process noted in the Lewiston Basin, as well as eight kilometers northwest of Touchet, was "faulting and fissuring by landslides in the Columbia River basalts" (Lupher, 1944, p. 1432).
5. The dikes "represent fillings of permafrost-related crevices" (Alwin and Scott, 1970, p. 58).
6. During dessication of the Touchet beds, the contraction resulted in fractures.
7. The dikes resulted from repeated injections of groundwater, each hydraulic injection caused by a large lowering of Lake Lewis (Newcomb, 1962; Jones and Deacon, 1966).
8. The rapid deposition of a turbidite made the underlying sediments unstable. Slumping was accompanied by extension fractures (Baker, 1973a).

Jenkins (1925) preferred downward filling when the cracks opened beneath the mud of a lake bottom. Lupher (1944) also believed that the fissures filled from above, mostly by streams, lake currents, and waves. Baker (1973a) implies that the extension fractures were filled from above by later turbidity flows. Alwin and Scott (1970, p. 58) stated that "primary structures indicate a downward filling of the dikes by sand and silt." Newcomb (1962) believes in upward injections. In the Walla Walla Valley, individual dikes thin upward and/or downward (Fig. 9.5), suggesting that a combination of processes may be responsible for the clastic dikes.

Figure 9.6. Portion of clastic dike (exposed in Burlingame Canyon) exhibiting vertical laminae of alternating silt and sand. Dike sediments are of the same general composition as the surrounding Touchet beds.

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