

Chapter 6

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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 sub-aerially 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 scab-

land 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 sub-fluvially 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,

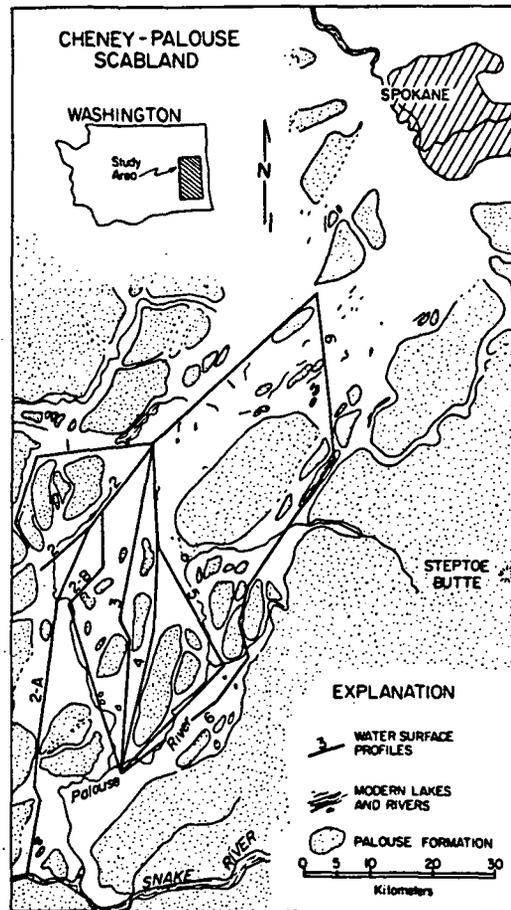


Figure 6.1. Highly generalized sketch map of the Cheney-Palouse scabland tract showing the location of various water-surface profiles that have been determined for the last major Pleistocene flood. Many of the small residual hills of Palouse Formation ("loess islands") have been omitted for simplicity.

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

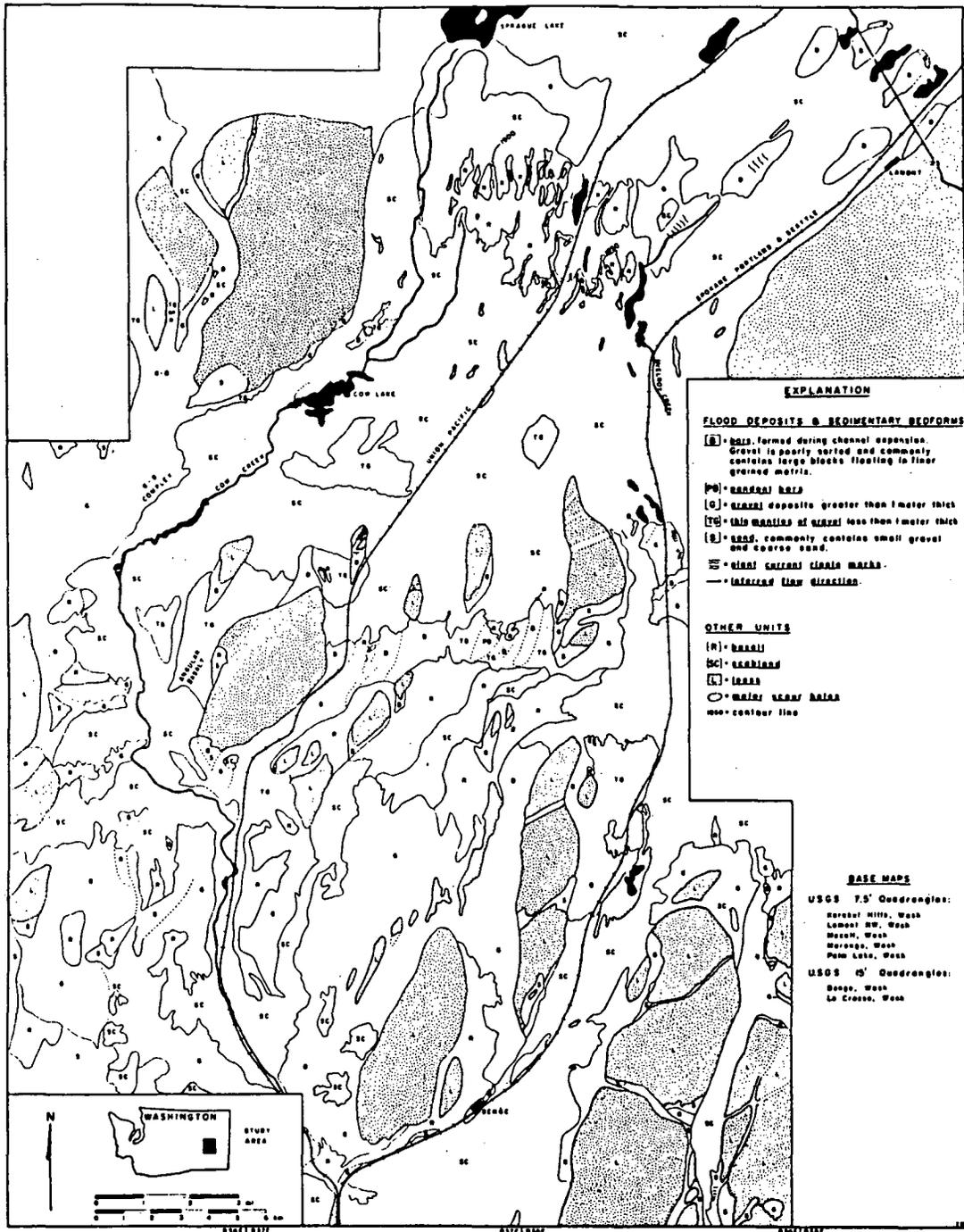


Figure 6.2. Geomorphic map of the central portion of the Cheney-Palouse scabland tract between Sprague Lake (north) and Benge (south). The map illustrates typical relationships between the various channel elements.

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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, abrasion, 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 lo-

calizing 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 pre-flood 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 bed-rock 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

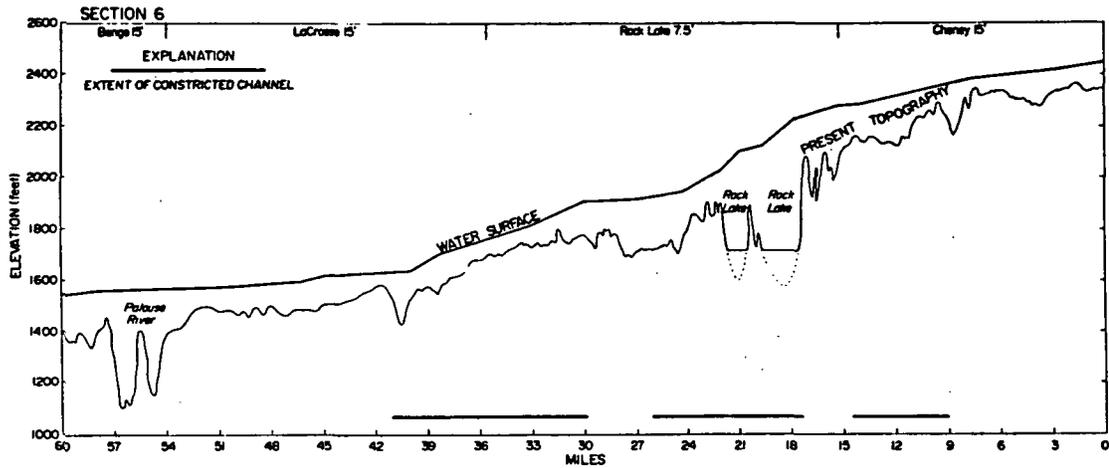


Figure 6.3. Profile 6 through the eastern part of the Cheney-Palouse tract (see Fig. 6.1 for location). Note the steepening of the water-surface gradient through constricted reaches.

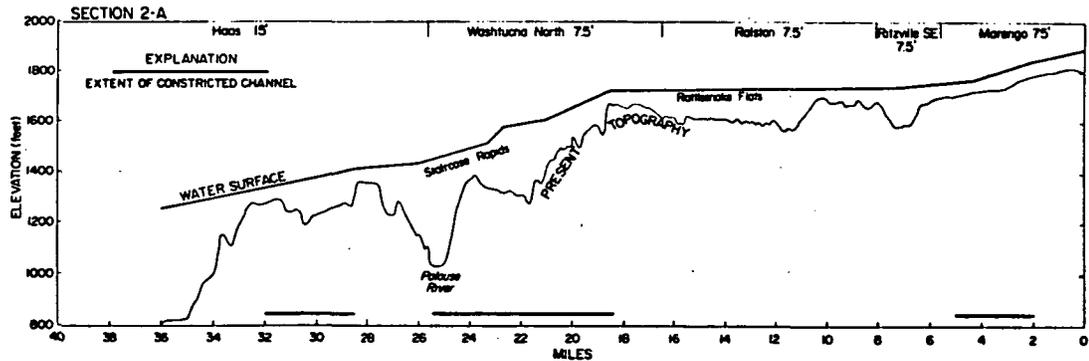


Figure 6.4. Profile 2-A through the western part of the Cheney-Palouse tract (see Fig. 6.1 for location). Note the pronounced ponding of water at Rattlesnake Flats and the steepening through Staircase Rapids.

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

Subfluviably 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

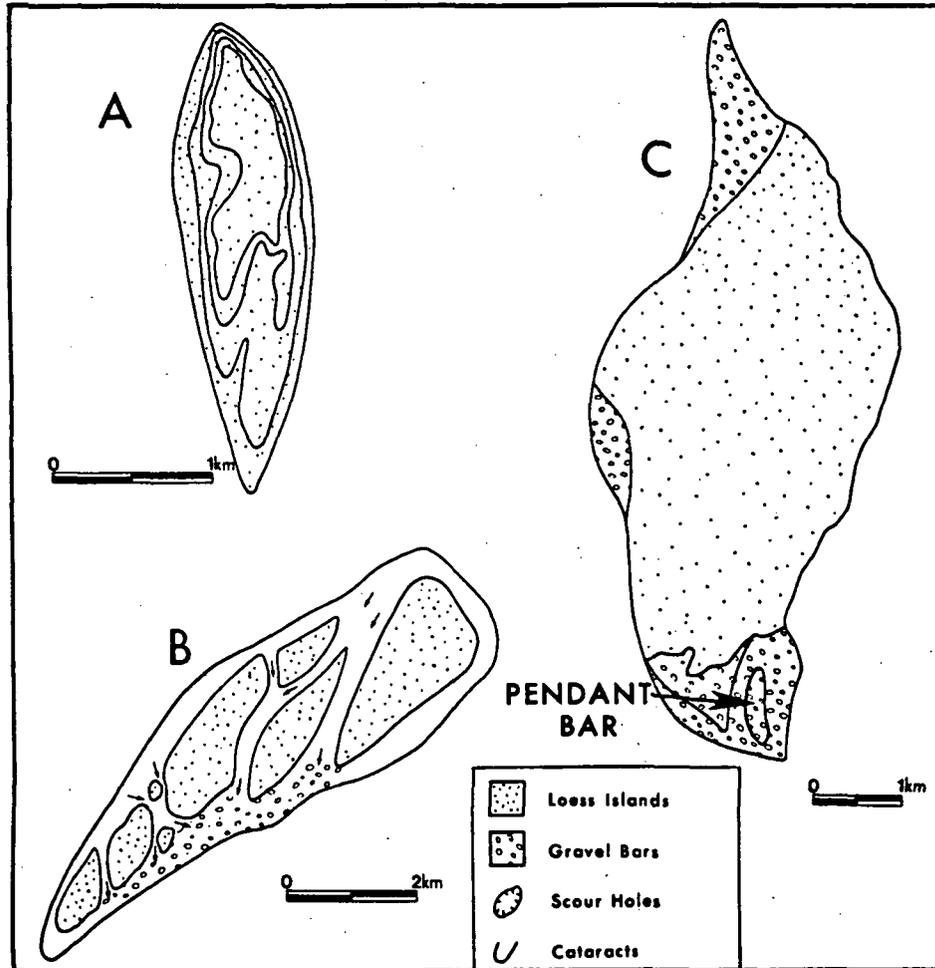


Figure 6.5. Loess island forms in the Cheney-Palouse scabland tract illustrating the three major morphologic types. A. Subfluviably eroded loess island located in the Karakul Hills (Secs. 2, 11, and 14; T. 19N., R. 36E.). B. Partially submerged loess island near Amber Lake

(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 interpreta-

tion of Bretz and others (1956). Many of Flint's "terraces" were gravel bars deposited in the reentrants 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 chan-

nel. 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 sedi-

ment 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

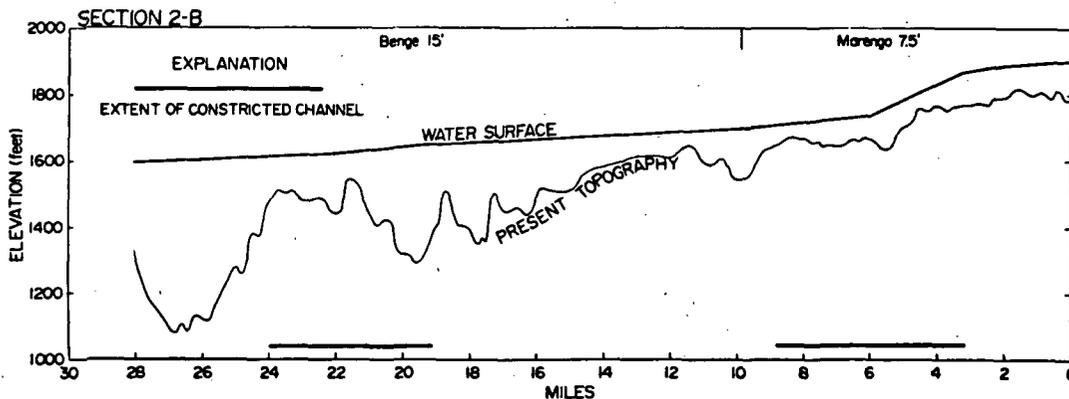


Figure 6.6. Profile 2-B along Cow Creek, a prominent inner channel of the Cheney-Palouse tract (see Fig. 6.1 for location).

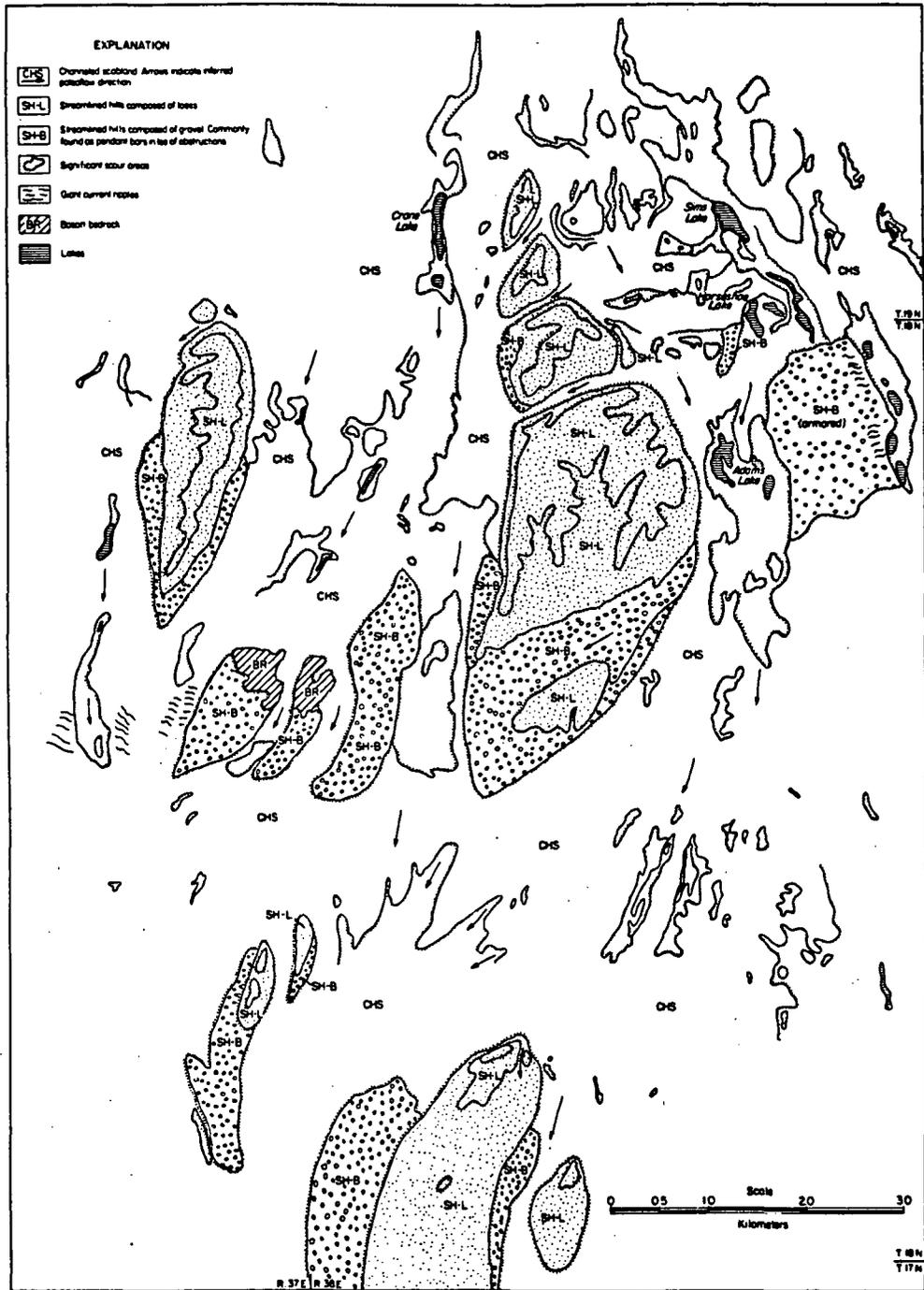


Figure 6.7. Geomorphic map of the Macall area in the central Cheney-Palouse tract.



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

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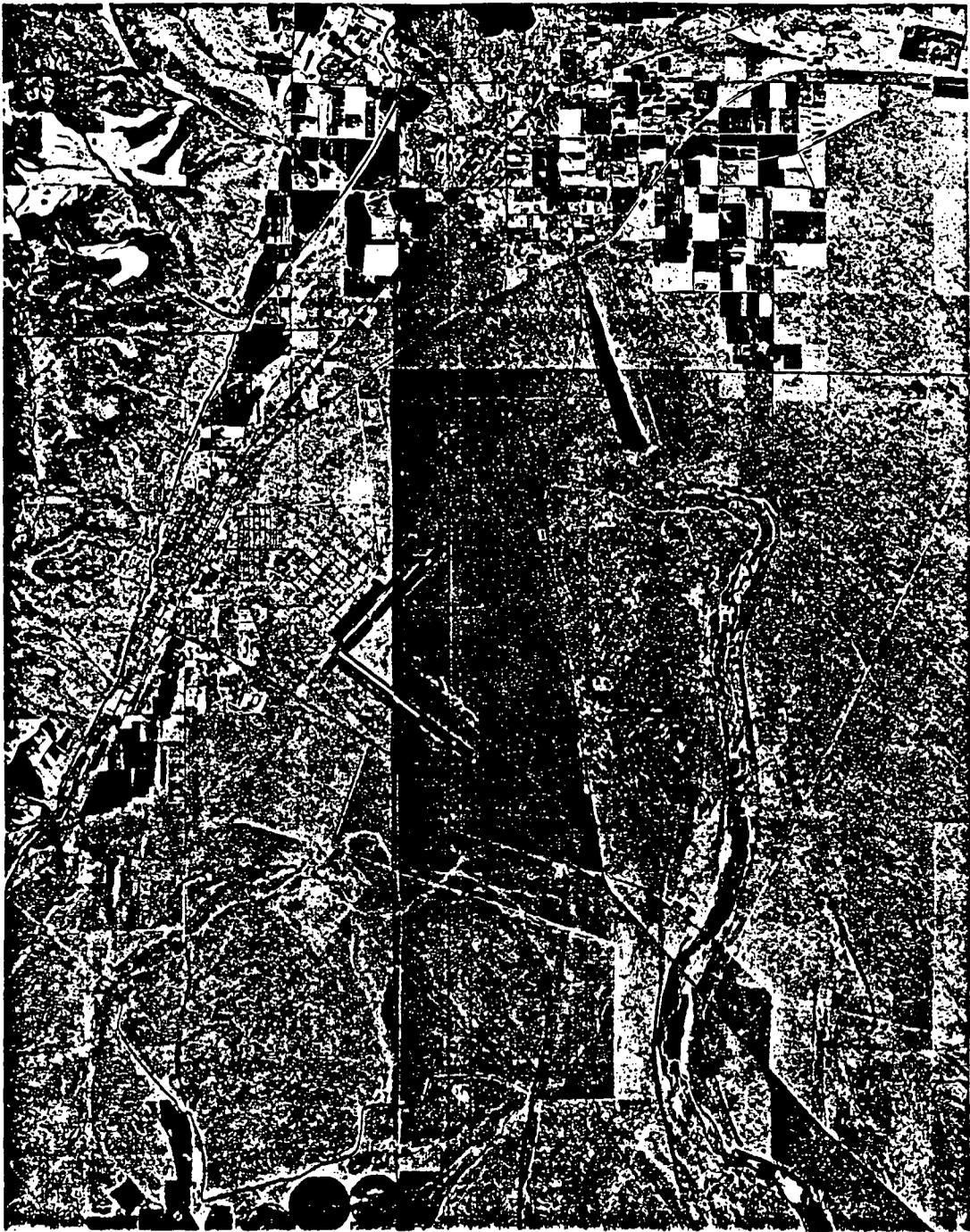


PLATE 2. Upper Ephrata fan. North is up.



PLATE 3. Lower Grand Coulee. North is up.

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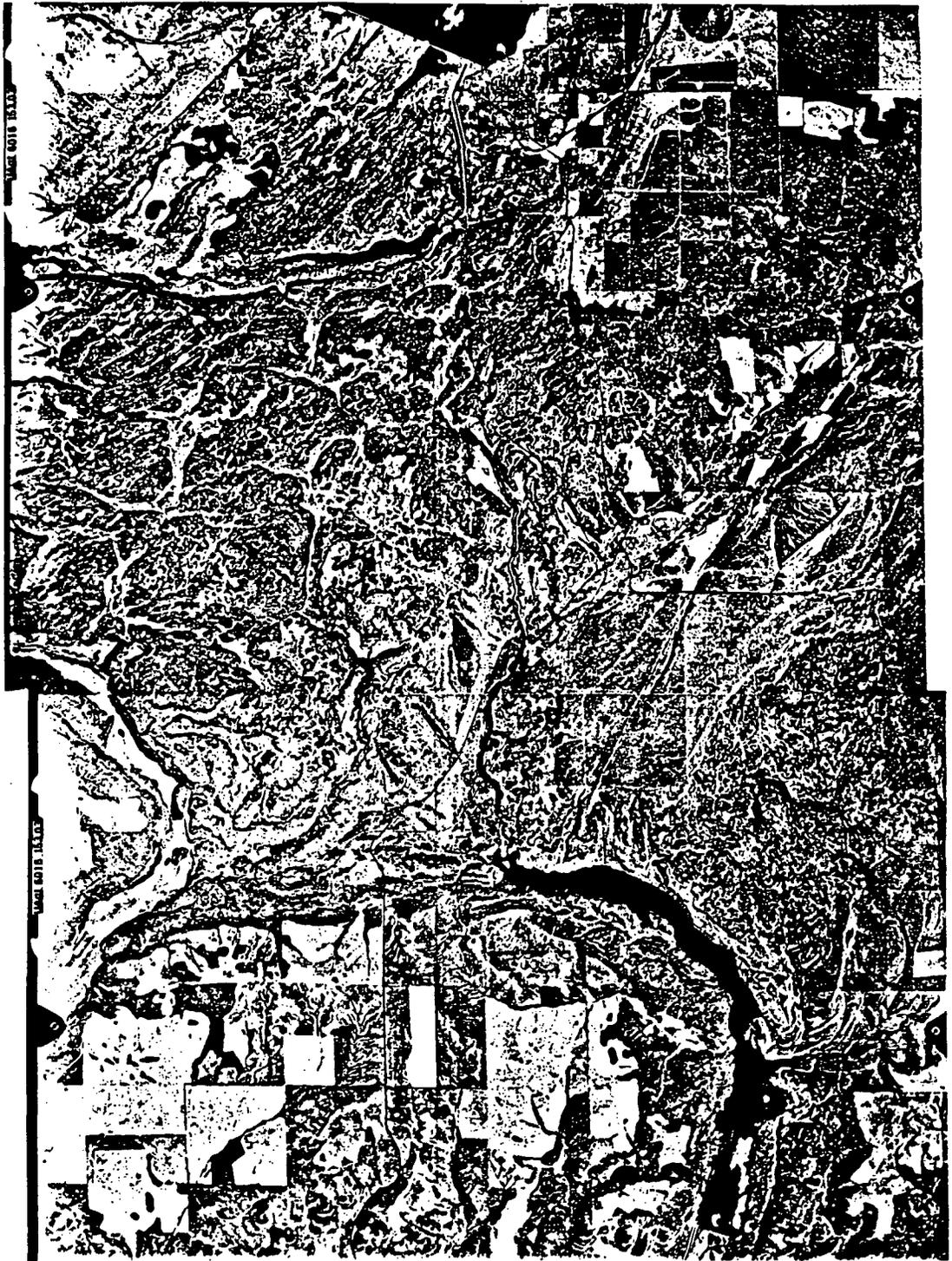


PLATE 4. Hayline Basin with Pinto Ridge. North is up.

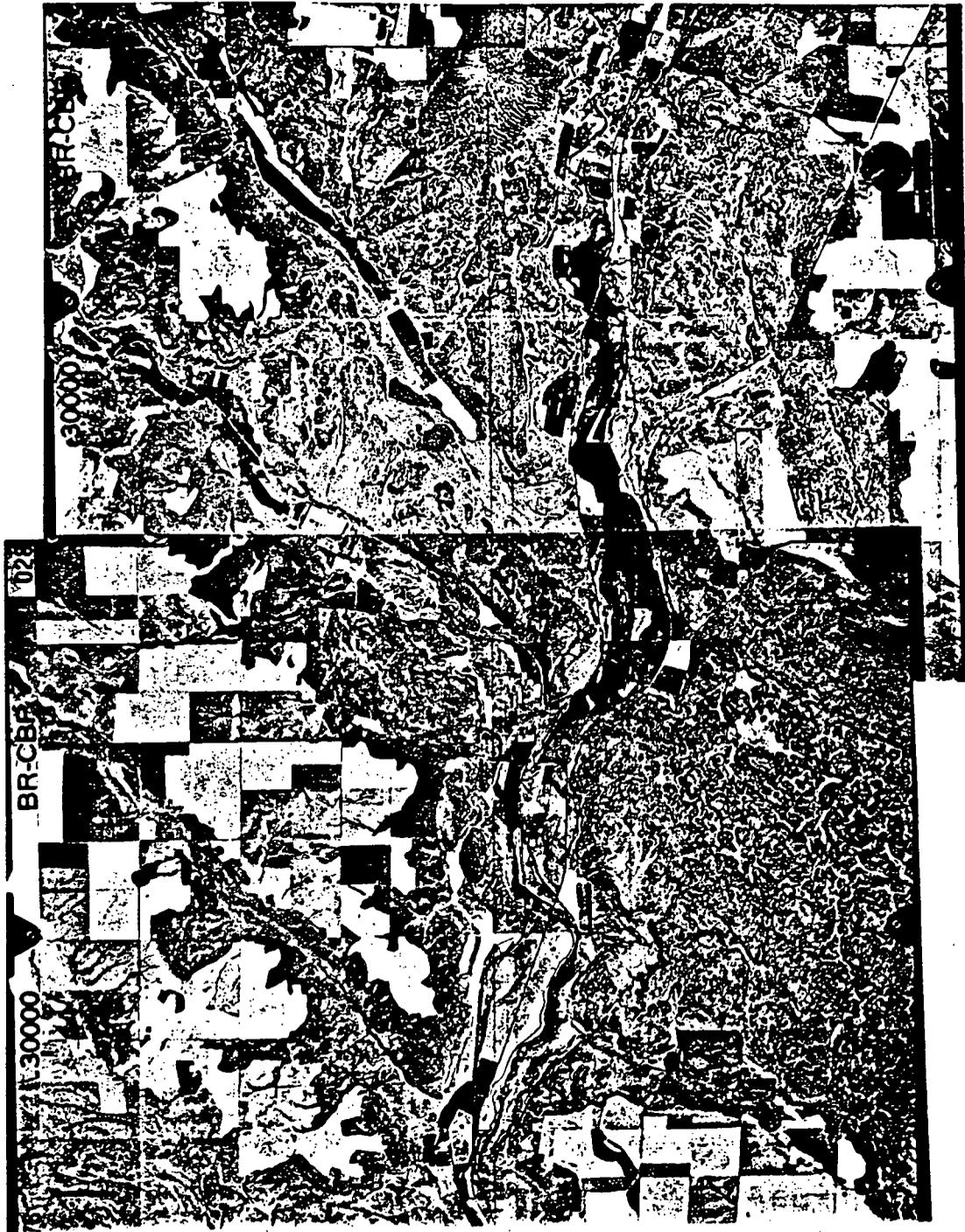


PLATE 5. Upper Crab Creek and Wilson Creek. North is to the left.

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



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

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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 slightly 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

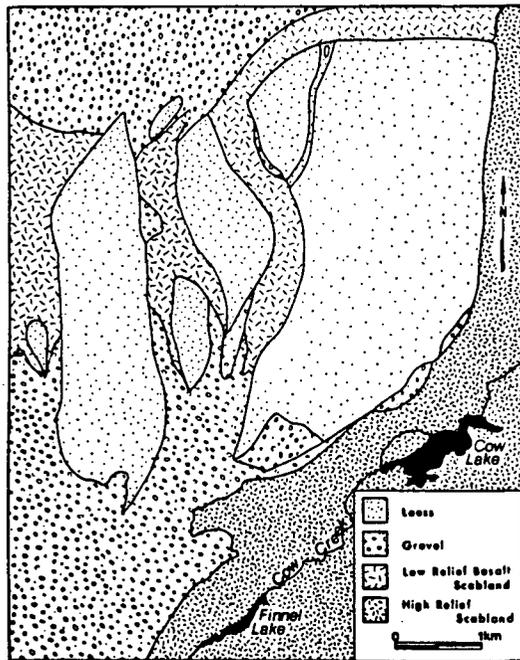


Figure 6.8. Map of the Karakul Hills loess islands.

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.

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-

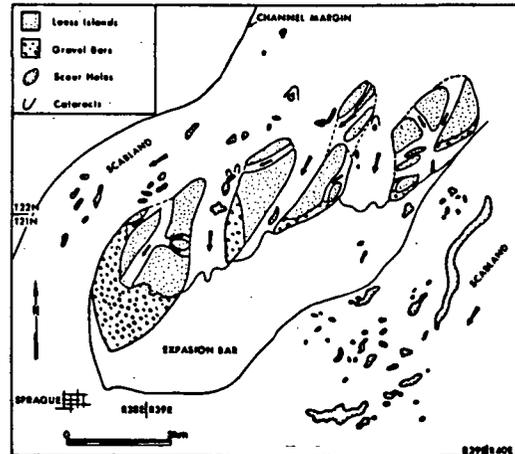


Figure 6.9. Map of the Sprague loess island assemblage.

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.