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Relation Between Geomorphic Stability and the Density of Large Shrubs on the Flood Plain of the Clark Fork of the Columbia River in the Deer Lodge Valley, Montana

Water-Resources Investigations Report 02-4070

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CONVERSION FACTORS AND ABBREVIATED UNITS

Multiply	By	To obtain
meter (m)	3.281	foot
kilometer (km)	0.6215	mile
square meter (m^2)	10.76	square foot
square kilometer (km^2)	0.3863	square mile
cubic meter per second (m^3/s)	35.31	cubic foot per second

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Abstract

A process-based procedure for investigating overbank flows during large floods in meandering fluvial systems is developed in this report in order to relate the stability of the Clark Fork of the Columbia River in the Deer Lodge Valley, Montana, to the density of large shrubs on its flood plain. This procedure permits the physical characteristics of large flood-plain shrubs to be used in a model for calculating shear stresses on flood plain surfaces. The flood-plain shrubs can vary continuously in density, from complete canopy coverage to none, and the shrubs can be grouped or uniformly dispersed. To investigate flood-plain erosion, the calculated boundary shear stress fields are compared to the threshold shear stress for erosion of the flood-plain surface. This threshold shear stress is composed of two parts; one for penetration of the high boundary shear stress through the surface organic cover, and one for erosion and transport of the flood-plain sediment. If wetland herbaceous plants cover the surface of the flood plain beneath the shrubs, then the threshold shear stress for penetration of the herbaceous vegetation can be very high. In contrast, if grasses with discontinuous sod cover the surface beneath the shrubs, the vegetative component of the threshold shear stress for erosion of the flood-plain surface is small and the flood plain will erode when the critical shear stress for sediment motion is exceeded by only a small amount. Once the flood-plain sediment is put in motion, the time scale for flood-plain erosion can be calculated using a standard skin-friction-based sediment transport method.

The density of large riparian shrubs on the flood plain of the Clark Fork in the Deer Lodge Valley has decreased substantially since the flood of record in 1908, according to a comparison of calculations for the 1908 condition and recent measurements. The nearly 400-year-recurrence-interval 1908 flood distributed an average of about 0.3 m of contaminated mine tailings on the flood plain of this meandering river upstream of the town of Garrison. The deposition of such a large amount of metals-rich, silt-sized material requires very low velocities and sub-threshold boundary shear stresses, which, in turn, indicates that there was a dense cover of large riparian shrubs in 1908. These tailings, however, subsequently killed or physiologically stressed much of the vegetation in the meander belt. At present, the large shrub flora on the Clark Fork flood plain in the Deer Lodge Valley is so sparse that (1) significant transport of the contaminated material into the river can occur during flows with very low boundary shear stresses, flows that can occur many times per century, and (2) this flood plain is capable at present of undergoing a catastrophic geomorphic transformation from a single-threaded, meandering fluvial system to a multi-threaded, braided system during a single flood. Such floods have a high probability of occurring several times per century.

INTRODUCTION

The Clark Fork of the Columbia River begins near Warm Springs, Montana, at the confluence of Silver Bow and Warm Springs Creeks (fig. 1). It then flows north through the broad, open Deer Lodge Valley to Garrison, where it is joined by the Little Blackfoot River. Downstream of Garrison, the river valley narrows and heads in a west-northwesterly direction

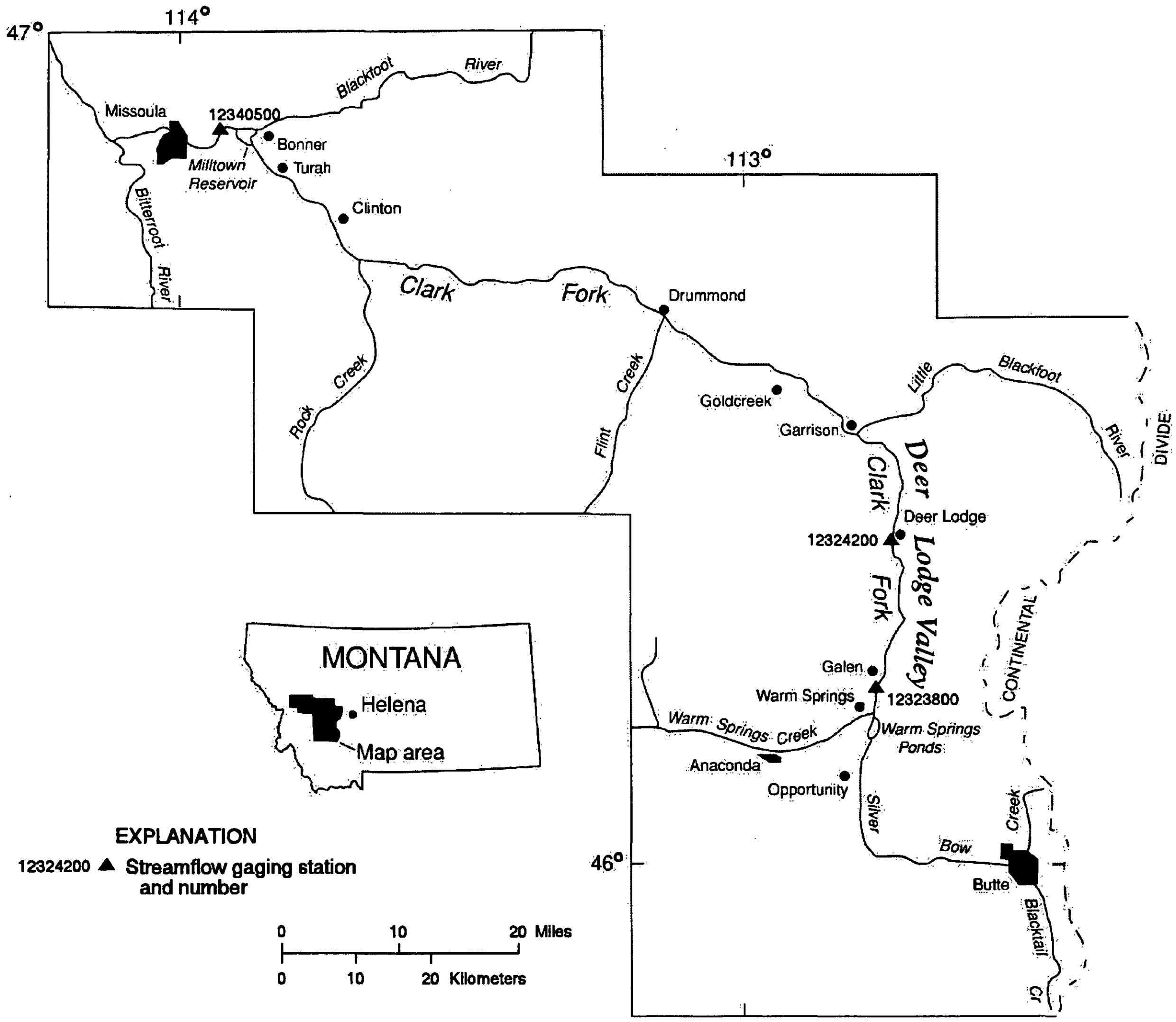

Figure 1. Location of the upper Clark Fork of the Columbia River, its major tributaries, its principal towns, and USGS gaging stations (modified from Smith and others, 1998).

toward Missoula. In the Deer Lodge Valley, the Clark Fork is a single-threaded, meandering river with an average sinuosity in the natural reaches of approximately 1.8 (Griffin and Smith, 2002); whereas downstream of Garrison, the river varies between single and multiple threaded and is much less sinuous (Smith and others, 1998). Between its origin and the Little Blackfoot River, only small tributaries enter the Clark Fork. Most of those flowing from the east are ephemeral and carry large volumes of sediment during short, high discharge floods. In contrast, those flowing from the west originally were mostly perennial but probably never carried much sediment. The west-side streams are now largely diverted for irrigation during the growing season.

The Clark Fork, upstream of Garrison, is typical of many small headwater streams in the semi-arid western United States in that it was bordered by lush, beaver-maintained willow carrs prior to the arrival of fur trappers of European ancestry (Smith and others, 1998). In 1831, a trapper named Warren Ferris visited the Clark Fork in the Deer Lodge Valley and described the flood plain as: "...decorated with groves and thickets of aspen, birch, and willow and occasional clusters of current and gooseberry bushes. The bottoms are rich and verdant and are resorted to by great numbers of deer and elk..." (Horstman, 1984). He continues, describing the river as "clear, deep, rapid and not fordable at high water", which means that the shrub flora on the banks had to be so dense that it was able to force the river to have an abnormally low width-to-depth ratio. Only at a few locations, primarily in the upper four miles, is this the case at present.

Subsequently, Father Pierre Jean De Smet, a Jesuit missionary and an early explorer of the American West, observed the Clark Fork and its tributaries in the Deer Lodge Valley in 1841. He wrote: "In no part of the world is the water more limpid or pure, for whatever may be the depth of the rivers, the bottom is seen as if there were nothing to intercept the view" (Horstman, 1984). This comment means that at the time of the observation, the river water was essentially devoid of suspended sediment, which, in turn, requires the riverbed in 1841 to have been gravel and to have contained an insignificant amount of silt and sand. Currently, clean gravel layers of this type crop out in cut banks in many places along the river. However, the flood plain of the Clark Fork in the Deer Lodge Valley is predominantly silty sand. The geomorphic implication of Father De Smet's observation, therefore, is that the river was narrow, deep (in agreement with the comments of Warren Ferris), and had banks so well vegetated with shrubs and herbaceous vegetation that the deep spring-snowmelt flows and occasional floods produced very low bank erosion rates and contributed little fine sediment to the river. In addition, there had to be a sufficient number of beaver dams impounding the river to prevent downstream transport of the silt and sand entering the Clark Fork from floods on the ephemeral east-side tributaries.

In sharp contrast to the fluvial system observed by Warren Ferris and Father De Smet, the channel of the Clark Fork in the Deer Lodge Valley today is broad, shallow, silty, and bordered in most places only by dispersed groups of mature willow shrubs and water birch. The average shrub density in the meander belt in 1997 provided less than 30% canopy coverage of the flood plain upstream of Deer Lodge (Griffin and Smith, 2002). Owing to a thick layer of mine tailings on the flood-plain surface, the Clark Fork in the Deer Lodge Valley is now part of a U.S. Environmental Protection Agency (EPA) Superfund site. The mine tailings have substantially reduced the vegetation, which has led to excessive bank erosion, high suspended-sediment concentrations, and high contaminant transport rates (Griffin and Smith 2001b; Smith and others, 1998). Vegetation typically protects the flood plains of meandering rivers from erosion. Therefore, the reduction in density of large shrubs and understory plants may have put the flood plain of the Clark Fork in danger of rapid surface erosion and extensive transport of contaminants into the river during floods. This issue needs to be evaluated and is examined quantitatively in this report. In addition, during floods with deep overbank flows, single threaded, meandering fluvial systems that lose their vegetative protection can become susceptible to catastrophic geomorphic transformation to broad, multi-threaded systems (Smith, 2001)

During the late 1800's there were numerous large floods on the Clark Fork above Missoula (Wheeler, 1974). These were documented primarily by newspaper articles, so their discharges could only be ranked. Nevertheless, this anecdotal information indicates that the flood of record in 1908 was the largest one of the sequence (Wheeler, 1974). During the 1908 flood, mine tailings from along Silver Bow Creek were transported downstream and deposited on the flood plain of the Clark Fork in the Deer Lodge Valley (Nimick and Moore, 1994). Further downstream (beyond Garrison), the majority of the tailings appear to have been mixed into the gravel bed of the predominantly multi-threaded river rather than being deposited on the flood plain (Smith and others, 1998). The newly constructed (1907) Milltown Reservoir just below the confluence of the Clark Fork and Blackfoot River and just upstream of Missoula was completely filled with tailings by the 1908 flood.

In the Deer Lodge Valley, tailings clearly mark the stage of the flood and the paths that it took. Surprisingly, for such a large flood, the river morphology hardly changed. The meandering channel remained intact with only a few cutoffs and short avulsions. This implies that the shear stresses at the bases of the cut banks and on the flood-plain surfaces had to have remained exceedingly small during the 1908 flood. Moreover, the thick deposits of tailings on the flood plain near the channel require the overbank-flow velocities to have been very low. Calculations using Rouse numbers based on settling velocities for medium and coarse silts suggest that the flow velocities at the depositional sites could not have exceeded 0.25 m/s and that they probably were much less than that.

Given the geomorphic characteristics of the fluvial system in 1908 and the reported characteristics of the flood-plain vegetation in the 1800's, the question arises as to whether the dense shrub flora observed in the 1800's could slow the deep overbank flow enough not only to prevent flood plain erosion, but also to promote the deposition of 0.3 m of tailings during the 1908 flood. To prevent significant flood plain erosion would require drag on the shrubs to reduce the shear stress on the Clark Fork flood plain of a 1.3 m deep flow from more than ten Newtons per square meter to less than one. To promote deposition of an average of 0.3 m of tailings (silt) in less than five days, drag on the shrubs would have to reduce the vertically averaged velocity by at least a factor of 5 (from potentially more than 2 m/s to less than 0.25 m/s). Further, if the dense willow and water birch on the flood plain proved to be the source of flood plain stability in the pre-1908 fluvial system, then the question arises as to what magnitude of flood the Clark Fork upstream of Deer Lodge could withstand given its present impoverished flora (Griffin and Smith, 2002).

The ultimate goal of this report is to investigate the geomorphic stability of the Clark Fork fluvial system in the Deer Lodge Valley in a quantitative manner. To do this has required the development of new fluvial geomorphic analysis tools. In order to answer the questions posed in the previous paragraph, it was necessary to develop new and more accurate procedures to calculate the responses of flood plains of meandering rivers in the semi-arid Western United States, such as the Clark Fork in the Deer Lodge Valley, to deep overbank flows. The procedure needed to be able to deal with flows over barren soils, over grasses with thin sods, through areas of mixed grasses and shrubs, and through shrub carrs of varying densities. Computation of flow properties over a flood plain requires the simultaneous and coupled calculation of the flow properties for the associated in-channel flow, because the partition of discharge between the channel and the flood plain depends on the characteristics of both regimes. Discharge of the overbank flow depends on the nature of the woody vegetation on the flood plain. The greater the resistance to flow over a broad flood plain, the smaller the fraction of discharge that covers it. Once a comprehensive procedure has been developed to calculate flow in a river channel coupled to flow over a flood plain as a function of (1) riverbed topography, (2) flood plain topography, and (3) stem densities of flood-plain shrubs, it will be used to investigate both the response of the Clark Fork of the Columbia River in the Deer Lodge Valley to the 1908 flood and the predicted response of the present fluvial system to floods with multi-decadal recurrence intervals.

In order to develop an appropriate theoretical procedure to examine the past and present stabilities of the fluvial system of the Clark Fork of the Columbia River in the Deer Lodge Valley, a certain amount of background information is required. First, the geomorphic features of that flood plain need to be characterized. Then it is necessary to examine the mechanisms by which flood plains of this type erode and to determine the threshold shear stress for erosion. After this preliminary information is presented, the characteristics of the computational procedure that has been employed are described. Then the new computational approach is applied to a series of situations. These are: (1) the 1908 flood, (2) tailings covered flood-plain segments, (3) grass and small shrub covered flood-plain segments, (4) large shrub patches on grass covered flood-plain segments, (5) grass covered flood-plain segments with bands of shrubs on their upstream sides, (6) grass covered flood plains with bands of shrubs on their downstream sides, (6) flood plains covered approximately uniformly with canopy-touching large shrubs, and (7) flood plains covered approximately uniformly with dispersed large shrubs. Once these situations have each been examined, the overall stability of the Clark Fork in the Deer Lodge Valley is discussed.

BACKGROUND

Characteristics of the Flood Plain of the Clark Fork in the Deer Lodge Valley

As a single-threaded river meanders through a valley, it forms a zone that is in equilibrium with the processes resulting from the flood history of the river. This zone is called the meander belt, and it extends from the outer bends of the meanders on one side of the valley to the outer bends of the meanders on the other side of the valley (fig. 2). Meander belts of low-order streams usually are fairly narrow and their axes correspond with the valley axis. When a meander belt is narrow, the river often oscillates more or less symmetrically back and forth across the valley axis in the shape of a piece of ribbon candy. This geometry is called a sine-generated curve (Leopold and Langbein, 1966; Smith and McLean, 1984; Kean and Smith, 2001). In this situation the segments of flood plain in the meander belt are bounded on three sides by the river and on the fourth side by the edge of the meander belt. These well-defined flood-plain segments are defined here as *flood-plain tabs*, because they are geometrically similar to the tabs of a puzzle (fig. 2). A typical series of meanders along the Clark Fork is shown in figure 2. Also displayed is an inset of one meander bend showing the angles with which the river crosses the valley axis.

In the case of the Clark Fork, the local width of the meander belt at a bend usually is the width of its inscribed flood-plain tab plus the width of the river at the bend. According to Griffin and Smith (2002), upstream of Deer Lodge the average width of the meander belt is 138 ± 56 m, and the average width of the river is 20.0 ± 2.9 m. For this reach, the average meander wavelength is 508 ± 168 m and the average angle at which the river crosses the valley axis is 77.6° ± 21.3° (Griffin and Smith, 2002). The average river slope is 0.00194 and the average valley slope is 0.00353. The *meander wavelength* and the *crossing angle* are the parameters necessary for approximating a meander sequence with a sine-generated curve (Kean and Smith, 2001). In this situation, where the river oscillates back and forth across the valley axis, the tabs extend nearly perpendicularly across the valley axis (fig. 2), so deep overbank flows are directed across the flood-plain tabs in the down-valley direction. The average tab width upstream of Deer Lodge is 118 ± 56 m and the average maximum-down-valley distance across the tabs is 139 ± 76 m (Griffin and Smith, 2002). The overbank flows would be approximately uniform in depth in the cross-meander belt direction if the tabs were homogeneous in their vegetation cover. Upstream of Deer Lodge, however, the average canopy cover is only 30%, and the shrubs are quite non-uniformly distributed in the cross-tab direction. The actual distribution of flow properties during a large flood, including flow depth, is sensitive to the distribution of the various types of large woody plants on the flood-plain tabs. During a large flood, the meander belt is the approximate zone of deep-overbank flow. Deep overbank flows, therefore, are driven by the valley slope, and the flow direction on the flood plain within the meander belt is approximately down valley. Over the flood plain, the flow depth is approximately uniform across the meander belt.

Along the Clark Fork it is common to have a single row of large shrubs fringing the river (Griffin and Smith, 2001b; __2002). Where this fringe exists, the floodwaters decelerate onto the flood plain during overbank flows so that mostly mass and not momentum is transported onto the flood-plain tabs from the river channel. That is, the added resistance from the shrubs in the fringe causes the water-surface elevation

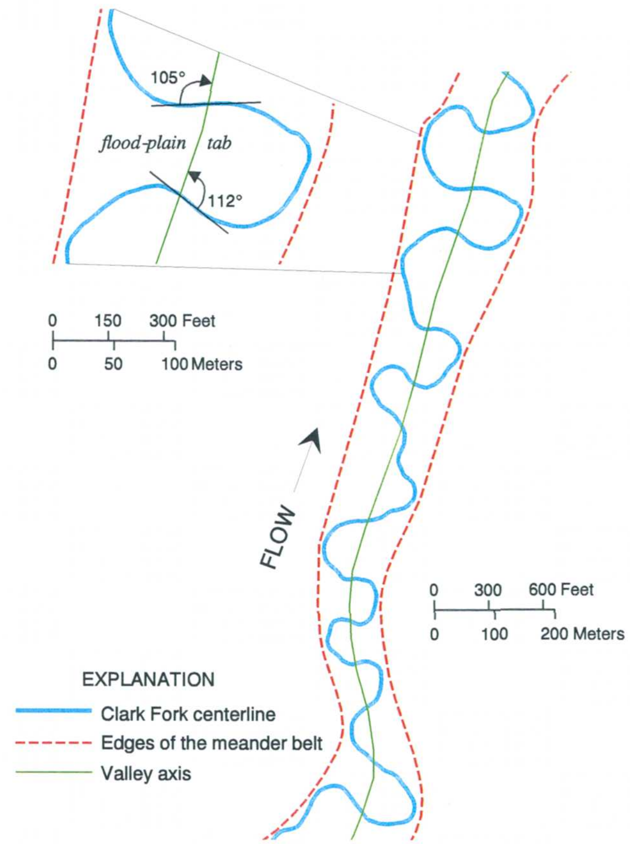

Figure 2. Sequence of meanders on the Clark Fork showing the meander belt and an inset of a typical flood-plain tab. The inset also shows the angles at which the selected bend crosses the valley axis.

to rise in the channel, then drop rapidly toward the flood plain as the flow onto the flood plain decelerates through the band of shrubs. Moreover, during large floods this situation is exacerbated by the entrapment of uprooted woody vegetation and other debris in the shrub fringe on the up-valley sides of the tabs, where the flow passes onto the flood plain. Although this enhanced resistance raises the stage in the channel, it is not of sufficient cross-tab extent to significantly alter the stage of the over-tab flow down valley of the shrub fringe. Inside the shrub fringes, many of the flood-plain tabs in the Deer Lodge Valley have a large fraction of their area covered by (1) barren tailings, (2) amended tailings with thin grass cover, (3) heavily grazed grasses with thin sods, and (4) non-irrigated grasses intermixed with sparse small shrubs (Griffin and Smith, 2002). None of these provide significant resistance either to overbank flow or to erosion of the tab surface. Although the flow resistance from these surface coverings is very small, the areas of the tabs that they encompass are large (Griffin and Smith, 2002). This means that the boundary shear stress on these tabs is controlled by (1) the depth of flow inside the upstream shrub fringe and (2) the valley slope until the flow reaches the shrub fringe on the down-valley sides of the tabs. On the down-valley sides of the tabs, flow acceleration becomes important.

The response of the flow through the downstream shrub fringe depends on the density of the shrubs. As mentioned previously, the shrubs on the edge of the Clark Fork in the Deer Lodge Valley usually are in a single row and rarely form rows greater than several shrubs in width. According to Griffin and Smith (2002), the shrubs in this location along the river are rarely closer than one canopy diameter, or about 4 meters center-to-center. Measurements of a small set of individual large shrubs suggest that the Bebb, Geyer, and yellow willows and water birch along the Clark Fork upstream of Deer Lodge can be modeled as if composed of a group of 36 stems supporting branches and leaves that form a 4-meter diameter canopy. The basal diameter of the group of 36 or so stems, which comprises a single shrub, is about 1/5 of canopy diameter, or about 0.8 m. The average basal diameter for shrubs on the Clark Fork flood plain is smaller than for this group of large shrubs (0.5 m rather than 0.8 m). Unless the stem clusters of an individual shrub are closer together than one stem-cluster diameter (tightly intermingled canopies) for many rows back from the river, the high velocity flow from across the tab will accelerate between the shrubs. The accelerating flow will then erode deep channels into the flood plain surface, countering the reduction in cross-sectional area of the flow caused by the blocking effect of the stem groups. This essentially negates the backwater effect that would arise from an extensive, dense shrub fringe. Erosion of channels beneath the accelerating flow is possible at the down-valley sides of the tabs because the velocity and boundary shear stress are both increasing along a flow path, producing a negative divergence in the boundary shear stress (as is required for the erosion of bed material). Deposition of the eroded sediment does not occur until it reaches the main channel and the sign of the divergence changes. The flow eroding the channels eventually undercuts the shrubs and removes them.

Mechanisms of Flood Plain Erosion

During large floods, removal of sediment from the flood plains of meandering rivers occurs in two primary ways. These are (1) lateral erosion of riverbanks by in-channel flow processes and (2) surface erosion of the flood plain by overbank flows. When the erosion rates for successive cut banks are the same, sediment eroded from their bases usually is deposited on the point bars immediately downstream. In this manner, the river moves laterally while maintaining a constant width and a constant cross-sectional area. In contrast, material added from other sources cannot be accommodated on the point bars without causing the river to narrow. Narrowing of the river in the vicinity of the point bar, however, causes a divergence of the boundary shear stress in the neighborhood of the downstream crossing. This, in turn, causes deposition on the riverbed of enough of the added sediment to restore the equilibrium downstream trend in the boundary shear stress field produced by the cross-sectional structure of the river.

As the river stage rises above the bankfull level, threads or sheets of overbank flow connect the main channel to the up-valley and down-valley sides of the flood-plain tabs. As the overbank flow deepens, the shear stresses on the flood plain beneath these threads increase until they exceed the critical shear stress for erosion of that surface. At that point, small channels begin to erode into the tab surfaces. As the channels erode, at first they curve around obstructions such as shrubs and clumps of sod. These features of the flood plain are removed later by undercutting after the channels have widened and deepened. Deepening of the flood-plain channels is a catastrophic process. The deeper the channels get, the higher the shear stresses on their bottoms, the higher the sediment transport rates, and the faster they erode. Like rills, as flood-plain channels deepen they also widen, leading to an ever-increasing export of sediment to the main river on the downstream sides of the tabs.

The large volume of sediment that is eroded through deepening and widening of channels on a flood-plain tab is deposited on the riverbed just downstream of that tab. Under bankfull and slightly overbank flow conditions, materials contributed to a river from a variety of sources usually are fairly well mixed by the time they are deposited. Smith and others (1998) used this principle to investigate the sources of contamination reaching the Clark Fork between Warm

Springs and Missoula. For deposition resulting from a rapidly eroding flood plain, however, this will not be the case. Incomplete mixing is almost certain because the volume of material being added to the river is so large that it cannot be transported very far. When such large volumes of material come from a single source, the composition of the material being deposited on the riverbed at the first available depositional site is likely to reflect the substantial contribution from that source. In particular, sand eroded from the channels on the flood-plain surface is likely to dominate the deposit on the channel bed just downstream of the tab from which it eroded. Even when the gravel of the riverbed is mobile at that site, it will first become mixed with sand from the flood plain as gravel stringers, but later it will be buried under the finer material from the tab.

As the flood plain channels deepen and the amounts of material exported from them increase, there is an ever-increasing transfer of material from the flood-plain tabs to the river channel. The decreased cross section of the river channel then forces a larger fraction of the channel flow over the next few tabs downstream. The result is runaway erosion of the flood-plain tabs and runaway filling of the original channel. Eventually, either the flood ends or the original river channel fills with sediment and is obliterated. If sediment fills the original channel, then the flowpaths on the flood plain tabs connect to become a set of new channels in a now elevated, multi-threaded fluvial system. Tabs with sparse, uniformly distributed shrubs are likely to be completely obliterated, whereas, segments of tabs with dense clusters of shrubs are likely to become islands between threads. Any sequences of tabs with sparse, uniformly distributed shrubs are likely to become braided sections. These braided sections would contain mobile bars composed primarily of the original flood-plain sediments.

As sands and silts are eroded from the flood plain they are put into suspension and sorted by their settling velocities. Consequently, materials with different settling velocities will end up at different locations in the evolving geomorphic system. The coarsest sediment will be deposited on the channel bottom and the finest sediment will move in the down-valley direction, following the overbank flow. Mixing of fluid and suspended sediment will occur each time the overbank flow interacts with the channel. Silts from the eroding flood-plain tabs will increase in concentration until the amount being eroded equals the amount being deposited in shrub bunches on the remaining flood plain or in the interstices of the sand grains added to the stream bed. Owing to the very small sizes of contaminated sediments such as those on the flood plain of the Clark Fork in the Deer Lodge Valley, these materials would first be deposited on any remaining parts of the flood plain that are densely populated with shrubs and understory herbaceous plants. If the shrub and understory flora of the flood plain were sparse, however, as on the flood-plain tabs of the Clark Fork upstream of Deer Lodge, the greatest mass of contaminated sediments would eventually be deposited on the bottom of the river. Once a fluvial system such as the Clark Fork in the Deer Lodge Valley changed from meandering to braided or partially braided, most of the contaminated sediment would be deposited in the upper part of the bed of the newly formed sandy channel bottom. These contaminated materials would be almost uniformly mixed into the dunes and bars of the newly evolved river bottom. In subsequent years, oxidation of the metal sulfides would occur when the river stage was low. Dissolved copper then would be released to the river as the now finer and more mobile bars moved during high stage flows.

Process-based Treatment of Bed and Bank Irregularities in Natural Flow Situations

Unlike flows in laboratory flumes, rivers have uneven beds, irregular banks, and non-uniform planiform geometries. In order to make experimental and theoretical examination of such flows tractable, various forms of spatial averaging are used. To simplify the effects of channel curvature, a curvilinear frame of reference is used. This procedure separates the convective accelerations associated with channel topography from the effects of channel curvature (Smith and McLean, 1984). The effects of channel topography then can be simplified by spatially averaging the flow over various length scales in the curvilinear frame of reference. If, for example, the flow through a series of nearly similar meander bends is averaged over one such bend (reach-averaged), then the result is a much simplified governing equation for the flow (downstream component of the fluid weight per unit volume balanced against the divergence of the Reynolds shear stress), but the expressions for friction on the channel bed and banks become very complicated.

The non-averaged fluid-mechanical formulation is governed by the full set of Reynolds equations for incompressible flow (Batchelor, 1967; Kundu, 1990). In general these equations are not closed and, thus, are not solvable, but in this case, the friction is fully specified by the grain roughness on the irregular boundary. Once the flow has been reach-averaged, however, this situation changes. By reach averaging, the boundary of the flow is made smooth, and the friction has to incorporate all of the form drag on all of the variations in geometry from the smoothed boundary. We define these irregularities as *topographic elements*. In the case under discussion, the topographic elements include the point bar in the meander bend, any dunes or scour-and-fill gravel structures on the channel bottom, and any irregularities of the stream banks relative to the smooth surfaces of the reach-averaged channel boundary. In a formulation using the Manning equation, friction on the sediment of the streambed and banks is enhanced by form drag on all of the topographic elements in the flow, and the sum of these effects is combined into one friction coefficient (often called the Manning coefficient or Manning's n). This procedure leads to great simplification of some fluvial hydraulics problems, but it leads to serious difficulties in others because it requires the friction coefficient to be determined empirically, and the value depends on the numbers and types of topographic elements in a particular reach. Moreover, the value of the boundary shear stress calculated using the Manning method is not related in a simple manner to the value needed to carry out erosion and sediment transport calculations.

If the flow across a flood plain tab were reach-averaged, then the topographic elements would be trees, shrubs, and irregularities of the flood-plain surface. In a formulation using the Manning equation, all of these effects would have to be combined into one coefficient, which would have to be determined empirically. The accurate determination of a Manning coefficient as a function of shrub density would be very difficult and, perhaps, impossible during a small flood. It would be even more difficult if the flood were large and the flood plain was eroding. Moreover, even if this coefficient were properly determined as a function of shrub density, it could not be partitioned into the components arising from trees, from shrubs, from flood-plain irregularities, and from friction on the actual flood-plain surface. An effective alternative would be to develop a means of predicting the components of flow resistance that combine to produce the Manning coefficient. A process-based procedure for separating the boundary shear stress into the necessary components is developed below.

In sediment transport calculations, only the component of boundary shear stress on the actual surface at a particular site is of importance. The forces on a sediment grain on the streambed depend on the boundary shear stress averaged over an area on the order of ten median grain diameters (defined here as *local skin friction*). If a flow is reach averaged, then the component of the total boundary shear stress related to local grain roughness often can be approximated by the spatial average of the shear stress on the actual (irregular) boundary. This is the spatial average of the local skin friction, and following the convention of Smith and McLean (1977) we refer to the averaged value as *skin friction*. It is skin friction on which erosion and sediment transport depend. In order to calculate erosion and sediment transport accurately when a flow has been reach averaged, skin friction must be separated from the total boundary shear stress. The need to partition the boundary shear stress for sediment transport calculations was recognized by Einstein (1950), but the procedure that he developed was not successful. To do the separation properly, form drag per unit area must be calculated for each of the classes of topographic elements. A procedure to do this for bedforms was devised by Smith (1977) and tested by Smith and McLean (1977) on a comprehensive set of measurements made over a field of large dunes in the Columbia River.

The fluid-mechanically-proper definition of skin friction is the boundary shear stress on the actual solid surface at a scale smaller than the topographic elements. The shear stress on a perfectly smooth sheet of glass is true skin friction. When applied to natural systems in a spatially averaged context, as done by Smith and McLean (1977), what is skin friction and what is form drag per unit area is a matter of scale. In this report, the shear stress on a smooth surface composed of sediment particles is called skin friction even though the skin friction in this case is due primarily to the sum of the form drags on each of the grains comprising the smooth array of sediment grains. In this sediment case, the actual skin friction is on the surfaces of the grains. The same problem arises when the concept is applied to a rusty steel plate. In this case, some of the boundary shear stress is still a consequence of form drag on surface irregularities. As mentioned above, a means of separating the shear stress on irregular sediment surfaces into the stress on the actual surface (skin friction) and that arising from spatially averaging (form drag per unit area) is essential for accurate sediment transport calculations. As one goal of this report is to calculate when flood-plain erosion will begin, the ability to make accurate skin friction calculations is essential.

DETERMINATION OF THE THRESHOLD OF EROSION ON FLOOD-PLAIN TABS ALONG THE CLARK FORK

Estimation of Threshold Shear Stress for Erosion of the Flood Plain of the Clark Fork

In order to investigate erosion of a flood-plain surface, the threshold boundary shear stress for erosion beneath threads of cross-tab flow must be known. Once erosion of a vegetated flood-plain surface begins beneath a thread of cross-tab flow, the erosion process is enhanced in the upstream and downstream directions. Consequently, the cross-flood-plain channels that are first to form follow paths of least flow resistance and maximum erodibility. Later, as the channel deepens and widens, the more resistant elements of the flood-plain surface are removed by undercutting. This means that the relevant threshold for erosion of a flood-plain surface is *somewhat less than the average value for the most erodible down-valley paths across the surface of that tab.* If a large population of randomly distributed erosion threshold measurements had been made for a tab of interest, then the threshold value could be suitably approximated by the value associated with one standard deviation of the population below the mean. Unfortunately, large populations (several hundred measurements) of randomly distributed erosion threshold values have never been collected on flood-plain tabs.

Owing to the difficulty of obtaining a sufficient number of direct field measurements of flood plain erodibility along the Clark Fork in the Deer Lodge Valley, an efficient approximate method that focuses on the salient characteristics of this flood plain was designed. For the most vulnerable half of the tabs on the Clark Fork in the Deer Lodge Valley, this surrogate method can be employed effectively. These tabs are ones on which mineral soil can be collected from the flood-plain surface in late May by reaching down and picking up large pinches of sediment from between blades or clumps of grass at random locations in four out of five attempts. Once the random samples have been collected and analyzed using a sand-size card (Compton, 1962; Birkeland, 1984), it is straightforward to find nearly straight (sinuosity less than 1.1), easily eroded, down-valley paths across the tabs and to determine the median size of the sediment along these paths. Moreover, the amounts of silt and clay and the presence of significant cohesion can be estimated with practice by rubbing the sample gently between the thumb and the index finger. These procedures were used by Smith on four randomly chosen, poorly vegetated tabs.

When the finger test is done only for soil characterization, it is carried out with a dry sample, but for critical shear stress estimates of flood-plain soils it must be carried out with both wet and dry samples. The wet finger test is reasonably accurate for demonstrating that cohesion is not an important factor, but it does not lead to a useful method for correcting the threshold shear stress for sediment motion when the estimated cohesion is significant. Except in areas where tailings crop out at the surface ("slickens"), the finger tests demonstrate that soil cohesion is minimal and is not likely to increase significantly the critical shear stress for erosion of flood-plain soils along the Clark Fork in the Deer Lodge Valley. Therefore, the threshold shear stress for significant erosion and transport of the flood-plain sediment along most of the Clark Fork in the Deer Lodge Valley can be calculated from the estimated median size of the sand fraction and its density. This is done using the method of Shields (Henderson, 1966; Graf, 1971; Yalin, 1972; Middleton and Southard, 1984; Wiberg and Smith, 1987). When the above-described methods are applied to typical tabs on the Clark Fork, threshold shear stresses ranging from 0.3 to 0.5 N/m^2 are found. This result was later confirmed by calculations, described below, based on field observations during the 1997 bankfull flow.

As mentioned in the previous paragraph, the finger test is not useful for quantitative inclusion of cohesion in threshold boundary shear stress calculations. Therefore, it was used to determine that the degree of cohesion was about the same for all slickens. Then, the threshold of erosion was measured for six randomly chosen samples in a small flume. The measurements indicate that the tailings comprising slickens appear to erode at surprisingly low boundary shear stresses, namely around one N/m^2. This result also was confirmed by calculations based on field observations during the 1997 bankfull flow, as discussed below. In all three types of critical shear stress determination, accuracies of only ± 30 percent were achieved consistently. The reason for the relatively low cohesion probably is the coarseness of the silt comprising most of the slickens.

Field Confirmation of Estimated Threshold Shear Stresses for Flood Plain Erosion

Owing to the nature of the soil, the field-based procedure described above for determining the threshold shear stress for erosion of channels on the most vulnerable half of the tabs along the Clark Fork through the Deer Lodge Valley is reasonably accurate. Nevertheless, an opportunity to test the method arose during a bankfull flow in June 1997. At several locations during this bankfull event, shallow flow crossed the flood-plain tabs. Erosion of the flood plain was observed at most of these locations and, thus, provided an opportunity for field testing the hypotheses that 1) the soil on the flood-plain tabs acted in a non-cohesive manner when subjected to a prolonged flow, and 2) form drag on the herbaceous plants was not a significant factor on poorly vegetated flood-plain tabs in June. Smith measured water depth at approximately ten locations in each of six of the shallowest of these cross-tab flow paths in which active erosion could be observed for a significant fraction, if not all, of the cross-tab distance. The measured stringers of flow were all located at sites where there previously had been no obvious channels across the tabs. From the measured flow depth in the shallow stringers of active erosion and the down-valley slope of the flood-plain tab, the average boundary shear stress beneath the sediment-eroding flow could be calculated and compared with the threshold shear stress for erosion calculated from the sedimentological characteristics of the local flood-plain soil.

Bed and bar morphologies and incipient-channel sinuosities were estimated for each of the six flood-plain channels and included in the calculation. Differences were insignificant so the final model treated the geometrical properties of the channels as fixed for each tab. The effective Manning coefficients calculated for each of the channels ranged from 0.042 to 0.048. Reach-averaged shear stresses were separated into components resulting from form drag on bed irregularities, form drag on channel scale irregularities, and skin friction using an updated version of the method of Smith and McLean (1977). As shown in Table 1, the skin frictions calculated from the measured flow depths were all about the same and were slightly higher than the critical shear stresses calculated from the soil parameters. The systematic discrepancy probably arises because all of the channels that were examined were actively eroding at the time the measurements were made. These measured boundary shear stresses place an upper bound on the threshold of flood-plain erosion on the measured tabs, and these tabs are estimated from aerial photographs to represent nearly half of the tabs along the Clark Fork upstream of Deer Lodge.

In the areas in which the stringers of cross-tab flow followed old channels on the flood plain the flow was much deeper, the boundary shear stresses were much higher than critical, and erosion was proceeding rapidly. These deeper flows also were straightening and widening by undercutting and removing clumps of better-developed sod and small shrubs. Subsequent observations at several sites indicated that erosion in the incipient channels that were previously used for erosion threshold measurements had proceeded partly by head cut formation next to various types of barriers. In subsequent years, all of the channels on the flood plain that had been observed to be active in 1997 had been filled with grass and small shrubs. Their traces, moreover, had been partially obliterated by trampling and other biologic activity. One year later, all of these channels were very difficult to identify.

Channelization of the flood-plain tabs along the Clark Fork in the Deer Lodge Valley appears to have occurred many times in response to small floods with shallow overbank flows during the decades since the 1908 flood. Nimick (1990) mapped the shapes of many large overbank-flow channels during his evaluation of the extent and thickness of the tailings deposits in the upper Deer Lodge Valley. These probably were associated with the largest floods in the last part of the 20th century, floods with a 10-year or so recurrence interval. Small- and moderate-sized overbank-flow channels do not persist for long. In open areas, they are sites of trampling associated with grazing and other forms of biologically produced diffusion of soil materials. This makes them discontinuous and allows sediment to col-

Table 1. Comparison of critical shear stresses for flood-plain erosion from field measurements of flow, from calculations based on grain size, and from flume measurements

[The values of skin friction calculated from the depths of incipient channels were determined at six sites on three different flood-plain tabs during the 1997 bankfull flow. These are compared to values of the critical shear stress calculated from the sizes of sediment comprising the flood-plain soil at two of those sites. Variables required to calculate the two shear stresses also are included in the table. At site C-1 no size analysis was done, because the soil was too cohesive to make calculation of the critical shear stress from the median grain size meaningful. Channels at location A are on a tab at a sharp bend at RM 19.7, downstream from the mouth of Dempsey Creek. Channels at locations B and C are on tabs downstream from the Perkins Lane bridge, in the vicinity of the Galen gage at RM 3.0.]

Location	A-1	A-2	A-3	B-1	B-2	C-1
Channel depth, m	0.06	0.05	0.06	0.05	0.07	0.014
Valley slope	0.0035	0.0035	0.0035	0.0035	0.0035	0.0043
Manning coefficient	0.046	0.048	0.046	0.048	0.045	0.042
Calculated skin friction, N/m^2	0.37	0.31	0.37	0.31	0.43	0.98
Median grain size of soil, mm	0.05	0.05	0.05	0.05	0.05	–
Critical shear stress for sediment erosion, N/m^2	0.32	0.32	0.32	0.33	0.33	–

lect in the pools during runoff events. In areas with higher shrub densities, these depressions in the flood-plain surface tend to be wetter and to be the locations of preferential grass and shrub growth, which can promote deposition of sediment from runoff or even small overbank flows. Those overbank-flow channels that occur in areas of moderate to dense willow coverage and that serve as cattle paths through the shrubs, tend to survive for a sufficient length of time to be reoccupied by subsequent overbank flows. The primary factor controlling the erosion of flood-plain channels, however, is a discontinuous line of thin or absent grass sod at the onset of the overbank flow. During a flood, overbank-flow channels often reform on the most erodible parts of the flood-plain tabs rather than following old paths that have become better vegetated. During deep overbank flows, the boundary shear stresses on the flood-plain surface are so high that new channels can easily form.

Comparison of Erosion-Threshold Estimates to Values Determined on an Eroded Flood Plain

Smith (2001) developed a more extensive version of the flow model described in this report and applied it to a site at which the flood plain of a narrow, single-threaded stream in Colorado was removed by a large flood in a few hours or less. Smith and Griffin (2002) later used the Smith (2001) model to examine the geomorphic transformation of that stream (an unnamed tributary of Carpenter Creek) in greater detail. In this situation, sandbar willows with an understory of wetland herbaceous plants protected the essentially unaltered sites, whereas, the fully eroded sites had vegetation and soil characteristics similar to those on the flood plain along the Clark Fork in the Deer Lodge Valley. Although subjected to over 3 m of overbank flow, the upper reaches of the unnamed tributary of Carpenter Creek were essentially unaltered, and the skin friction stresses remained less than 0.4 N/m^2. In contrast, where the sandbar willow shrubs were sparse and grazed range grasses dominated the flora, the flood plain was subjected to skin friction shear stresses in excess of 2 N/m^2. The estimated threshold shear stress in these non-wetland areas prior to removal of the flood plain by the large flood is about 0.4 N/m^2, similar to those estimated above for the flood plain along most of the Clark Fork in the Deer Lodge Valley. Catastrophic geomorphic transformation of the unnamed tributary of Carpenter Creek began in wetland areas with higher shrub canopy percentages and higher concentrations of herbaceous plants than in most areas along the Clark Fork in the Deer Lodge Valley. This catastrophic geomorphic change occurred at skin friction shear stresses ranging from 1.0 to 2.0 N/m^2, depending on the estimated shrub densities used in the calculation at the site of initiation of unraveling.

THEORETICAL CONSIDERATIONS

Model for Flow on a Shrub-Covered Flood Plain in a Meandering Fluvial System

Quantitative evaluation of flow process on flood plains during large floods requires the assistance of predictive, process-based models. Measurements made during large floods are likely to be inaccurate and the data sets are probably going to be incomplete, if any useful measurements can be made at all. After-the-fact observations also are fraught with errors and provide an incomplete understanding of the time-dependent flow and sediment-transport processes active during the flood. As a consequence, the hydraulics of large floods needs to be characterized in comprehensive fluid-mechanical terms through process-based models that tie together the available measurements into a tight scientific understanding. Owing to the paucity of accurate, high-discharge information that can be used for calibration of flood-plain models, it also is dangerous to use simulation models for deep-overbank flows. Consequently, the most accurate approach for investigating large floods on shrub covered flood plains is to employ predictive models that embody the salient fluid- and sediment-mechanical processes and to use all available data to evaluate the accuracy with which the model represents the relevant aspects of the measured floods.

The first step in the construction of a predictive process-based model is to identify the salient processes that must be accurately characterized. It also is desirable to eliminate all non-essential processes from the model. When dealing with complex natural systems it also is desirable to use scaling relations among important variables whenever possible. The salient characteristics of flow over shrub- and grass-covered flood plains result from the various ways that the fluid motion is resisted by obstructions on the bed and in the interior of the overbank flows. The obstructions on the flood-plain surface and in the interior of the flow both can be simplified using scaling relations.

On the beds of rivers, the topographic elements that produce flow resistance scale with the width or depth of the channel and with the median diameter of the bed material. For example, bars have heights that are related to bankfull flow depth, and they have lateral and downstream lengths that are related to river width (Smith and McLean, 1984). Similarly, dunes have heights and wavelengths that scale with flow depth (Smith, 1970). Small-scale topographic features on the bottoms of gravel bedded streams scale with the median bed-particle diameter at that location on the streambed, and the larger scale topographic irregularities on the beds of gravel bedded streams have depressions that scale with median particle-diameter and spacings that scale with flow depth. In the case of shrub covered flood plains, the systematic topographic elements on the flood plain have lengths that are related to the spacing between shrubs in the down-valley direction and heights that are related to the diameter of the group of stems that comprise a single shrub. Topographic elements on the channel bottom and on the flood-plain surface, when their heights are small relative to the flow depth, are treated here using the method of Smith and McLean (1977). The drag coefficients that were used for these bed features are as determined by Smith and McLean (1977) and as calculated by Kean (1998) using the measurements of Hopson (1999). In all cases where the bed features have heights that are small relative to the flow depth, the flow is assumed to separate at the crests of the topographic elements.

The other main source of resistance to overbank flow is drag on the stems and branches of the shrubs. The average physical properties of the shrubs also follow scaling relations. For a large group of shrubs these relations permit a relatively simple model to be constructed from the appropriate set of non-dimensional lengths and the average stem diameter. Proper specification of these non-dimensional lengths and the average diameter of the stems from field data are essential inputs to the model. A typical willow or water birch shrub along the Clark Fork in the Deer Lodge Valley contains an average group of about 36 stems in a clump under a single canopy, and the ratio of canopy diameter to the diameter of the group of stems that comprises a shrub is about five. The ratio of stem-group diameter to stem diameter is about 13, and in the lower meter, which is the zone of primary concern in this report, the average cross-sectional area of a clump remains approximately constant with height. Therefore, a typical shrub can be modeled as a group of 36 cylindrical stems, each with a diameter of 40 mm, with a stem-group diameter of 0.5 m and a canopy diameter of 2.5 m. As the shrubs become dense, their canopies intermingle and it is no longer possible to resolve individual shrubs accurately on high resolution aerial photographs. Moreover, the ratio of canopy diameter to stem-group (shrub) diameter decreases. In a dense carr, the stem groups can be closer than one group diameter apart.

Computation of flow around and through an individual shrub is possible but unnecessary for the present purpose. Once there are many shrubs in all directions, the flow responds to the average properties of the stand, not to the individual shrubs. In this case, the stems of the groups act in the same manner as when they are uniformly dispersed. Flow accelerations can be included in the model, but this is not necessary unless a shrub band less than five-average-canopy-diameters thick (12 to 13 m) is of concern. As a consequence, a relatively simple model was constructed using the basic principles of fluid drag.

When a flow is blocked by bedforms or other topographic elements that have heights short relative to the flow depth, these topographic elements produce an outer velocity field in which the topographic elements act as roughness elements and an inner velocity field in which the topographic elements act to accelerate and decelerate the flow.

The theoretical approach followed in this report is similar to that of Smith (2001), except that there is no need to employ a three-layer model for the Clark Fork case. The floods of concern in this report are less than 2 m deep and do not overtop the large willows. When, in addition, the trees and shrubs do not bend over, as is the case with most riparian trees and many types of willows that are associated with active and extant beaver ponds in the Northern Rocky Mountains, then only one flow layer need be considered. This is the situation on the flood plain of the Clark Fork through the Deer Lodge Valley.

In an unaccelerated, homogeneous-flow, stress divergence balances the downstream component of the fluid weight per unit volume. In the case of flow through a field of stems and trunks that penetrate the free surface, this force per unit volume balance needs to be modified by subtracting the spatially varying drag forces per unit volume on the shrubs and trees. This drag force acts like a pressure gradient that opposes the flow and reduces the velocity and the boundary shear stress fields. For a steady, horizontally uniform flow, the force balance per unit volume becomes

$$\frac{\partial \tau_{zx}}{\partial z} = \rho g(\sin\alpha) + \sigma_D \frac{\tau_b}{h}, \tag{1a}$$

where

$$\sigma_D = \frac{F_D}{\lambda^2 h\left(\frac{\tau_b}{h}\right)} = \frac{C_D}{2k^2}\left(\left(\ln\frac{h}{z_0}\right) - 0.74\right)^2 \cdot \frac{hD_s}{\lambda^2}. \tag{1b}$$

In (1a), τ_{zx} is the shear stress on planes parallel to the average bed of the flow, z is the distance above the average bed, ρ is the density of the fluid, g is the acceleration due to gravity, α is the slope of the bed relative to the horizontal (using the conventional mathematical definition so that a downward sloping bed gives α a negative value), σ_D is a drag function defined by (1b), $\tau_b = (\tau_{zx})_b$ is the shear stress on the average bed, $z = s$ is the equation for the free surface, $z = \eta$ is the equation for the bed, and $h = s - \eta$ is the thickness of the water layer protruded by the trunks of trees and the stems of shrubs. In (1b), F_D is the drag force on the woody vegetation, λ^2 is the mean area affected by a stem, C_D is the drag coefficient for a stem or branch, $k = 0.408$ is von Karman's constant, z_0 is the roughness parameter for the mean bed, and D_s is the mean diameter of the stems producing the drag on the flow.

Integrating (1a) from $z = 0$, where $\tau_{zx} = \tau_b$, to the free surface where $z = s$ and $\tau_{zx} = 0$, and solving for τ_b, gives

$$\tau_b = \frac{(-\rho g h(\sin\alpha))}{1 + \sigma_D}. \tag{2}$$

Next, integrating (1a) from z to s, rearranging, and equating τ_{zx} to the shear ($\partial u/\partial z$) using an eddy viscosity (K) yields

$$\tau_{zx} = \tau_b\left(1 - \frac{z}{h}\right) = \rho K \frac{\partial u}{\partial z}. \tag{3}$$

Following Smith (2001), the eddy viscosity can be written in two parts. Near the bottom of the flow, the turbulent eddies increase in scale with distance from the boundary (Schlichting, 1979), and dimensional analysis requires that the eddy viscosity vary linearly with this distance. Further away from the boundary, the effect of the boundary on the turbulent eddies becomes small and the length scale for the turbulence becomes the flow depth. In this region, the eddy viscosity becomes constant with distance from the boundary. Using the method of Rattray and Mitsuda (1974), the two velocity profiles can be matched at 0.20*h*. It should be noted, however, that the analysis employed here gives a slightly different and somewhat more general result than that obtained by Rattray and Mitsuda (Smith, 1999). The fully specified, two-part eddy viscosity is as follows:

$$K = ku_*z\left(1-\frac{z}{h}\right), \text{ when } z \le z_m = 0.20h \qquad (4a)$$

and

$$K = \frac{ku_*h}{\beta}, \text{ where } \beta = 6.24, \text{ when } z \ge z_m = 0.20h \qquad (4b)$$

where $u_* = (\tau_b/\rho)$ is the shear velocity.

Substituting (4a) into (3) and integrating from $z = z_0$ to $z \le z_m$ gives

$$u = \frac{u_*}{k}\left(\ln\frac{z}{z_0}\right), \text{ for } z \le z_m. \qquad (5a)$$

Similarly, substituting (4b) into (3) and integrating from z_m to $z \le$ s gives

$$u = \frac{u_*}{k}\left(\beta\left(\left(\xi-\frac{1}{2}\xi^2\right)-\left(\xi_m-\frac{1}{2}\xi_m^2\right)\right)+\left(\ln\frac{\xi_m}{\xi_0}\right)\right) \qquad (5b)$$

for $z \ge z_m$. Here, $\xi = z/h$, $\xi_m = z_m/h$, and $\xi_0 = z_0/h$.

The average velocity field given by (5b), between z_0 and s, is

$$u = \frac{u_*}{k}\left(\left(\ln\frac{z}{z_0}\right)-0.74\right). \qquad (6)$$

Equation (6) already has been used in (1b) to represent the velocity field in the expression for form drag on the stems of the shrubs. In any problem involving form drag on an object in a flow, the choice of a reference velocity is not at all arbitrary, but rather it is dictated by basic fluid mechanical principals. The *only* proper reference velocity is *the velocity averaged over the region encompassed by the object were the object removed.* If there were only one stem obstructing the flow, then the proper velocity to use would be that of the undisturbed flow at the appropriate location, but if many stems obstruct the flow, then use of the undisturbed flow is not appropriate. In this latter case, the proper velocity to use is that affected by all of the other stems. When there are many stems present, the removal of only one of them has an inconsequential effect on the rest of the flow. Therefore, the appropriate velocity to use in a many object drag problem is the velocity that results from removing the flow momentum through the drag on *all* of the obstructions.

In the case of equation (6), the shear velocity carries all of the information on what type of flow is being considered. If the appropriate reference velocity were from the undisturbed flow, then the shear velocity would be obtained from the boundary shear stress calculated using the depth-slope product. In contrast, for the overbank-flow problem of concern in this report, the appropriate shear velocity is obtained from the boundary shear stress given by equation (2). The former boundary shear stress is for the unaltered flow and the latter is reduced by the momentum lost through form drag on all of the stems. In drag problems involving natural systems, proper choice of the reference velocity makes the difference between a successful model and a useless one. It is mandatory that the correct reference velocity be used in a predictive, drag-based model. Were the stems close enough together for the accelerating flow around one stem to enhance the velocity at the location of another stem, then this effect also would have to be included in the model. Once stems are three or more diameters apart, however, this effect can be neglected. Similarly, when individual shrubs are closer than two stem-clump diameters apart, accelerations around the clumps need to be taken into account. Although the calculations in this report border on requiring inclusion of both of these accelerative effects, the cases of interest have the clumps more dispersed and the inclusion of accelerations around the stems and the stem-clumps is of little ultimate benefit.

Coupling of Overbank and In-channel Flows in Meandering Fluvial Systems

When examining flow over a flood plain during an overbank event, flow in the channel cannot be ignored. If a flood plain is broad relative to the width of the channel, such as is the situation with the meander belt of the Clark Fork in the Deer Lodge Valley, then a small increase in stage produces a large increase in overbank discharge. In the channel, however, this increase in stage produces only a small increase in discharge. Consequently, as shrub density on such a flood plain decreases, the flow depth decreases and velocity over the flood plain increases. This, in turn, results in an increase in discharge over the flood plain and a decrease in discharge through the channel. This effect is demonstrated in a subsequent section for the Clark Fork in the Deer Lodge Valley. In this report, coupling between channel hydraulics and flood-plain hydraulics is only through discharge. Although some mixing occurs at the edges of the flood plain, this does not have a profound effect on either of the two types of flows and is not a zero order issue. Similarly the effects of channel curvature also are of higher order. The approach used herein is to calculate flow in an artificially deepened channel and on the flood plain separately and then require the stage to be the same. The solution is iterated until the two discharges sum to the prescribed total value.

COMPUTATION OF FLOW IN THE CLARK FORK IN THE DEER LODGE VALLEY DURING FLOODS

Application of the Model to the 1908 Flood in the Vicinity of the Galen Gage

The flood of record in 1908 was not well documented in the written records of that period. Discharge was measured at a gage near Missoula (USGS gaging station number 12340500), below the newly constructed Milltown Dam. As a consequence, it included flow from both the Blackfoot River and the Clark Fork. Owing to the long duration and large magnitude of the flood, it is reasonable to partition the flow into drainage basins using drainage area. Following this approach and using the flood-frequency analysis for the upper Clark Fork presented by Smith and others (1998), we estimate the recurrence interval of the flood to be 370 years at Perkins Lane near the Galen gage (USGS gaging station number 12323800). This location is used because detailed topography is available in the form of a 0.6-m (2-foot) contour map for this part of the flood plain. That map extends slightly downstream of Perkins Lane (Schafer and Associates, 1996). With this contour map and the delineated edges of the tailings deposit, a cross-section and the wetted perimeter of the 1908 flood can be constructed. Although the tailings become indistinct near their margins, the careful work of Nimick (1990) and Schafer and Associates (1997) in this area permit reasonable flow margins to be determined.

It is assumed that tailings originally were deposited as a drape right up to the flow margins and then the margins became diffuse through various forms of bioturbation. This assumption gives an overbank flow depth next to the main channel of 1.6 m. The cross-sectional geometry and the flow depth probably are better known than the discharge, which at this location was calculated from the uniform-rainfall analysis to be 130 m^3/s. Perhaps the poorest known variable associated with the 1908 flood is the state of woody vegetation on the flood plain. Owing to the comments of Warren Ferris and Father De Smet, it is clear that the riparian vegetation in the mid 1800's was very dense, and it is likely that the riparian shrubs were still quite dense in 1908. With estimates of cross-sectional area, wetted perimeter, and discharge, shrub density can be back calculated using the above-described model. When this is done the calculated shrub density turns out to be quite high.

For these calculations, it is assumed that during a large flood, the topography of the channel will adjust to the higher stage as it apparently did during the 1908 event. Calculations done for the 1997 bankfull flow in the neighborhood of the Galen gage indicate that the river, at present, has an exceedingly smooth bed. Specifically, the measured discharge at the gage can be matched with a reach-averaged model using only grain roughness. That is, the topographic elements are all so small that form drag on them provides a negligible contribution to the total boundary shear stress. In contrast, the optimal model for the 1908 flood in the channel of the Clark Fork requires that the channel be roughened by the bed topography of a typical gravel bed stream with an occasionally mobile bed. The 1908 model included well-developed point bars and two typical types of topographic elements on the riverbed. These were (1) short wavelength features that have lengths and heights that both scale with bed particle size and are best called pebble clusters and (2) longer features that have heights related to pebble size but have lengths that are related to water depth. These latter features are about twice as high as the pebble clusters and appear to be related to intense coherent turbulent structures called sweeps. Smith thinks that they are poorly formed, incipient gravel ripples that begin to develop during the sweeps because that is the only time the sediment transport rates are high enough for gravel ripples to form. Both features are typical of streams with occasionally mobile gravel beds and must be included in predictive, process-based models. Otherwise, the gravel beds do not produce enough friction and have sediment transport rates that are too high. The wavelength-to-height ratio for the pebble clusters in the model is 10 and that for the longer features is 30.

Even during the 1908 flood skin friction was so low that clasts on the present bed of the Clark Fork in the upper Deer Lodge Valley would have barely moved. At present the larger pebbles and cobbles rarely move and when they do move it is because they are surrounded by sand. The marginal transport keeps the point bars stunted. As the gravel approaches the significant transport regime and the clasts start rolling, the bed starts to become uneven and the topographic elements extract enough momentum to return the bed to marginal transport. In this manner the skin friction remains low until fully developed gravel bars and gravel bedforms evolve. One way to model this situation is to hold the skin friction constant in the marginal transport regime and let the topographic elements on the gravel bed increase in height just enough to remove the excess momentum. In this type of process-based model, the pebble clusters develop. The gravel bars develop next, and, finally, the gravel ripples start to form. Both types of channel models were tried for the Clark Fork in the Deer Lodge Valley during the 1908 flood and both gave essentially the same result.

The flood-plain component of the coupled flood model was a straightforward application of the equations presented in the section on theory. As mentioned in that section, spacing of the topographic elements was determined by the distance between shrubs. The height of these elements was held constant at 0.10 m. For the dense shrub situation in 1908, this approach worked well, but for subsequent calculations pertaining to the present flood plain, which is sparsely vegetated with shrubs, this approach would have yielded a flood plain that was too smooth. Therefore, a residual roughness with a length-to-height ratio of 10 was left independent of shrub spacing for these latter cases. Although this fixed residual roughness is higher than warranted for the present Clark Fork flood plain, it was used to provide a conservative estimate of the present flood-plain vulnerability.

Calculations for the 1908 flood, constrained by the measured stage and estimated discharge yielded a most probable shrub canopy in excess of 80%, with an estimated range of 75 to 90%.

Application of the Coupled Model to Potential Future Floods on the Clark Fork

The approach that forms the basis for this report has been used successfully to predict the initiation of catastrophic geomorphic change of a headwater drainage basin of East Plum Creek, Colorado (Smith and Griffin, 2001). Although the available data for each of the East Plum Creek test cases were not as comprehensive and precise as would have been desirable, the process-based model predicted no catastrophic geomorphic change where none occurred and predicted catastrophic geomorphic change at the sites where it did occur. The model also correctly predicted the threshold location downstream of which catastrophic geomorphic change occurred along the unnamed tributary of Carpenter Creek. In addition, the model provided what appears to be a reasonable estimate of shrub density along the Clark Fork during the 1908 flood, based on the reports of two early explorers. Better data on catastrophic geomorphic change or resistance to geomorphic alteration during large floods in fluvial systems similar to the Clark Fork are not presently available, so no more accurate tests of the model are possible. Nevertheless, the computational approach used in these two very different situations was successful, and the procedure appears to provide an accurate means of evaluating flood plain vulnerability. In this section, the computational procedure is used to evaluate the present state of the fluvial system of the Clark Fork in the Deer Lodge Valley with respect to vulnerability to large floods. Computations with the model employ the same structure as was used to investigate the 1908 flood, but the calculations are carried out for floods with recurrence intervals ranging from 5 years to 300 years and for a range of shrub densities.

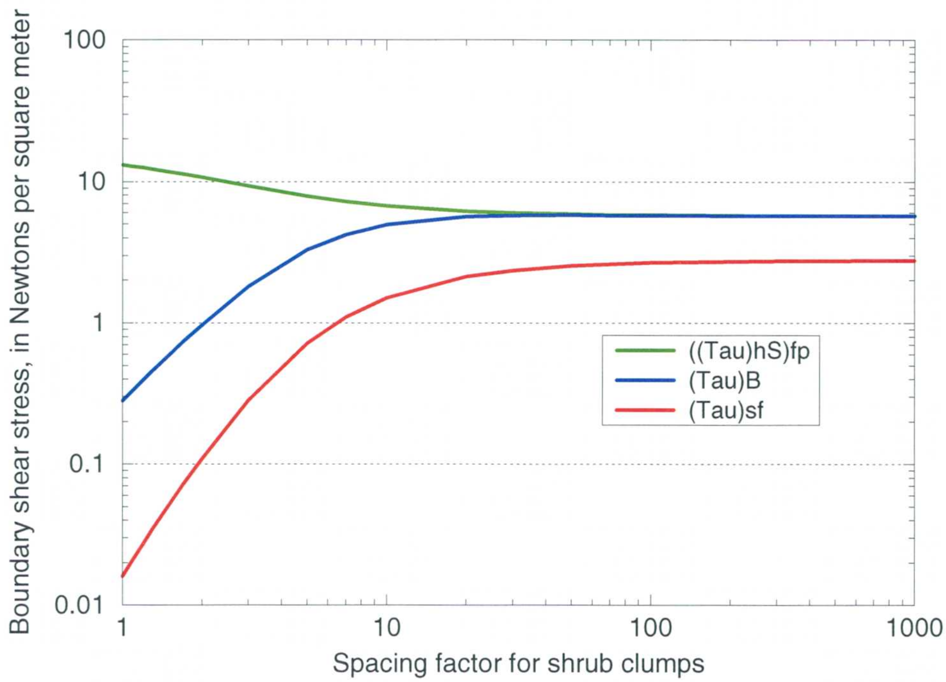

Figure 3. Comparisons of three types of the boundary shear stress as a function of shrub spacing factor during a 5-year-recurrence-interval flood. The top curve is the boundary shear stress from the depth-slope product ((Tau)hS), using the flow depth on a shrub-covered flood plain. This (total) boundary shear stress is reduced by form drag on the shrubs to give the boundary shear stress on the average surface of the flood plain, denoted τ_B = (Tau)B. The velocity over most of the flow depth is scaled by this part of the boundary shear stress; therefore, it must adjust with shrub spacing to accommodate the discharge boundary condition. When form drag on the topographic elements of the flood plain surface is removed from τ_B the remaining component of the boundary shear stress is the skin friction ((Tau)sf).

The primary issue is whether or not the skin friction exceeds the threshold shear stress for flood-plain erosion for any or all floods with these recurrence intervals when shrub densities typical of large, contiguous segments of the flood-plain tabs on the Clark Fork in the Deer Lodge Valley are used. In order to provide more insight and generality, the results of the calculations are presented as continuous functions of shrub spacing. A linear measure of spacing between shrubs is used. This then can be converted to canopy cover. Owing to the linear nature of the spacing factor it must be squared and divided into the reference state to get canopy cover. As shrubs get closer together, their canopies intermingle and what appears to be continuous canopy cover on an aerial photograph can have a range of shrub spacings on the ground. The reference state used here is a shrub spacing of one shrub diameter. The canopy cover does not begin to drop below 90%, however, until the shrub spacing exceeds a value of five. The shrub spacing factor used in this report (SpF) is 4 for this case. This shrub spacing factor is a linear measure of distance between shrubs beginning with a reference state in which the shrubs are located one stem group diameter (basal shrub diameter) apart (SpF = 1). Adding one to the spacing factor then multiplying it by the basal diameter of the stem group in meters gives the actual spacing between shrub centers in meters. Although the use of a spacing factor is not as easy to relate to aerial photographs, it is more representative of the actual physical situation and, thus, is more accurate.

Figure 3 compares skin friction with the boundary shear stress on the average surface of the flood plain for a 5-year recurrence interval flood. The model treats the topographic elements as roughness elements. It also shows the boundary shear stress calculated from the depth-slope product using the actual depth. In reality, the boundary shear stress would never be any higher than the asymptotic value because the depth is elevated for a given discharge by the substantially reduced velocity. The difference between the skin friction and the other two types of boundary shear stresses at high spacing factors is a result of the artificially roughened flood plain. Some grass-covered flood plains may be this rough, but usually the grass bends over and drastically smooths the flood plain surface.

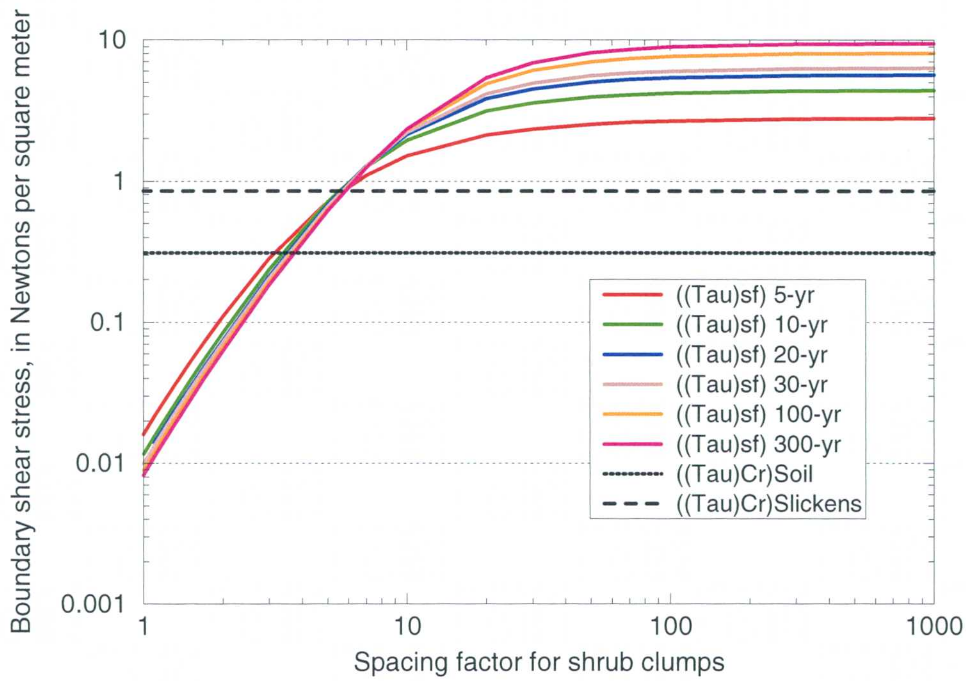

Figure 4. Skin friction ((Tau)sf) as a function of shrub spacing factor for floods with recurrence intervals ranging from 5 to 300 years. The information in this figure is the same as that displayed in figure 5, but the logarithmic ordinate emphasizes the structure of the skin friction with small spacing factors. The lower horizontal line represents the critical shear stress for the flood-plain soil (((Tau)Cr)Soil), and the upper one represents the critical shear stress for the tailings in the slickens (((Tau)Cr)Slickens).

This particularly is the case early in the growing season. Although figure 3 is for a flood with only a 5-year recurrence interval, the analogous graphs for floods with longer recurrence intervals are similar. The critical shear stress for the erosion of fine sand is less than 0.3 N/m^2, and the skin friction is below that value only for shrub spacing factors of less than approximately four.

Recall that a spacing factor of four means that the shrubs are still only about five clump diameters, or 2.5 m, apart and their individual canopies are just beginning to lose contact. For spacing factors greater than eight (spacings of 9 clump diameters or 4.5 m), the skin friction is greater than one and, thus, erosion of both the flood plain soil and tailings would occur even for this very low recurrence interval flow. A spacing factor of 4 translates to a canopy cover of over 90%, but a spacing factor of 8 translates to a canopy cover of less than 30%. Griffin and Smith (2002) report that 30% canopy cover is the average value for all tabs upstream of Deer Lodge. However, most tabs upstream of Deer Lodge have a few areas of higher canopy densities and broad cross-tab strips that display much lower canopy covers making them good candidates for rapid flood-plain erosion. It appears from figure 3 that for small floods, the transition from a well-protected flood plain to a poorly protected one begins when the canopies of the individual shrubs begin to lose contact.

From the point of view of flood plain stability, uniform cover with a dense shrub flora is the most desirable state. In contrast, dense bunches of shrubs with wide cross-tab strips devoid of shrubs is the most unstable condition. In the latter case, the dense bunches funnel the flow across the tab, enhancing the shear stress on the areas devoid of shrubs. This same condition results if there are only a few rows of shrubs on the down-valley sides of the tabs. Here the flow is accelerated through the gaps between shrubs, exerting very high shear stresses on the flood-plain surface and initiating head cuts that can propagate across the tab surface very rapidly. In contrast, if there is a large number (many tens) of shrubs along a flow path, the resistance of the shrubs becomes sufficient to reduce the velocity and boundary shear stress below the critical shear stress for erosion even along paths between individual shrubs and, thereby, to prevent head cutting. The single, double, and triple lines of shrubs that line the river banks now in many places are effective for reducing bank erosion but are detrimental to protection of the flood plain from surface erosion during overbank flows. Fortunately, most low-recurrence-interval floods do not last long enough to cause the tabs to erode significantly in such areas. For floods caused by rain or rain-on-snow, flood duration tends to be positively correlated with recurrence interval, and the larger floods are more likely to last long enough to promote catastrophic geomorphic change.

Figure 4 shows skin friction as a function of the spacing factor for floods with recurrence intervals ranging from 5 to 300 years. A dense shrub carr (SpF < 4) is extremely effective at protecting the flood plain from floods of all recurrence intervals. Once a shrub carr becomes dense, the hydraulics of the flow depends on stem density rather than flow depth. For very dense carrs, velocity and boundary shear stress become completely independent of flow depth until the shrubs are overtopped. The lower horizontal line on figure 4 is an estimate of the critical shear stress for erosion of the flood plain soil, as described in a previous section, and the upper horizontal line is an estimate of the critical shear stress for erosion of slickens. If a conservative critical shear stress of 1.0 N/m^2 is used, then this yields a critical spacing factor of approximately 8. Therefore, for canopy covers that exceed approximately 30% and in which the shrubs are uniformly distributed over the flood plain, the flood plain is protected and the recurrence interval of the flood is of consequence only because it is related to flood duration. In this regime the skin friction depends primarily on stem density. For spacing factors between 8 and 30, the hydraulics of the flow depends on both stem spacing and discharge (flow depth), and for spacing factors greater than 30, the flow hydraulics depends on flow discharge (flow depth) but not on stem spacing. A flood-plain surface may or may not erode when the SpF is greater than 30 (a spacing of 16 m and a canopy cover of less than 0.1% for the Clark Fork in the Deer Lodge Valley) depending on the state of the herbaceous vegetation on its surface. The flood plain cannot erode once 90% canopy cover is achieved and might not erode if the canopy cover is substantially above 30%.

Below the horizontal line representing the critical shear stress for slickens, the boundary shear stresses for the various recurrence-interval floods are all about the same. This indicates that the dominant variation is with shrub spacing and not flow depth. In fact, these curves reverse their order relative to that above the critical shear stress for slickens, with the largest floods having the lowest boundary shear stresses. This reversal is a consequence of the partition between the flow in the channel and that over the flood plain. In the dense shrub situation, as the discharge increases, a decreasing fraction of the flow goes over the flood plain. In this asymptotic regime the flow velocity is the same for flows of all recurrence intervals. The decreasing discharge, therefore, leads to a decreasing depth and, hence, to a decreasing boundary shear stress.

Figure 5 presents the same information as did figure 4, but the linear ordinate emphasizes the transition between the dispersed shrub situation and the dense shrub case. This figure emphasizes the asymptotic dependence on discharge and depth rather than the dependence on stem spacing. Figure 5 also shows more clearly the transition from control by stem spacing to control by flow depth. Specifically, it demonstrates that the primary difference between small floods and large ones is that the small floods retain the lower skin frictions as these woody plants become more widely spaced. Although shear stresses rise more sharply for the larger floods in the moderate shrub spacing region, the differences in boundary shear stresses between small floods and large ones are primarily in the sparse shrub domain. As mentioned above, larger floods tend to erode more sediment partly because they last longer. However, once the shrubs are numerous, dense and uniformly spaced, erosion is impossible and the duration of the flood is of no consequence. The range of skin friction is nearly three orders of magnitude (from 0.01 to 9.0 N/m^2).

Figure 6 is analogous to figure 5, but it presents vertically averaged velocity rather than boundary shear stress. In turbulent boundary layers, shear velocity controls the velocity field, so it is not surprising that figures 5 and 6 are similar. The shear velocity that controls the vertically averaged velocity is that on the average flood-plain surface and not the skin friction; nevertheless, from figure 3 it is clear that the shape of the skin friction and the shape of the boundary shear stress on the average flood-plain surface are similar. All of the points made with regard to figure 5 pertain also to figure 6. In addition, it is worth noting that the velocity on the flood plain is less than 0.25 m/s when the shrub spacing factor is less than three. The range of vertically averaged velocity is from 0.15 to 2.25 m/s.

Owing to the high friction from the shrubs, it is of interest to determine how the flow depth varies with the spacing factor (fig. 7). The largest changes in flow depth are for the largest floods. That is, dense shrubs are most effective in decreasing the velocities and boundary shear stresses produced by the largest over-

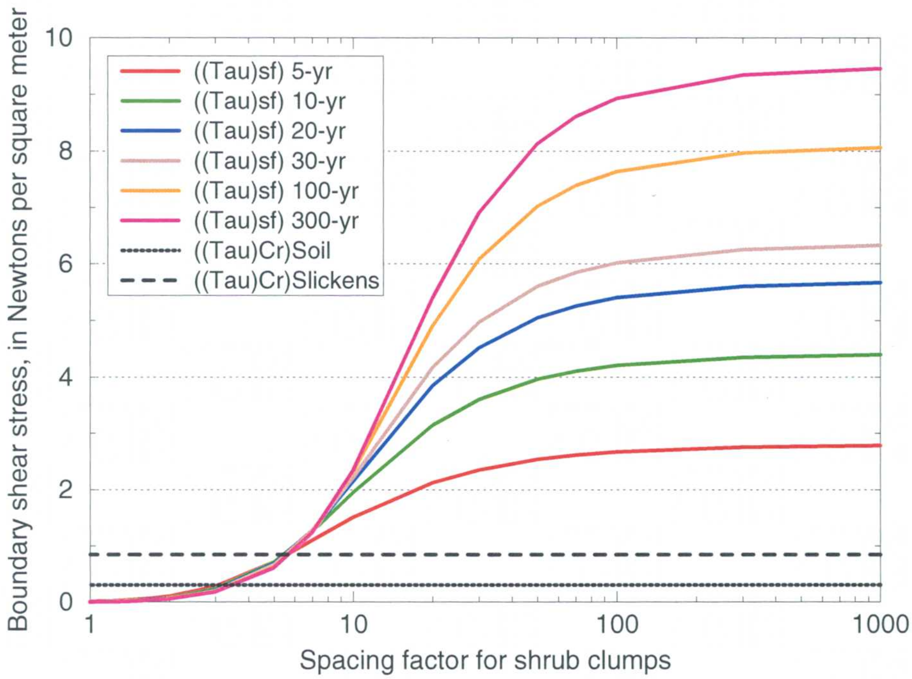

Figure 5. Skin friction ((Tau)sf) as a function of shrub spacing factor for floods with recurrence intervals ranging from 5 to 300 years. The information in this figure is the same as that displayed in figure 4, but the linear ordinate emphasizes the structure of the skin friction with intermediate and large spacing factors. The lower horizontal line represents the critical shear stress for the flood plain soil (((Tau)Cr)Soil), and the upper one represents the critical shear stress for the tailings in the slickens (((Tau)Cr)Slickens).

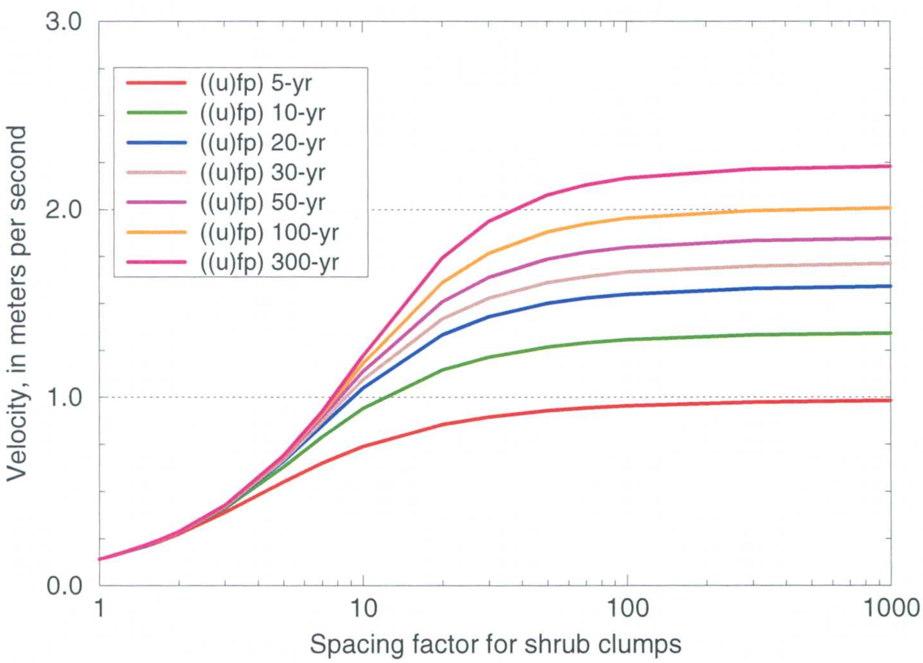

Figure 6. Vertically averaged velocity ((u)fp) as a function of shrub spacing factor for floods with recurrence intervals ranging from 5 to 300 years.

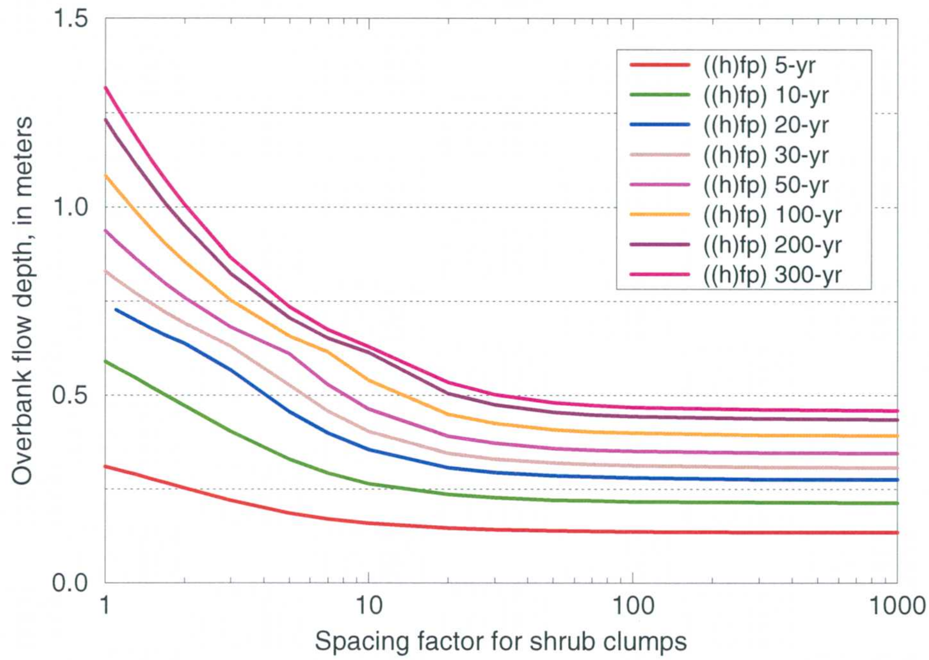

Figure 7. Flow depths ((h)fp) as a function of shrub spacing factor for floods with recurrence intervals ranging from 5 years (bottom) to 300 years (top).

bank flows. Nevertheless, they also are quite effective at reducing these variables for small floods. Removing dense shrubs increases the flow velocity and reduces the flood-plain flow depth from 1.30 to 0.46 m for a 300-year-recurrence-interval event. For a 10-year-recurrence-interval flood, removing the shrubs decreases the flow depth from 0.3 to 0.13 m. From this diagram it is clear that shrubs have some effect on the overbank flow depth until the shrub spacing drops below approximately 30. As mentioned above, a spacing factor of 30 is equivalent to a physical distance of 16 meters between 0.5 m diameter plants with canopies approximately 2.5 m in diameter. The short wavelength variations in curve shape near the spacing factor of ten are a result of the topography of the flood plain. The 5- and 10-year recurrence interval floods never get deep enough to be affected by this abrupt increase in the overbank flow width.

Figure 8 shows how the discharge partitions as a function of the spacing factor for 10- and 100-year recurrence interval floods. The upper straight line is the total discharge for the 100-year flood and the lower straight line is the discharge for the 10-year flood. The lines that slope down to the right represent the component of discharge over the channel bed and the lines that slope up to the right represent the component of discharge over the flood plain. Note that the discharge over the flood plain increases as the shrubs get sparse. The changes in discharges over the flood plain are shown for eight different recurrence intervals in figure 9. These curves are not geometrically similar because of the topography of the flood plain. The longer recurrence interval floods display the same variations in flow partitioning as in flow depth. The value of the spacing factor at which the shrubs lose their importance rises from about 10 to about 30 with increasing flood magnitude. This occurs because drag on the stems increases linearly with flow depth.

DISCUSSION

Owing to the predictive, process-based nature of the fluid mechanical model used in this report, the results presented in the previous section do not depend on coefficients with unspecified values that must be set empirically. As a consequence, the trends displayed in figures 3-9 should be considered reliable. In contrast, the numerical outputs as applied to the Clark Fork depend on the accuracy of the input parameters. For the core model, these range from the physical characteristics of the shrubs to the geometric properties of the bed features to the cross-sectional geometry of the flood plain. In addition, converting discharge to recurrence interval requires an accurate flood frequency function, and converting skin friction to erosion rate

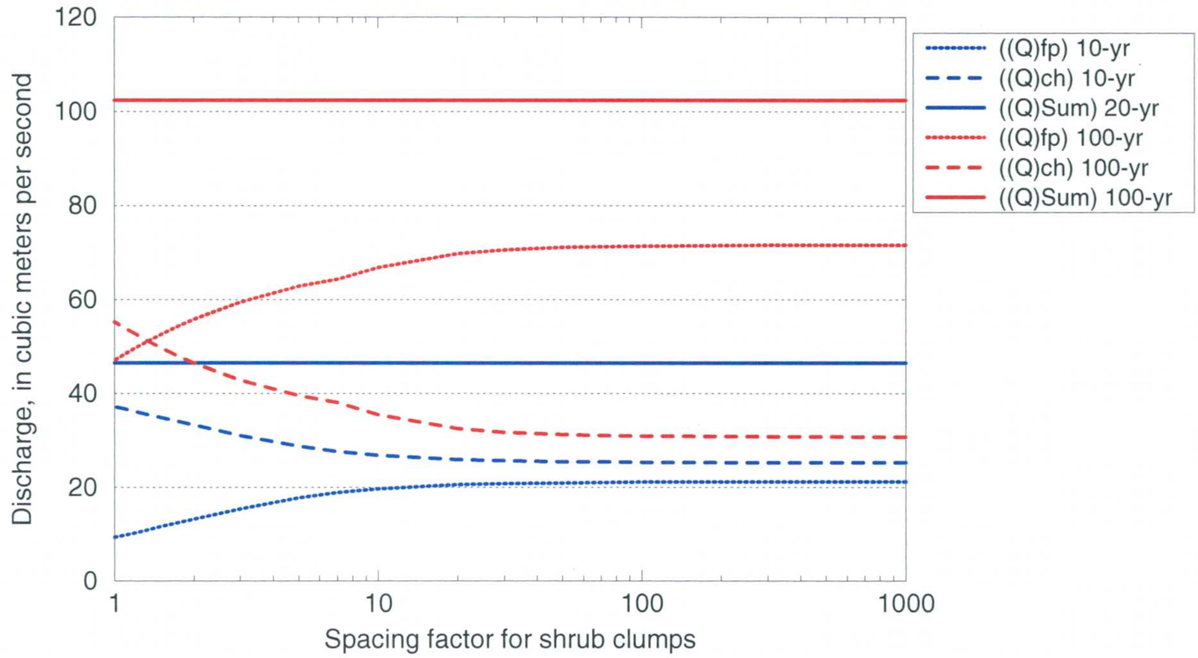

Figure 8. Through-channel and overbank components of discharge as a function of shrub spacing factor for floods with 10-year and 100-year recurrence intervals. The horizontal, straight lines represent the sums of the two discharge components. The downward sloping curves represent the component of discharge through the channels and the upward sloping curves represent the component of discharge over the flood plain.

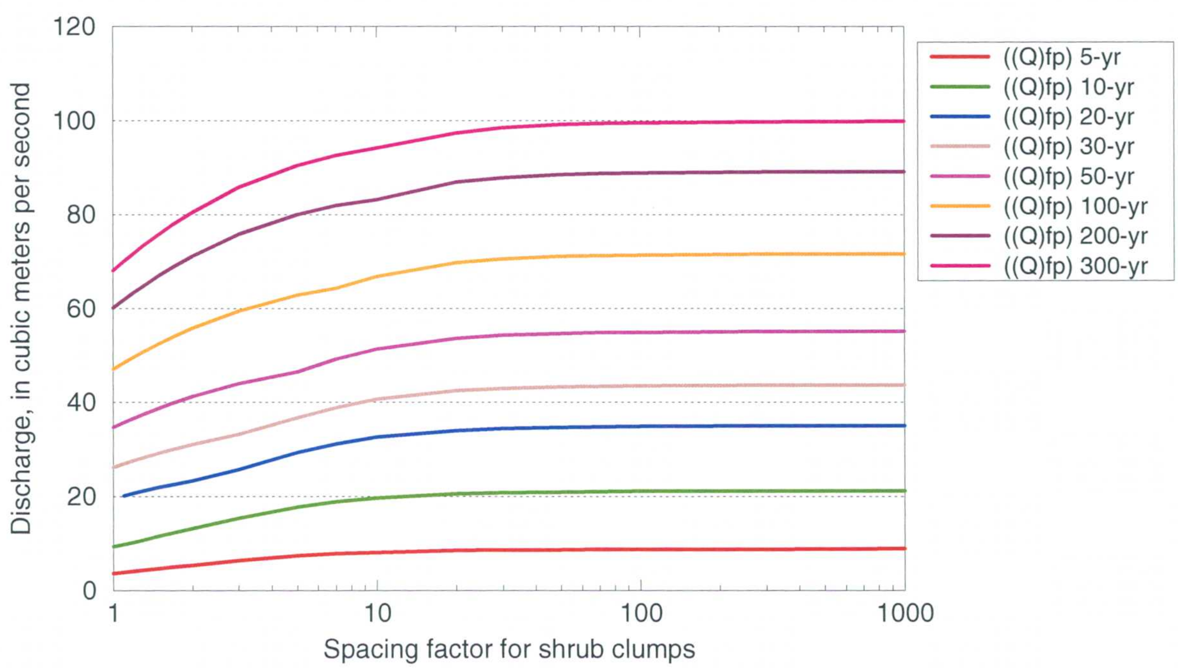

Figure 9. Overbank components of total discharge ((Q)fp) as a function of shrub spacing factor for events with recurrence intervals ranging from 5 to 300 years.

requires accurate estimates of the two components of the threshold shear stress. The values used in the calculations all come from measurements made on the Clark Fork upstream of Deer Lodge. Most of these measurements emphasize the area near the Galen gage because this is the only location at which an accurate valley cross section could be reconstructed. Further examination of the Clark Fork flood plain upstream of Deer Lodge would greatly benefit from precision meander-belt topography determined using LIDAR (light detection and ranging systems). All of our input parameters should be considered estimates. The model based on them gives the general flavor of the fragility of the Clark Fork to floods, but if an accurate evaluation of either the entire fluvial system or of particular sites is desired, then a comprehensive sampling program is essential.

It was not the aim of the investigation on which this report is based to carry out a comprehensive sampling program on the Clark Fork. Only a small amount of field work was done, and most of the field work was in the upper part of the Deer Lodge Valley. The application in this report is considered by the authors to be reasonably accurate for the Galen area and to be fairly speculative for the area between Galen and Deer Lodge. To do a significantly better job at locations other than near the Galen gage would require a comprehensive sampling program. This would involve (1) surveying tens of cross-valley cross sections or using LIDAR data to get the meander belt topography throughout the fluvial system, (2) obtaining the micro topography of the surfaces of the flood-plain tabs and relating these to their origins, (3) sampling many hundred shrubs on several tens of flood-plain tabs to determine the general physical properties of the shrubs and relating these to the individual and aggregate characteristics of their canopies, and (4) making hundreds of inverted flume measurements of the erosion threshold shear stress on tens of flood-plain tabs. With these types of data, however, a very accurate application of the model would be possible.

The results of this flood-plain analysis still would have to be stated in terms of discharge, not flood recurrence interval, because the flood frequency analysis probably cannot be improved much. The primary problem is that for large recurrence intervals, flood frequency functions depend heavily on an empirically determined skew and, for the Clark Fork, there is really no source of data to determine the skew accurately. Ultimately hydrologists will have to turn to meteorology for assistance in flood prediction, but this interfacial field is still in its infancy, and few useful predictive, process-based models are yet available for dealing with large floods. Fortunately in healthy riparian systems, as shown above, the dominant dependence of skin friction is on shrub spacing, not on flood discharge.

Once the riparian shrubs become sparse, then the nature of the flood-plain stability problem changes. The hydraulics becomes dependent on stage rather than shrub density and the flood plain must be protected by herbaceous vegetation. A process-based model for momentum extraction by grasses is under development by Smith, but process-based models for shrub-covered flood plains are much easier to formulate in careful fluid-mechanical terms than are ones for grass-covered systems. Until predictive, process-based models for extracting momentum by form drag and skin friction on herbaceous plants have been developed, the effects of herbaceous surface cover probably will have to remain in the domain of enhanced threshold shear stresses, and not in the domain of momentum extraction. This forces the critical shear stress problem to remain an empirical one. For the Clark Fork through the Deer Lodge Valley, this analysis leads to the conclusion that little further progress is likely to be made on the stability issue until much additional field data are collected, including hundreds of inverted flume measurements of threshold shear stress for flood-plain erosion. Any inverted flume measurements will have to be carefully analyzed to develop methods for generalizing the results in terms of flood-plain topography, sedimentology, vegetation and land use characteristics, flood-plain botany, and land-use history. Without this sort of analysis, the flood plains will remain undersampled, and a purely statistical approach will lead to biased results. A handful of threshold shear stress measurements made out of context using an inverted flume will not lead to much gain over the field procedures employed to procure the crude measurements described in this report. The most immediate need for examining the vulnerability of the Clark Fork to large floods is a careful, comprehensive inverted flume investigation focusing on the effects of non-irrigated grasses in protecting the flood-plain soil from erosion (that is, increasing the threshold shear stress for erosion).

The results presented in this report indicate that the flood plain of the Clark Fork in the upper Deer Lodge Valley has become highly vulnerable to surface erosion during large floods and that this vulnerability has reached the point where catastrophic geomorphic change can occur during a 100-year-recurrence-interval flood and probably during a prolonged 30-year-recurrence-interval flood. The present fluvial system cannot withstand skin frictions of several Newtons per square meter.

CONCLUSIONS

The predictive, process-based model for flow over shrub-covered flood plains developed for this investigation and tested using data for a geomorphology-altering flood on East Plum Creek, Colorado, provides an essential tool for the evaluation of flood plain response to large floods. Its greatest attribute, however, is its ability to provide continuous and accurate variations of the key hydraulic parameters as functions of stem density for a variety of flood-plain shrubs. This is essential because once stem density is high enough for continuous canopy coverage over flood-plain tabs, it replaces stage as the primary independent hydraulic variable. Under these conditions, flow velocity and boundary shear stress are essentially independent of flood discharge and flow depth, making catastrophic geomorphic change virtually impossible until the shrubs are overtopped by the flow, and perhaps even for overbank flow depths that are well in excess of the average shrub height (Smith, 2001).

Only a model that can treat continuous variations in stem density as well as in discharge can provide a satisfactory understanding of the interplay of stem density and stage on flood plain tabs that are only partially covered with shrubs. Without such a model it would be impossible to understand flood plains like that of the Clark Fork in the Deer Lodge Valley.

At present, the flood plain of the Clark Fork in the Deer Lodge Valley is susceptible to catastrophic geomorphic change during a moderate flood, owing to its impoverished riparian shrub flora. In contrast, historical data indicate that it was heavily vegetated with shrubs and aspen trees in the mid-nineteenth century, and our calculations for the 1908 flood indicate that the flood plain of the Clark Fork probably was extremely resistant to geomorphic alteration at that time. Currently, a 100-year flood and possibly a prolonged 30-year flood will cause the Clark Fork through the Deer Lodge Valley to undergo catastrophic geomorphic change from a single-threaded, meandering river to a broad, shallow, multi-threaded one. Even smaller floods will channelize the flood plain tabs and transport large amounts of contaminated sediment to the river. The solution to these problems is to re-vegetate the entire meander belt of the Clark Fork through the Deer Lodge Valley with large shrubs.

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