Magnitudes and implications of peak discharges from glacial Lake Missoula

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ABSTRACT

New field evidence and discharge calculation procedures provide new estimates of maximum late Pleistocene glacial Lake Missoula flood discharges for two important reaches along the flood route. Within the Spokane Valley, near the point of release, the peak discharge probably exceeded 17 ± 3 million $m^3 \cdot sec^{-1}$. Downstream at Wallula Gap, a major point of flow convergence, peak discharge was about 10 ± 2.5 million $m^3 \cdot sec^{-1}$. Flow duration was on the order of several days. These are the largest known terrestrial fresh-water flows.

Consideration of these discharge values constrains models for the failure of glacial Lake Missoula. The maximum discharges estimated here are larger than theoretical and empirical predictions of maximum subglacial jökulhlaup-style releases for Lake Missoula. We postulate, consistent with geological relations in the glacial Lake Missoula basin and in the Channeled Scabland, that the largest late Wisconsinan Missoula Flood resulted from a cataclysmic failure of the impounding ice dam of glacial Lake Missoula. This large release may have been the result of a complete rupture of the ice dam. Subsequent multiple flows of lesser magnitude may have resulted from repeated subglacial releases from the lake.

INTRODUCTION

Pleistocene discharges from ice-dammed glacial Lake Missoula produced a suite of spectacular flood features along multiple flow pathways between western Montana and the Pacific Ocean (Fig. 1). Since the first decades of this century, scientific controversy has continued regarding the interpretation of these features and deposits, including the origin, chronology, magnitude, frequency, and lake-release mechanisms associated with possible repeated emptying of Lake Missoula. Baker and Bunker (1985) and Waitt (1985) reviewed the development of hypotheses regarding these issues. Here we report results of field studies and hydraulic reconstructions of maximum discharges associated with the Missoula floods at two important sites, and we discuss the implications of these results to hypotheses proposed and models advanced concerning the nature of Lake Missoula flooding.

METHODS

The primary purpose of this study is to provide limits, based on physical geologic evidence of flood stages, on the maximum discharges that issued from glacial Lake Missoula. Field studies were conducted in the Wallula Gap reach where there is little reported information regarding maximum flood stages. Further studies were also performed in the Rathdrum Prairie and Spokane Valley, where existing information (Baker, 1973) was re-evaluated and augmented. The evidence for maximum stages is combined with step-backwater flow modeling to estimate peak discharges through each of the reaches.

Geologic Evidence of Maximum Flood Stages

Altitudes of flood landforms and deposits provide minimum values for the maximum flow stages, because the depth of water above the flood feature is unknown. In contrast, nearby nonflooded terrains indicate the maximum possible local altitude of peak flow stage, because the actual water surface must have been lower. Evidence of flooding includes divide crossings, "scabland" topography, scarps cut into loess, ice-rafted erratics, and accumulations of flood-transported sediment (Baker, 1973, p. 13–14). Divide crossings, loess scarps, and scabland topography are most useful as paleostage indicators in reaches of relatively steep flow gradient and can generally be recognized on topographic maps and aerial photographs. Ice-rafted erratics, exotic rock types beyond the limits of glaciation, are the most consistent indicators of maximum flood stages in ponded areas. Field studies were undertaken in both study reaches to confirm map and photo interpretations, to evaluate existing data, and to determine the narrowest possible altitude range of maximum flow stages. Altitudes of stage evidence were determined by mapping onto U.S. Geological Survey 7.5-minute topographic maps. Altitudes were determined conservatively for features identified in this study and those of Baker (1973); that is, positions inferred to be flooded were assigned altitudes on the basis of the highest contour line determined to be below that position. Therefore, the reported altitude of maximum flow stage is as much as one contour interval (6 to 12 m) below the maximum flow stage. Likewise, reported altitudes of features inferred not to have been inundated are as much as one contour interval above the inferred maximum flow stage.

Discharge Calculations

Step-Backwater Modeling. To estimate discharges associated with the field evidence of maximum stages, we used the U.S. Army Corps of Engineers' HEC-2 Water Surface Profiles computer program (Feldman, 1981; Hydrologic Engineering Center, 1985). This algorithm computes energy-balanced water-surface profiles for steady, gradually varied flow on the basis of the one-dimensional energy equation (a form of the Bernoulli Equation) with the step-backwater method of profile computation (Chow, 1959, p. 265–268). For flow conditions (discharge, stage, and channel geometry) specified at one cross section, stage is calculated for an adjacent cross section of known geometry and reach length by conserving mechanical energy and accounting for estimated flow-energy losses associated with roughness and channel expansions and contractions. Calculated

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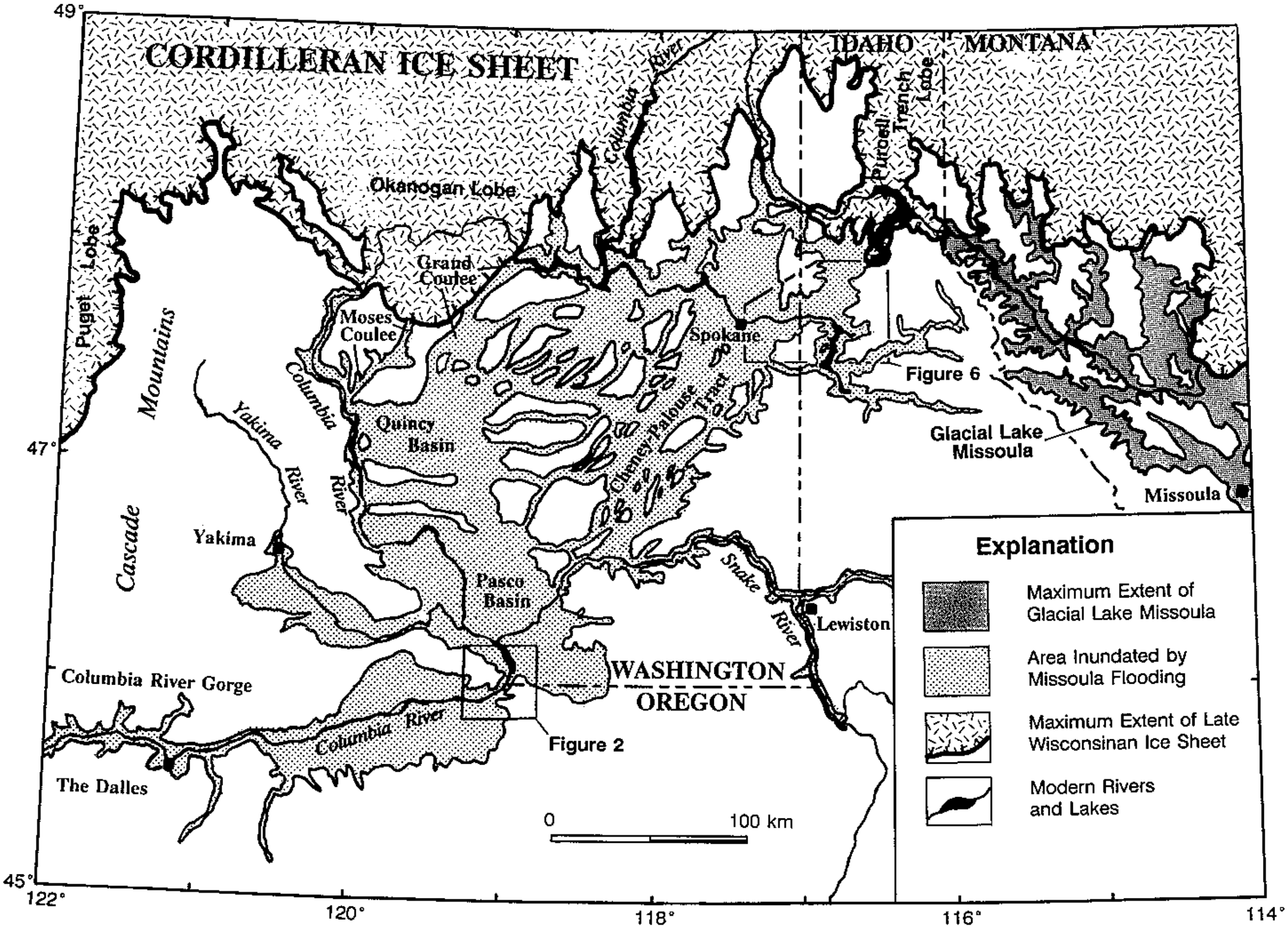

Figure 1. Major features associated with Missoula flooding in eastern Washington, northern Idaho, and western Montana. Geology from Baker (1973), Waitt and Thorson (1983), and Atwater (1986).

conditions at the second cross section are then taken as "known" for the next incremental subreach. In this manner, the stage can be determined for each cross section, and the water-surface profile for a specified discharge can be determined for an indefinite length of channel. For subcritical flow, profiles are calculated successively in the upstream direction.

Application for Paleohydrologic Analysis. For geologic applications, where the object is to estimate the discharge associated with paleostage evidence, water-surface profiles are computed for different combinations of discharge values and energy loss coefficients until a calculated profile is achieved that most closely matches the water-surface profile indicated by the maximum stage evidence (Ely and Baker, 1985; Craig and Hanson, 1985; Jarrett and Malde, 1987; O'Connor and Webb, 1988). This method of flow retrodiction probably represents an improvement over most previous discharge calculations in the Channeled Scabland for the following reasons. (1) It does not require the assumption of uniform flow in a highly nonuniform environment as does simple application of the Chezy Equation (Bretz, 1925). (2) In contrast to the slope-area method (Dalrymple and Benson, 1967; Baker, 1973, p. 17–18), computed step-backwater profiles are not dependent on the precise altitudes of the stage evidence; therefore, discharge estimates are not particularly sensitive to local uncertainties in actual stage altitudes.

Assumptions intrinsic to step-backwater analyses are that the flow is steady, gradually varied, and one-dimensional. The study reaches were selected partly to minimize deviations from these assumptions; the following reach descriptions discuss the suitability of each site for step-backwater calculations. Additional assumptions are introduced for flows at this scale and for flows that occurred in the past. Importantly, it is uncertain whether using the empirical Manning Equation to evaluate local friction slope (intrinsic to most indirect methods of discharge estimation) is appropriate for flows that are several orders of magnitude larger than any flow that has been directly measured. We know of no way to evaluate this assumption except to note that the Manning Equation has proven to be appropriate for flows over several (albeit much smaller) orders of magnitude. Another assumption is that the existing measured cross sections closely approximate the channel geometry at peak flood stages; that is, there hasn't been significant post-peak downcutting, filling, or valley widening. This assumption must also be evaluated at each site, as described below.

For both reaches, cross sections were measured from U.S. Geological Survey topographic maps and were positioned and spaced (1) to characterize downstream changes in channel geometry and (2) to limit changes in flow conditions between sections. Manning roughness coefficients were assigned for appropriate portions of each cross section according to the estimated degree of flow efficiency across the surface. Although there is no rational method available to determine roughness coefficients for flows of this scale, the chosen values were selected on the basis of systematic criteria and experience with modeling similar flows (O'Connor, 1990). Sensitivity tests were performed to estimate the effect of uncertainties in roughness coefficients on the resultant water-surface profiles. Flow energy losses resulting from flow expansions and contractions (form roughness) are calculated as being proportional to the change in kinetic energy from one cross section to another (Hydrologic Engineering Center, 1985). Different values of these coefficients were also used to test their effect on the calculated water-surface profiles, as described below.

STUDY REACHES

Wallula Gap

Wallula Gap is a narrow, 1.5-km-wide, 250-m-deep gorge through the Horse Heaven Hills anticline in south-central Washington (Fig. 2). The constriction is the southern boundary of the Pasco Basin, a basin that has accumulated as much as 300 m of late Tertiary and Quaternary

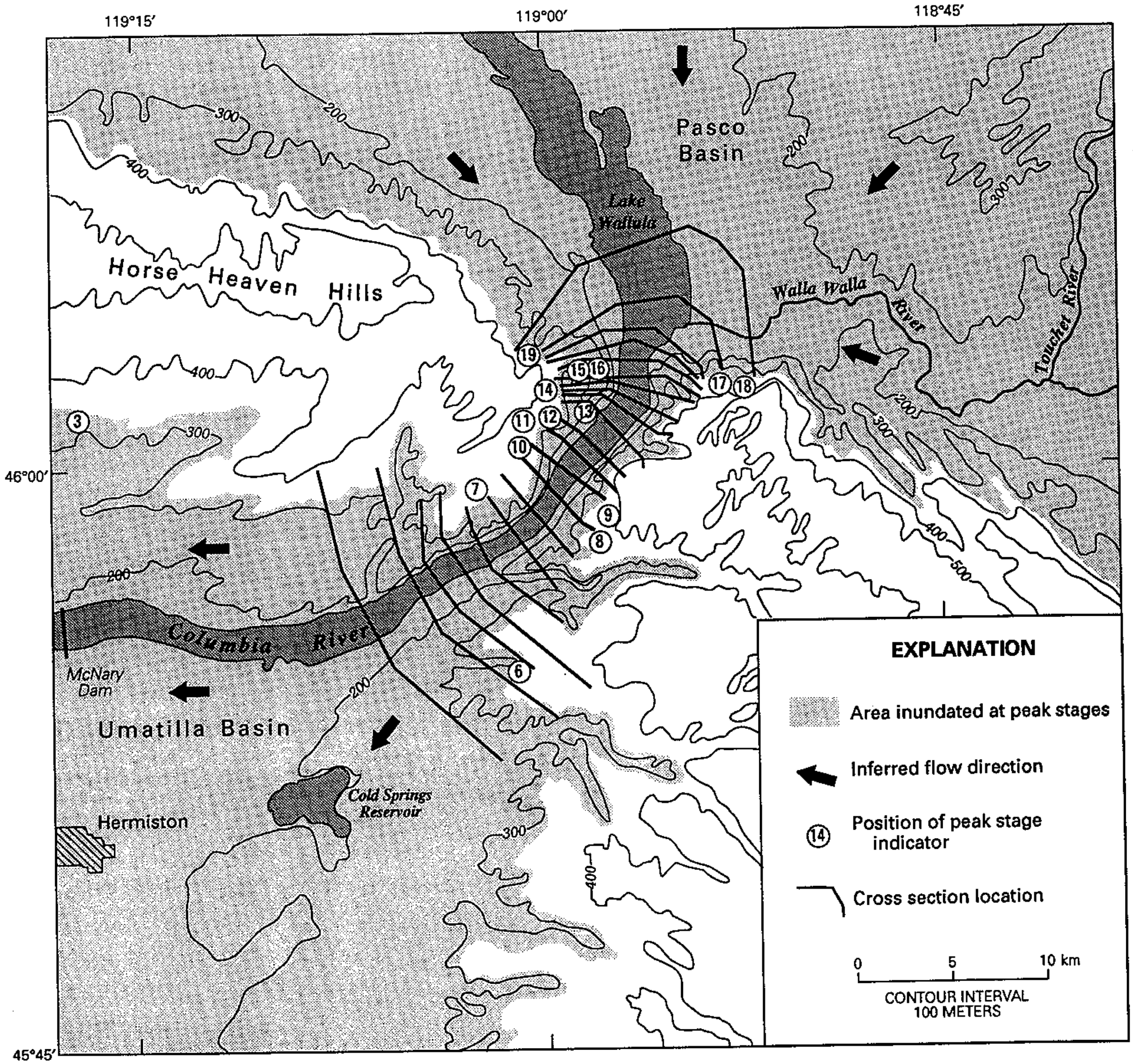

Figure 2. Topography, locations of evidence for maximum stages (Table 2), cross-section locations, and approximate area of inundation for the Wallula Gap reach at maximum flood stages. Note that some of the evidence for maximum stages is outside the map area.

TABLE 1. DISCHARGE ESTIMATES FOR THE MISSOULA FLOODS

Location	Discharge (10^6 $m^3 \cdot sec^{-1}$)	Reference	Calculation procedure	Comments
Wallula Gap	1.9	Bretz (1925, p. 257–258)	Chezy equation	Calculation performed by D. F. Higgins. Utilized cross-section areas much smaller than those associated with present understanding of maximum flood stages.
Wallula Gap	9.1	Baker (1973)	Contracted opening equation	Assumed a maximum ponding level of 350 m in the Pasco Basin. Assumed critical flow through Wallula Gap.
Wallula Gap	12.5	Craig and Hanson (1985, p. 52–53)	Step backwater	Assumed a maximum ponding level of 350 m in the Pasco Basin. Assumed critical flow through Wallula Gap.
Wallula Gap	10	This study	Step backwater	Profile constraints include 335-m ponding level in the Umatilla Basin and a 370- to 385-m ponding level in the Pasco Basin. Flow not assumed to be critical in the constriction.
Sentinel Gap	9.5	Craig and Hanson (1985, p. 53–66)	Step backwater	Downstream stage estimated on the basis of an extrapolated water-surface slope from Wallula Gap. Upstream ponding level assumed to be 366 m. Discharge estimate applies only to water that followed the course of the Columbia River (does not include flow in the eastern scabland tracts).
Moses Coulee	1.1–9.8	Hanson (1970, p. 59)	Manning equation	Limiting values considering estimated maximum and minimum cross-section areas at peak stages for two different cross sections.
Grand Coulee (near Soap Lake)	4.5	Baker (1973, Pl. 1)	Contracted opening equation	Assumed critical flow through the constriction at Soap Lake. Upstream water surface estimated from geologic evidence of maximum stage.
Grand Coulee (near Steamboat Rock)	0.13	Atwater (1986)	Sediment transport competence criteria	Based on extrapolation of a local velocity estimate across the upper Grand Coulee. Discharge associated with flood deposits within glacial Lake Columbia sediments.
Cheney-Palouse scabland tract	8.1	Baker (1973)	Slope-area method	Based on water-surface slope defined by geologic evidence of maximum stages.
Spokane Valley–Rathdrum Prairie	21.3	Baker (1973)	Slope-area method	Based on water-surface slope defined by geologic evidence of maximum stages.
Spokane Valley-Rathdrum Prairie	17	This study	Step backwater	Based on highwater evidence of Baker (1973) and additional field observations.
Lake Missoula breakout point	13–15.3 2.6–3.5	Clarke and others (1984)	Simulation model	Model simulating subglacial release of Lake Missoula from the maximum lake level. High values are for conditions of no downstream backwater and immediate conveyance of lake water to the release point. Low values are for conditions of downstream backwater and delayed conveyance associated with the complex lake geometry.
Lake Missoula breakout point	2.1	Beget (1986)	Empirical relationship	Extrapolated relationship between ice-dammed lake volumes and associated jökulhlaup discharges to predict discharge associated with the release of the maximum volume of Lake Missoula.
Eddy Narrows	10.9	Pardee (1942)	Chezy and Manning equations	Used high-level flood features to estimate cross-section area. Used present channel slope as a proxy for the water surface (and energy slope).

fluvial and lacustrine sediments (Newcomb, 1958; Newcomb and others, 1972; Waitt and Swanson, 1987), including glacial Lake Missoula flood deposits. Wallula Gap is an important site for discharge estimation, because it is the most upstream point of convergence for the multitude of flow routes that enter the Pasco Basin (Fig. 1). Previous discharge estimates at Wallula Gap were calculated by Bretz (1925), Baker (1973), and Craig and Hanson (1985) (Table 1).

Wallula Gap is particularly suitable for flow reconstruction by the step-backwater method. The steady-flow assumption is probably satisfied because of hydraulic damming at the constriction and consequent upstream storage and flow attenuation within the Pasco Basin. A radiometrically dated basalt flow close to present river level near The Dalles indicates that the Columbia River has been near its present altitude for at least 900,000 yr (Bjornstad and others, 1991), and flood-related downcutting was minor. Because the constriction is formed in resistant rocks, flood-related widening probably does not affect discharge estimates. As noted by Bretz (1969), the existence at river level of channel-margin flood sediments associated with maximum stages indicates that there has not been major channel widening after the passing of maximum discharges. Bretz (1969, p. 535) further stated, "[The] gorge [immediately] downstream from the gap itself therefore was deep as it is today. . . ."

Previous workers in the area have defined, on the basis of ice-rafted erratics, maximum flood stages in the Pasco Basin (370–385 m; Bjornstad and others, 1991) and in the Umatilla Basin at the downstream end of the study reach (335 m;

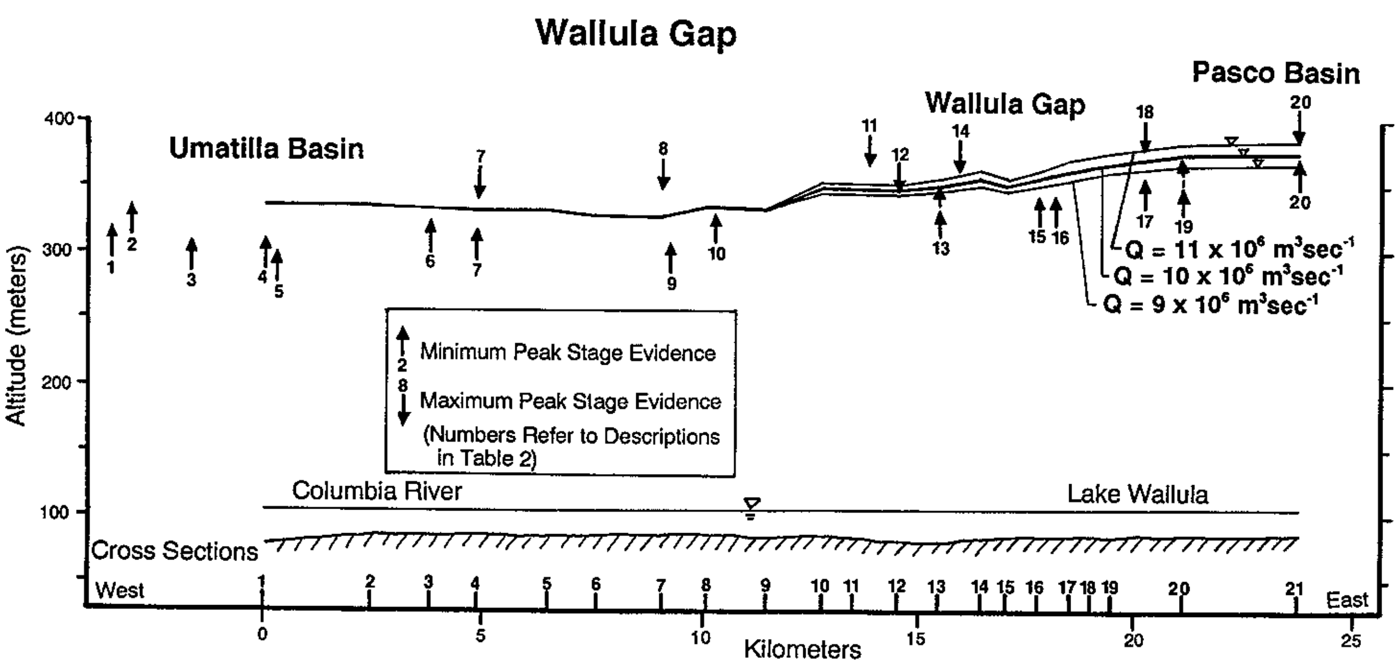

TABLE 2. MAXIMUM STAGE EVIDENCE FOR THE WALLULA GAP REACH

Number	Location (and USGS 7.5' quadrangle)	Altitude (m)	Type	Reference	Comments
1	W½ S26 T. 2 N. R. 25 E. Strawberry Canyon SW (OR)	320	Ice-rafted erratic	Allison (1933, p. 689)	Altitude as reported by Allison.
2	Umatilla Basin	335–350	Fine-grained deposits and small erratics	Allison (1933, p. 685)	Upper limit of "thin cover of the pebbly and bouldery glacial silt." Lower altitude estimate indicated in Figure 4. (We have found such evidence of flooding only to an altitude of 335 m.)
3	SW¼ SW¼ S17 T. 6 N. R. 28 E. Umatilla (WA-OR)	305	Fluvial sands and silts	This study	Felsic boulders (ice-rafted?) at similar altitudes on adjacent hillslopes.
4	NW¼ S8 T. 2 N. R. 30 E. Nolin (OR)	310	Ice-rafted erratic	Allison (1933, p. 685)	Altitude as reported by Allison.
5	NW¼ NE¼ S17 T. 2 N. R. 30 E. Nolin (OR)	300	Ice-rafted erratics	This study	
6	W½ S13 T. 5 N. R. 30 E. Juniper (OR-WA)	323	Scabland topography	This study	Stripped basalt and eroded loess.
7	NE¼ NE¼ S28 T. 6 N. R. 30 E. Juniper (OR-WA)	315	Base of loess scarp	This study	Top of loess scarp is at 342 m and provides upper limit for maximum flood stage.
8	SW¼ SW¼ S30 T. 6 N. R. 31 E. Juniper Canyon (OR-WA)	349	Top of loess scarp (upper limit)	This study	Non-flooded loess-covered terrain.
9	NW¼ SW¼ S30 T. 6 N. R. 31 E. Juniper Canyon (OR-WA)	305	Divide crossing	This study	
10	NW¼ NE¼ S13 T. 6 N. R. 30 E. Nine Canyon (WA)	329	Divide crossing	This study	
11	SW¼ SW¼ S6 T. 6 N. R. 30 E. Wallula (WA-OR)	366	Divide not crossed (upper limit)	This study	
12	NE¼ SE¼ S6 T. 6 N. R. 31 E. Wallula (WA-OR)	343	Uneroded loess-covered mesa (upper limit)	This study	
13	SW¼ NE¼ S5 T. 6 N. R. 31 E. Wallula (WA-OR)	336–349	Scabland topography	This study	Boulder deposits to an altitude of 336 m.
14	NW¼ NW¼ S5 T. 7 N. R. 31 E. Wallula (WA-OR)	360	Uneroded loess-covered mesa (upper limit)	This study	
15	SW¼ SW¼ S32 T. 7 N. R. 31 E. Wallula (WA-OR)	342	Divide crossing	This study	
16	SW¼ NE¼ S32 T. 7 N. R. 31 E. Wallula (WA-OR)	342	Ice-rafted erratics	This study	
17	SW¼ SW¼ S35 T. 7 N. R. 31 E. Wallula (WA-OR)	355	Boulder bar and divide crossing	This study	Altitude is of boulder bar.
18	SW¼ SW¼ S35 T. 7 N. R. 31 E. Wallula (WA-OR)	378	Top of loess scarp (upper limit)	This study	Immediately above 17.
19	SW¼ SW¼ S30 T. 7 N. R. 31 E. Wallula (WA-OR)	348–366	Divide crossing	Bretz (1969); this study	Bottom of divide is at 348 m. Bretz postulated a maximum flood stage of 366 m.
20	Pasco Basin	370–385	Ice-rafted erratics	Bjornstad and others (1991)	Maximum ponding level in the Pasco Basin.

Allison, 1933). These values are consistent with our observations of maximum stages inferred from the field evidence within Wallula Gap (Table 2).

Wallula Gap Modeling and Results. Flow profiles were calculated through 21 cross sections over a length of 30 km (Figs. 2 and 3). The maximum flow stage was assumed to be 335 m at the downstream cross section. Manning roughness coefficients were assigned as follows: 0.04 over the valley floor; 0.05 over the steep valley margins; 0.1 for flow over the high uplands. These values are larger than roughness coefficients typically assigned to more normal scale floods. Following Jarrett and Malde (1987), in their application of step-backwater modeling techniques to the Bonneville Flood, we used contraction-expansion coefficients of 0.0 and 0.5. With these parameters, a discharge of 10 million $m^3 \cdot sec^{-1}$ provides the best fit, by eye, to the geologic evidence of maximum Missoula Flood stages through Wallula Gap and results in a ponding level of 375 m in the Pasco Basin (Fig. 3). Utilizing the same energy-loss coefficients, an 11.2 million $m^3 \cdot sec^{-1}$ discharge is required to achieve the 385-m ponding elevation reported by Bjornstad and others (1991) as an upper estimate of maximum ponding levels in the Pasco Basin. In our estimation, the 10 million $m^3 \cdot sec^{-1}$ estimate provides a better fit to the maximum stage evidence through the constriction and serves as a minimum estimate for the energy-loss coefficients specified above. For comparison, the maximum measured discharge for the Columbia River was 35,000 $m^3 \cdot sec^{-1}$ at The Dalles during the 1894 flood.

The 10 million $m^3 \cdot sec^{-1}$ discharge estimate is minimally affected by uncertainties in the Manning roughness coefficients. Varying Manning roughness coefficients by ±25% requires adjustments of less than 5% to the prescribed discharge value to maintain a 375-m stage in Pasco Basin (Fig. 4). Modeling results are more sensitive to the expansion and contraction coefficients. Increasing these coefficients to the values of 0.3 and 0.7, values larger than typically assigned for estimating this form of energy loss, the required discharge to achieve a 375-m ponding level decreases 20% to about 8 million $m^3 \cdot sec^{-1}$. Considering uncertainties in the maximum stages achieved in the Pasco Basin and uncertainties in assigning energy-loss coefficients, discharge values between 7.5 and 12 million $m^3 \cdot sec^{-1}$ are consistent with maximum stage evidence at the upstream and downstream ends of the reaches (Fig. 4).

Figure 3. Step-backwater calculated water-surface profiles for the Wallula Gap reach. A 10 million $m^3 \cdot sec^{-1}$ discharge results in a calculated profile that matches most of the highwater evidence along the reach and results in a ponding level in the Pasco Basin consistent with the altitudes of the highest ice-rafted erratics.

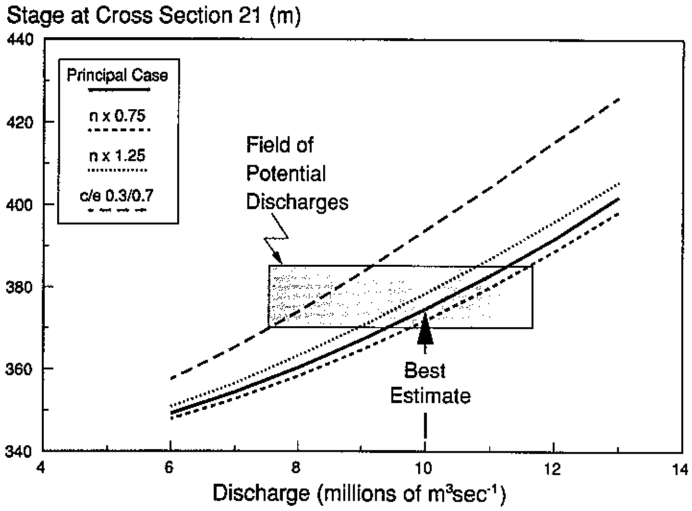

Figure 4. Rating curves for calculated stages in the Pasco Basin (cross section 21 of Fig. 3) for different values of energy-loss coefficients and for a ponding level of 335 m in the Umatilla Basin. The effects of varying Manning roughness coefficients are minimal. The principal case was calculated with contraction and expansion energy-loss coefficients of 0.0 and 0.5. Using larger coefficients reduces the discharge required to achieve a 375-m stage in the Pasco Basin to about 8 million $m^3 \cdot sec^{-1}$. Considering the highwater evidence identified in this study and the uncertainty of maximum Pasco Basin stages and reasonable ranges of the energy-loss coefficients, the maximum Missoula Flood discharge through Wallula Gap was between 7.5 and 12 million $m^3 \cdot sec^{-1}$.

The 10 million $m^3 \cdot sec^{-1}$ discharge proposed here is somewhat higher than the discharge estimates calculated in the studies by Baker (1973, p. 17–22) and Bretz (1925, p. 257–258) but lower than the 12.5 million $m^3 \cdot sec^{-1}$ estimate of Craig and Hanson (1985, p. 52–53) (Table 1) (see also Craig, 1987, p. 324). Previous discharge calculations were postulated on assumed ponding levels in the Pasco Basin lower than the maximum level of 370–385 m presently recognized (Bjornstad and others, 1991). Furthermore, the analyses of Baker (1973) and Craig and Hanson (1985) assumed critical-flow conditions within the Wallula Gap constriction. Even though it is likely that critical-flow conditions were achieved during the waxing phase of the flood, it seems that during maximum stages, ponding in the Umatilla Basin to an altitude of 335 m impeded flow through the Wallula Gap. On the basis of the altitudes of a variety of flood features described by Allison (1933) and Newcomb (1969), it appears that constrictions 250 km downstream in the Columbia River Gorge (Fig. 1) hydraulically impounded flow to the extent that Wallula Gap did not act as a free constriction. If, however, the flood wave traveled at such a speed that downstream backwater effects did not choke the constriction at Wallula Gap until after maximum stages were achieved, then the assumption of critical flow (Baker, 1973; Craig and Hanson, 1985) may be valid. For critical flow conditions within the constriction and a ponding level of 375 m in the Pasco Basin, a discharge of 14 million $m^3 \cdot sec^{-1}$ could have funneled through Wallula Gap with the energy-loss coefficients specified above (Fig. 5). Because critical flow at Wallula Gap is the most efficient flow condition, this discharge estimate serves as a maximum value, given our present understanding of maximum ponding levels in the Pasco Basin.

Spokane Valley and Rathdrum Prairie

The Spokane Valley and the contiguous Rathdrum Prairie is a large, steep-walled valley extending southwest of the southern end of Pend Oreille Lake, Idaho, to Spokane, Washington (Fig. 6). In the most constricted location (at cross section 5 of Fig. 6), the valley narrows to a width of 6 km before opening up into the Spokane Valley. As much as 150 m of Pleistocene glaciofluvial and flood deposits underlie the valley floor (Bolke and Vaccaro, 1981). Some of these sediments have been molded into flood forms, including bars and trains of "giant current ripples" (Baker, 1973, p. 48–64). Eddy bars, blocking pre-existing valleys, impound many of the lakes marginal to the study reach (Kiver and Stradling, 1982, 1989).

This reach is adjacent to the area of release for the Missoula floods near the south end of Pend Oreille Lake, and the valley served as the primary conduit of flood waters exiting Lake Missoula. As such, discharges within this reach were probably the largest with respect to the entire Channeled Scabland complex. Previous Missoula flood discharge estimates pertinent to this reach include the slope-area calculation of Baker (1973, p. 14–22) based on field evidence of maximum stages, and the discharge calculations for the maximum Lake Missoula release by

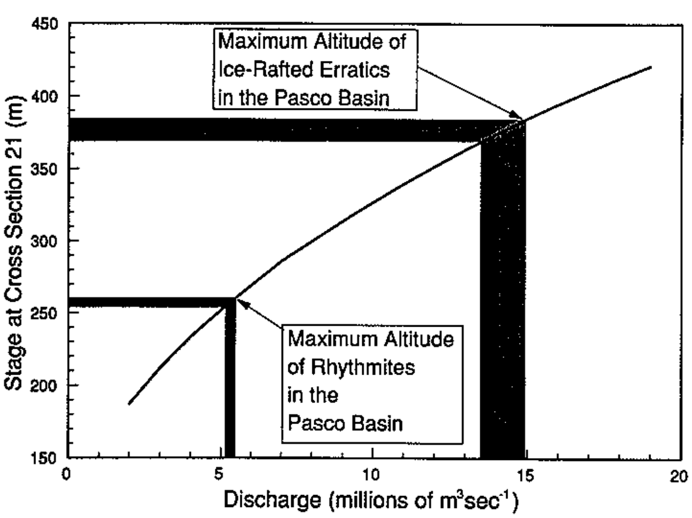

Figure 5. A calculated stage-discharge rating curve for the Pasco Basin associated with critical flow conditions within Wallula Gap and the relationship of this curve to the maximum reported altitudes of ice-rafted erratics and rhythmite deposits in the Pasco Basin. The rating curve was calculated by assuming critical flow at cross section 16 (Fig. 3) and computing the upstream profiles with the same (upstream) cross sections used for calculating the water-surface profiles for the entire reach. Assigned energy-loss coefficients were the same as for the principal case of Figure 4. This rating curve should closely approximate maximum possible discharge values associated with the corresponding stages in the Pasco Basin. Craig (1987) provided a similar rating curve based on a different set of cross sections (Craig and Hanson, 1985).

Figure 6. Topography, locations of maximum stage evidence (Table 3), cross-section locations, and approximate area of inundation for the Spokane Valley reach. Note that some of the evidence for maximum stages is outside the map area.

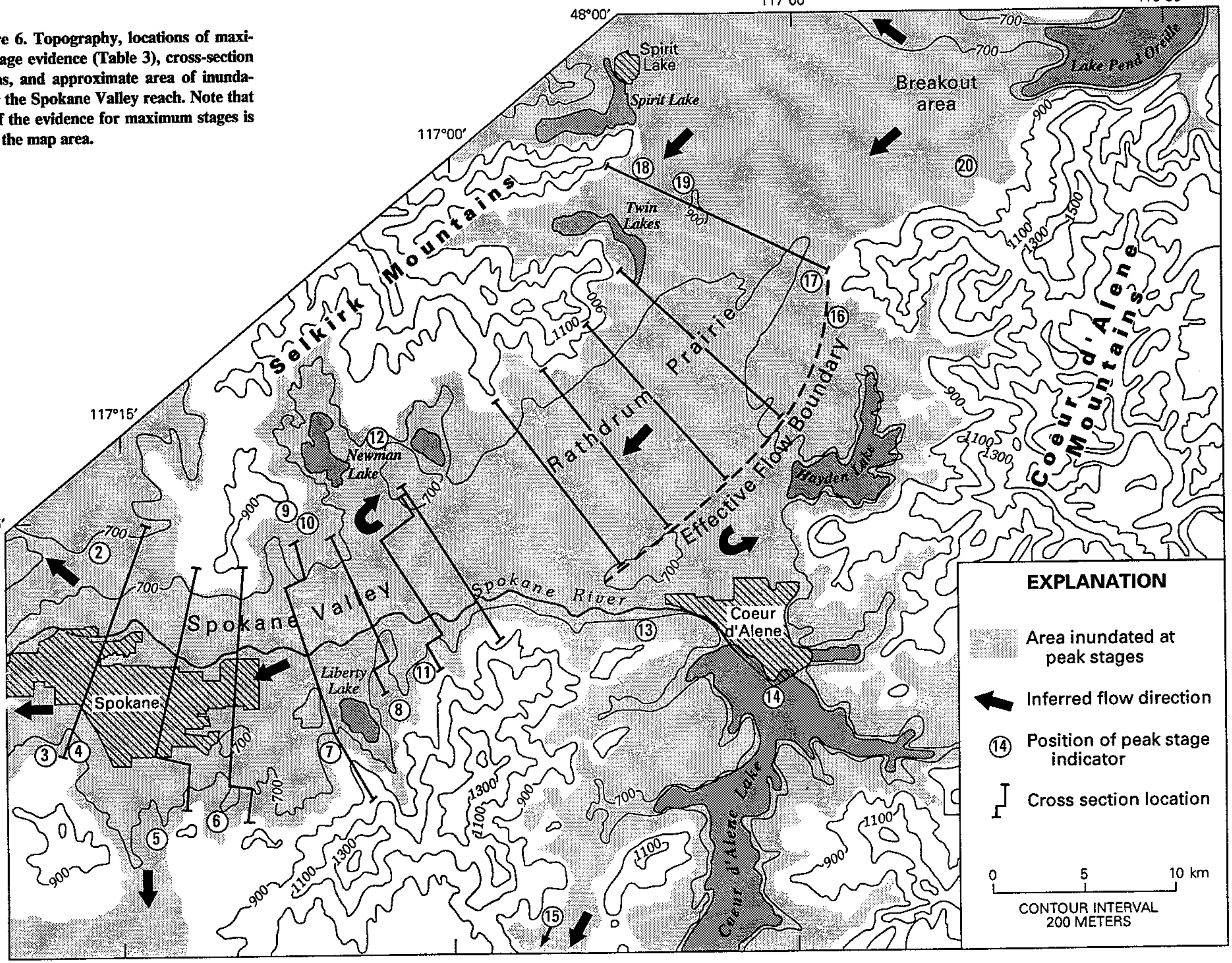

TABLE 3. MAXIMUM STAGE EVIDENCE FOR THE SPOKANE VALLEY-RATHDRUM PRAIRIE REACH

Number	Location (and USGS 7.5' quadrangle)	Altitude (m)	Type	Reference	Comments
1	Sanpoil River Valley	750	Ice-rafted erratics	Atwater (1986 p. 7)	Maximum, "flood-swollen" level of Glacial Lake Columbia.
2	W½ S28 T. 26 N. R. 44 E. Spokane NE (WA)	735	Divide crossing	This study	Flood deposits immediately downstream at 730 m.
3	NW¼ SW¼ S30 T. 25 N. R. 44 E. Spokane NE (WA)	726	Eddy bar	This study	
4	NE¼ SE¼ S30 T. 25 N. R. 44 E. Spokane NE (WA)	756	Scabland topography	This study	
5	W½ SW¼ S10 T. 24 N. R. 44 E. Freeman (WA)	743	Divide crossing	Baker (1973, no. 61); this study	
6	SW¼ NE¼ S12 T. 24 N. R. 44 E. Freeman (WA)	744	Ice-rafted erratic	This study	
7	SE¼ S27 T. 25 N. R. 45 E. Liberty Lake (WA)	762–775	Divide crossing	Baker (1973, no. 63); this study	Bottom of the divide at 762 m. Maximum altitude of rounded boulders is 775 m.
8	SE¼ SE¼ S24 T. 25 N. R. 45 E. Liberty Lake (WA-ID)	781	Divide crossing	Baker (1973, no. 64); this study	Small gravel delta downstream of divide.
9	SW¼ SW¼ S16 T. 26 N. R. 45 E. Greenacres (WA)	762	Divide crossing	This study	
10	NW¼ SE¼ S21 T. 26 N. R. 45 E. Liberty Lake (WA-ID)	811	Divide not crossed (upper limit)	This study	
11	NE¼ NW¼ S19 T. 50 N. R. 6 W. Liberty Lake (WA-ID)	830	Divide not crossed (upper limit)	Baker (1973, no. 65); this study	
12	NE½ S36 T. 27 N. R. 45 E. Newman Lake (WA-ID)	799	Divide crossing	This study	
13	SW¼ SE¼ S8 T. 50 N. R. 4 W. Coeur D'Alene (ID)	778	Scabland topography	This study	
14	Coeur D'Alene Lake Basin	813	Ice-rafted erratics	Richmond and others (1965)	Ice-rafted erratics located on "adjacent mountain slopes" flanking Coeur d'Alene Lake.
15	E½ S32 T. 49 N. R. 6 W. Setters (ID-WA)	787	Divide crossing	Richmond and others (1965)	Flow depth unknown.
16	NE¼ NW¼ S32 T. 52 N. R. 3 W. Hayden Lake (ID)	799	Divide crossing	This study	
17	SW¼ NE¼ S17 T. 52 N. R. 3 W. Hayden Lake (ID)	793	Scabland topography	Baker (1973, no. 71)	
18	N½ S29 T. 53 N. R. 4 W. Spirit Lake E. (ID)	799	Flood bar	This study	Large bar with boulders larger than 2 m.
19	SW¼ SW¼ S35 T. 53 N. R. 4 W. Spirit Lake E. (ID)	817	Scabland topography	This study	Upper extent of eroded topography in the vicinity of Round Mountain. Bar downstream is at an altitude greater than 793 m.
20	SE¼ NE¼ S30 T. 53 N. R. 2 W. Bayview (ID)	824	Gravel bar	This study	

Clarke and others (1984) and Beget (1986) concluded from theoretical and empirical analyses, respectively, of flood releases from ice-dammed lakes (Table 1).

Evidence of Maximum Flood Stages. Baker (1973, p. 66–67) reported several sites of maximum stage evidence along this reach, of which several were first observed by P. L. Weis (personal commun. with V. R. Baker, 1969). Many of these were revisited for this study, and their altitudes were re-evaluated. Richmond and others (1965) reported evidence of maximum flood stages along this reach, as did Kiver and Stradling (1989). Additional observations from our field studies have also been incorporated into the analysis (Table 3; Fig. 7). Altitudes of maximum stage evidence at the downstream end of the reach approach 750 m, which is similar to the altitude determined by Atwater (1986, p. 7; 1987) for the maximum flood-swollen stage of glacial Lake Columbia. At the upstream end of the study reach, gravel bars and ice-rafted erratics achieve altitudes near 820 m.

Discharge Calculation. The ability to produce reliable discharge estimates here is tempered by the complex flow conditions that must have existed through this reach. Initial flow was assuredly unsteady, perhaps manifested as a dynamic wave (Craig, 1987). Unsteadiness in the flow can lead to discharge overestimation in that the flood features related to the calculated steady-flow water-surface profiles may have been emplaced by the crest of a transient wave moving through the study reach. Many of the high flood features are large constructional landforms, however, that we infer to have required formation times longer than that associated with passage of a dynamic flood wave. Key such features include the eddy bar southeast of Spokane at 726 m (stage indicator no. 3 on Table 3) near the downstream end of the study reach and, at the upstream end, the large bars near Spirit Lake and Round Mountain (stage indicators no. 18 and no. 19) at 799 m. Furthermore, the tortuous geometry of sub-basins and constrictions within glacial Lake Missoula would have regulated outflow from the lake into this reach (Pardee, 1942; Clarke and others, 1984; Craig, 1987) after the initial floodwave had passed. We propose that the well-developed flood features preserved along the study reach represent flow conditions that approached steady. Kiver and Stradling (1989) reported evidence of flood erosion much higher than the stage indicators used in this study. Perhaps this high evidence records the transient effects of a dynamic wave.

Another complicating factor in determining the total glacial Lake Missoula discharge in this reach is that multiple flow routes probably operated at peak stages; at least portions of the latest Pleistocene floods bypassed the Spokane Valley and Rathdrum Prairie and flowed northwestward and then south into the Little Spokane River drainage and Spokane basin (Kiver and Stradling, 1982; Waitt, 1984). Flow also bypassed the Spokane Valley by overtopping the Coeur d'Alene Lake basin (Richmond and others, 1965) and spilling southward over divide crossings south of Spokane. Furthermore, during peak stages, unconsolidated sediment along the valley bottom was probably excavated, creating larger cross-section areas than now present. Water depths approaching 65 m in Coeur d'Alene and Hayden Lakes, lakes that occupy valleys blocked by flood deposits, may provide an indication of the altitude of Rathdrum Prairie when the deposits were emplaced. Flow at maximum stages may have entered Glacial Lake Co-

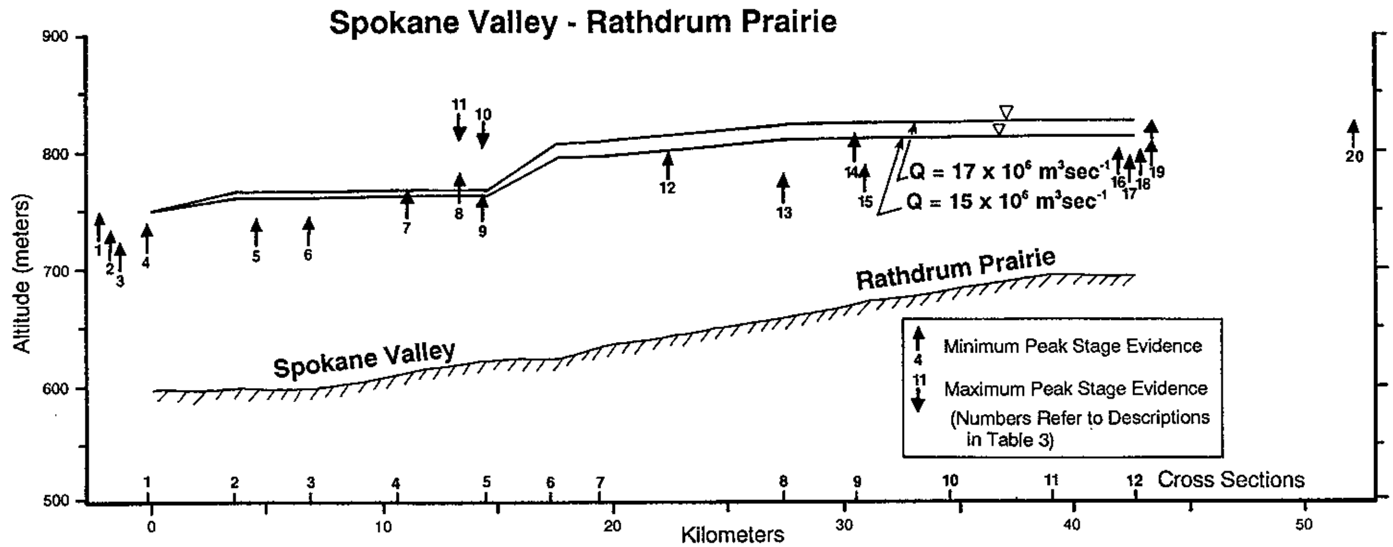

Figure 7. Calculated profiles for the Spokane Valley reach. A 17 million $m^3 \cdot sec^{-1}$ is the minimum discharge that results in a calculated profile that matches or exceeds all of the evidence for maximum flood stages. These profiles were calculated on the basis of the present valley topography.

lumbia, a lake formed by the Okanogan ice lobe blocking the Columbia River. At its maximum, stable level of 715 m (Atwater, 1986, p. 6; 1987), ponding extended up the Spokane Valley and inundated most of the study reach.

Because of these complexities and uncertainties, we have focused on determining a *minimum* estimate for the maximum Lake Missoula discharge in this reach. Because we are using the present topography for the flow modeling, calculated cross-section areas are probably smaller than those extant at peak discharge. The high-water indicators used for the flow reconstructions are consistent with the downstream presence of glacial Lake Columbia; however, equally large discharges would have coursed through the study reach at lower stages (at the downstream end of the reach) if driven by a steeper energy gradient associated with unimpounded conditions. The energy-loss coefficients used to calculate the water surface profiles for this reach are larger than those typically used in indirect discharge estimates. As a result, the water surface profile that best matches the stage evidence should be associated with a minimum discharge estimate.

Water-surface profiles were calculated through 12 cross sections over a distance of 45 km, assuming an initial downstream stage of 750 m. Manning roughness coefficients of 0.04 and 0.1 were chosen to estimate roughness losses over the valley floor and valley walls, respectively. We used contraction/expansion loss coefficients of 0.3 and 0.7.

Spokane Valley Modeling Results. For the above parameters and the present topography, a discharge of 17 million $m^3 \cdot sec^{-1}$ provides the best eye fit to the geologic evidence of maximum stages (Fig. 7). Because the constriction at cross section 5 exerts strong control on the upstream profile at high stages, the presence or absence of glacial Lake Columbia (at a stage of 715 m) has no effect on the discharge value required to emplace the maximum stage evidence at the upstream end of the reach.

Model variable sensitivity analyses and limits on a minimum estimate for the maximum discharge were postulated on a maximum flow stage of between 817 and 824 m at the upstream end of the reach (constraints imposed by stage indicators 19 and 20), with 824 m assigned as the most likely value of the maximum stage (Fig. 8). Similar to results at Wallula Gap, varying the assigned Manning roughness coefficients by ±25% has minimal (less than 10%) effect of the calculated discharge. Using smaller values of the expansion and contraction coefficients has a larger effect, raising the discharge required to achieve an 824-m altitude at the upstream end of the reach to about 20 million $m^3 \cdot sec^{-1}$. Con-

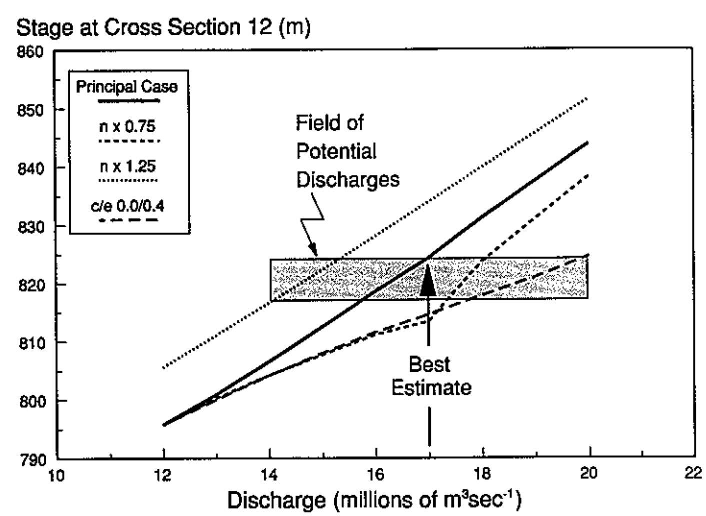

Figure 8. Stage-discharge curves for cross section 12 of the Spokane Valley reach for present topographic conditions. Considering uncertainties in maximum stages and reasonable ranges of energy-loss coefficients, a bracketing range for a minimum estimate of the maximum discharge within this reach is between 14 and 20 million $m^3 \cdot sec^{-1}$. For calculations using low energy-loss coefficients, flow conditions were predicted to be near critical at cross section 12.

sidering these ranges of model parameters and uncertainties in the maximum upstream stage, a minimum of 14 to 20 million $m^3 \cdot sec^{-1}$ flowed through the Spokane Valley. We emphasize that, because of our approach to reconstructing flow conditions for this reach, this is a *minimum* estimate for *maximum* flow through this reach, as well as being a minimum estimate for maximum glacial Lake Missoula outburst discharges.

This point is substantiated by evaluating the importance of uncertainty in the valley floor altitude at maximum stage. We performed a model trial assuming that the entire valley bottom was excavated to a depth of 60 m relative to its present altitude (consistent with the depths of the marginal lakes). For these conditions, the discharge that produces the required stages is about 27 million $m^3 \cdot sec^{-1}$. Clearly, the valley topography at the time of peak discharge is a major factor in establishing accurate discharge estimates in this reach. It is unlikely, however, that the valley floor was higher than its present altitude during maximum flood stages; hence our 17 million $m^3 \cdot sec^{-1}$ estimate is probably a valid minimum estimate.

The 17 million $m^3 \cdot sec^{-1}$ discharge determined here is less than the 21 million $m^3 \cdot sec^{-1}$ estimate calculated by Baker (1973) with the slope-area method. This discrepancy is probably generated by higher energy-loss coefficients used in this study and the more complete characterization of flow conditions afforded by the step-backwater calculation procedure.

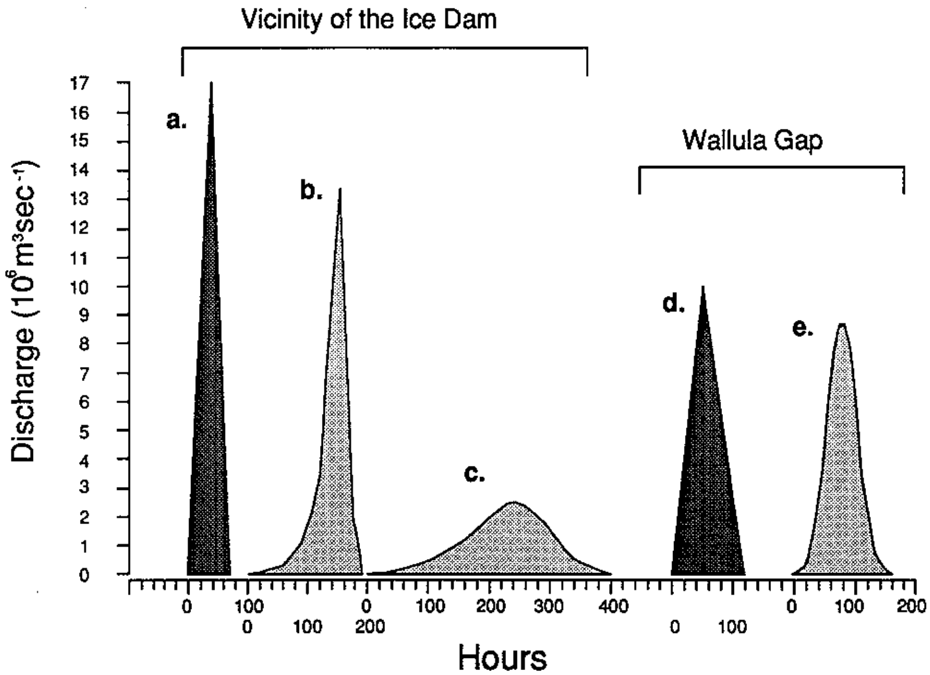

Figure 9. Postulated hydrographs for the largest Missoula Flood(s) (assuming complete emptying of Glacial Lake Missoula from its maximum level). All of the hydrographs represent the same total volume (2,184 km^3). a and d are those postulated in this study on the basis of maximum discharges near the outlet and at Wallula Gap. b and c are those proposed by Clarke and others (1984) as resulting from a jökulhlaup release. b represents the extreme case of no "tailwater ponding" (flow impeded by downstream ponding in the Spokane Valley) and immediate conveyance of all of the lake water to the breakout location. c is the more realistic situation of downstream hydraulic conditions affecting the rate of release as well as the delayed response at the breakout point because of the complex lake geometry. e is the hydrograph proposed for Wallula Gap by Craig (1987).

GENERAL IMPLICATIONS

Flood Hydrographs

Peak discharge estimates at Wallula Gap and in the Spokane Valley, in conjunction with known storage volumes in Lake Missoula and the Pasco Basin, permit some simple conclusions regarding the flood hydrographs at both localities. For this analysis, we make the following simplifying assumptions. (1) The flood hydrographs were triangular. This is consistent with the general shapes of flood hydrographs produced by dam failures and by some types of jökulhlaups (Costa, 1988). (2) Maximum discharges at both Wallula Gap and in the Spokane Valley were associated with complete emptying of Lake Missoula from its maximum highstand (1,265 m). (3) An insignificant quantity of water bypassed the Spokane Valley flow route. Important physical parameters include the volume of Lake Missoula available for flooding: 2,184 km^3, according to Clarke and others (1984); and the volume of hydraulically ponded water behind Wallula Gap: 1,210 km^3 at the 375-m ponding level as determined from the stage-storage curve of Craig (1987).

Constructing a hydrograph for the Spokane Valley reach that satisfies our peak discharge estimate and the volume of Lake Missoula requires a flow duration of about 70 hr (Fig. 9). Because the 17 million $m^3 \cdot sec^{-1}$ estimate is a minimum discharge value, it is possible that the actual flow duration was even less. A similar hydrograph can be drawn for Wallula Gap with the added limitation that at peak stage (375 m in the Pasco Basin) there is 1,210 km^3 of ponded storage upstream of Wallula Gap. This imposes a degree of asymmetry to the hydrograph: the waxing part of the hydrograph must have been less than 54 hr, and the waning period was at least 67 hr, with a total flow duration of at least 121 hr (Fig. 9).

Although these hydrographs must be viewed as first-order approximations, some conclusions can still be drawn. Flow duration associated with such a large discharge was probably remarkably short, on the order of three days within the Spokane Valley. Moreover, the peak discharge seems to have attenuated significantly between the outburst area and Wallula Gap. A large part of peak discharge diminution was probably the result of storage in the Pasco Basin. Stages in the Pasco Basin apparently rose at least 20% faster than they declined. Craig (1987) addressed some of the dynamic aspects of Missoula Flood hydrographs for portions of the flood route, including Wallula gap, and presented hydrographs with shapes that accord with his dynamic model (Fig. 9). Obtaining a complete dynamic picture of the maximum flood(s), however, remains difficult because of the plexus of flow paths across the Channeled Scabland and uncertainty regarding the degree of blockage of the Columbia River valley in north-central

Washington by the Okanogan lobe of the Cordilleran ice sheet.

The Mode of Ice-Dam Failure

Consideration of the discharge estimates obtained in this study for the Spokane Valley reach may have implications regarding the release mechanism associated with drainage from the highest level of glacial Lake Missoula. In an independent analysis that focused on the dynamics of the failure of the ice dam, Clarke and others (1984) used a physically based computer-simulation model (Clarke, 1982) to predict deductively the range of potential discharges associated with an assumed subglacial style of lake emptying (jökulhlaup) for the glacial Lake Missoula ice dam. Postulating various geometric models and combinations of parameters, they predicted a physically plausible range of maximum discharge values of between 2.5 and 15.3 million $m^3 \cdot sec^{-1}$. For their modeled conditions that most closely approximate the conditions of our step-backwater calculations—downstream stages of 750 m and a more steady contribution of outflow from the entire lake (their conditions of tailwater ponding and a trapezoidal duct conveying outflow to the release area)—Clarke and others (1984) estimated peak outflow rates of 2.6 to 3.5 million $m^3 \cdot sec^{-1}$.

Beget (1986), in commenting on the analysis of Clarke and others (1984), provided an empirical relationship between lake volume and peak discharge based on data from historic jökulhlaups. Extrapolating this relationship to the immensely larger lake volume associated with glacial Lake Missoula, he predicted a maximum Lake Missoula peak discharge of 2.1 million $m^3 \cdot sec^{-1}$. Beget (1986, p. 138) concluded that this estimate "compares favorably with the most conservative and realistic computer model runs of Clarke et al." and further implied that the field-based discharge estimate of Baker (1973, 21 million $m^3 \cdot sec^{-1}$) may be an order of magnitude too large.

Implicit in the analyses of Clarke and others (1984) and Beget (1986) is that the ice dam impounding glacial Lake Missoula, at its highest level, failed by buoyant lifting and enlargement of subglacial tunnels. Our best estimate of 17 million $m^3 \cdot sec^{-1}$ for the maximum glacial Lake Missoula discharge is a minimum estimate; nevertheless, it is larger than the estimated discharge values associated with a subglacial release, especially the more realistic models. We propose that at least one outburst may have been the result of a more cataclysmic failure of the ice dam than an outburst resulting from bouyant lifting and subglacial draining. A cataclysmic failure leading to the discharges estimated in this report may have involved complete rupture and removal of the blocking lobe of ice. Although there is no direct evidence of the mode of ice-dam failure, geologic evidence in the Channeled Scabland and in the glacial Lake Missoula basin is consistent with a late Wisconsinan cataclysmic failure of the ice dam followed by a phase of multiple floods associated with periodic subglacial releases.

Evidence in the Channeled Scabland. Waitt (1980) originally proposed that a minimum of about 40 late Wisconsinan floods, recorded by sets of rhythmite deposits in the Pasco Basin (commonly referred to in the literature as "Touchet beds"), were jökulhlaups from a regularly self-dumping glacial Lake Missoula. This hypothesis has been found to be consistent with recent interpretations from additional sites of similar stratigraphy preserved in backwater areas along the flood route (Waitt, 1984, 1985). A particularly extensive record is preserved in the Sanpoil River valley, where deposits interpreted to represent at least 89 floods are intercalated within lacustrine deposits of glacial Lake Columbia during blockage of the Columbia River by the Okanogan lobe of the Cordilleran ice sheet (Atwater, 1984, 1986). Interpretation of pertinent radiocarbon dates and varve counts indicates that these floods occurred between 15,550 ± 450 and 13,050 ± 650 yr B.P. Varve counts, flood-bed thickness, and grain-size variations exhibit tendencies consistent with a general up-section decrease in flood magnitudes and recurrence intervals (Atwater, 1986) in a manner consistent with repeated subglacial releases from a thinning ice dam impounding glacial Lake Missoula (Waitt, 1985; Atwater, 1986).

It is apparent from the sedimentologic record that these floods ranged widely in magnitude. Nevertheless, it is not clear that any of the floods recorded in the rhythmite sections is the one that emplaced the highwater evidence used to calculate peak discharge associated with the largest release. For example, in backflooded areas marginal to the Pasco Basin, multiple rhythmites are discernible only to an altitude of 255–260 m (Waitt, 1980, p. 672; 1985, p. 1285; Bunker, 1980, p. 60), whereas bedded silt, interpreted to have been flood-transported but not rhythmically bedded, is found up to altitudes of 350 m (Bretz, 1930, p. 408–415; Flint, 1938, p. 493–494; Newcomb and others, 1972, p. 17, Pl. 1). If a stage of ~260 m was the maximum achieved by the series of flows that deposited the highest rhythmites, a maximum corresponding discharge value can be estimated from the step-backwater modeling at Wallula Gap. For critical flow conditions in the Wallula Gap constriction, a 5.5 million $m^3 \cdot sec^{-1}$ discharge results in a 260-m ponding level in the Pasco Basin (Fig. 5). The corresponding volume of storage in the Pasco Basin is 340 km^3 of water (from the stage-storage relationship of Craig, 1987), less than 30% of the volume associated with a 375-m ponding level.

Pertinent conclusions that can be drawn from the relations in the Pasco Basin are (1) at least one late Wisconsinan flood achieved a stage of 370–385 m, and (2) interpreting each of the highest rhythmites as representing individual floods requires several flows exceeding stages of 255 m. The lack of evidence for multiple flows above 260 m does not rule out higher stages having been achieved repeatedly; however, the disparity between the discharge required to emplace the highest evidence of flooding and those required to emplace the evidence for multiple, repeated flows is consistent with a large flood resulting from a more rapid or complete failure of the ice dam followed by multiple floods that were released by buoyant lifting of the ice dam. A candidate for such a flood may be the large (about 350 m deep near Wenatchee), Wisconsinan-age flood that coursed down the Columbia River prior to complete blockage by the Okanogan lobe of the Cordilleran ice sheet (Waitt, 1977, 1982, 1987). This flood may predate the record of flood sediments in glacial Lake Columbia.

Evidence in the Glacial Lake Missoula Basin. Geologic relations in the glacial Lake Missoula basin are also consistent with the idea that there was a single large flood resulting from complete or rapid failure of the ice dam followed by multiple smaller flows that drained subglacially. Pardee (1942) documented cataclysmic flood features associated with a single episode of rapid drawdown from the maximum 1,265- to 1,280-m level of Lake Missoula. Superimposed on some of these features are shorelines indicating refilling of the basin to a maximum altitude of 1,080 m. Pardee (1942, p. 1598) stated, "in this new body of water no unusual currents seem to have developed." Pardee (1942) ruled out cataclysmic emptying of the younger lake, because of the unmodified nature of the high flood features upon which the shorelines are developed. Chambers (1984) concluded from stratigraphic studies of 40 cycles of bottom sediments within the Lake Missoula basin (Chambers, 1971) that glacial Lake Missoula drained repeatedly with at least 22 partial drainages of the lake to altitudes below 985 m. Chambers (1984), however, found "no evidence that each drainage was catastrophic." Depositional evidence of multiple water-level fluctuations is confined to only the lowest parts of the lake basin, thinning up the valley sides (Chambers, 1984). Chambers (1984) also noted that multiple cycles of lacustrine deposits are preserved within the troughs of giant current ripples in the Camas Prairie, further illustrating the disparity between a large, early release and later, lower-level fluctuations recorded by the cycles of lacustrine deposits. The volume of water available for flooding from a 1,080-m glacial Lake Missoula was about 1,055 km^3 (estimated from the hypsometric data of Clarke and others, 1984), less than 50% of the maximum lake volume. These observations are consistent with the view that only once in its history did the late Pleistocene Glacial Lake Missoula rise to the 1,265- to 1,280-m level and that the release from this level was more cataclysmic than subsequent releases.

Conditions at the Ice Dam. The failure mechanics of the ice dam depend on the local glacier profile, the geometry of the blockage, and the state of the glacier-bed interface. None of these conditions is known with certainty. The stratigraphic evidence for multiple, periodic, late Wisconsinan floods of systematically varying magnitude and frequency is consistent with a self-dumping, subglacial release mechanism for the flows that left these deposits. This mechanism has been observed for a number of historic jökulhlaups (for example, Thorarinsson, 1939; Post and Mayo, 1971; Sturm and others, 1987). Clarke and others (1984) and Waitt (1985), guided by these examples, quantitatively evaluated the possible conditions associated with such a failure mechanism for glacial Lake Missoula.

There are, however, mechanisms that could have resulted in a more rapid failure and a larger peak discharge. As Waitt (1985, p. 1282) noted, "preliminary calculations" (presumably associated with maximum lake-level conditions) indicate that the enlarging tunnel (due to dynamic thermal erosion) could expand to the point that the "tunnel roof collapsed and allowed most of glacial Lake Missoula to escape." Clarke and others (1984, p. 294) mentioned the possibility "that the tunnel roof collapses during the flood and the dam is swept away." The potential of such a failure is greater during a release from the maximum glacial Lake Missoula level because of the much greater amount of energy available for thermal erosion while the lake emptied through the ice dam. The energy release associated with passage of the maximum (1,265-m stage) glacial Lake Missoula through the outlet is about three times that of a 1,080-m stage glacial Lake Missoula.

Another possibility is that the ice dam failed because the shear strength of the lobe of ice blocking the Clark Fork River was less than the hydrostatic force exerted by the impounded water. This hypothesis is difficult to evaluate because the geometry of a blocking lobe of ice is not known for the time of maximum lake level. At its maximum extent during late Wisconsinan time, the Purcell Trench lobe occupied the entire Lake Pend Oreille basin. With such an arrangement, the ice dam would probably have been stable against such a failure mechanism. At less advanced positions, however, the western margin of the glacier would likely have calved into the Pleistocene predecessor to Lake Pend Oreille. Under such conditions, the breadth of the blocking lobe of ice could have been reduced rapidly if it became ungrounded, leading to a brittle and rapid failure of the ice dam. This type of failure would be somewhat analogous to the largest known historic glacial outburst flood that occurred in October 1986, when Russell Lake burst through Hubbert Glacier in southeastern Alaska (Mayo, 1988, 1989).

CONCLUSIONS

The field evidence for maximum Missoula Flood stages, in conjunction with step-backwater water-surface profile reconstructions, indicates a peak discharge of about 10 million $m^3 \cdot sec^{-1}$ at Wallula Gap and a minimum of about 17 million $m^3 \cdot sec^{-1}$ in the Spokane Valley (near the point of release). Considering uncertainties in the calculation parameters and the geologic evidence, the discharge at Wallula Gap can be limited to between 7.5 and 12 million $m^3 \cdot sec^{-1}$. The maximum discharge at the point of release is likely to have been larger than the 17 million $m^3 \cdot sec^{-1}$ discharge estimated for the Spokane Valley.

The discharge estimate for the Spokane Valley is generally larger than independent analyses of the peak discharges of maximum glacial Lake Missoula outbursts derived from theoretical and empirical evaluations of a subglacial release mechanism. This leads us to propose that at least one failure of the ice dam may have been by a more cataclysmic mechanism than the subglacial jökulhlaups postulated (Waitt, 1980, 1984, 1985; Atwater, 1986) for emplacement of rhythmic strata at diverse sites throughout the flooded region. The hypothesis of a larger, more cataclysmic, late Wisconsinan flood preceding multiple subglacial jökulhlaups is consistent with geologic evidence in the Channeled Scabland and in the glacial Lake Missoula basin.

Finally, it should be made clear that we are not attempting to disparage the well-constructed "case for periodic, colossal jökulhlaups." We are simply concerned that, in our view, anomalies still exist between some aspects of the field evidence and the conceptual models that have been advocated. The position that "the scores-of-floods hypothesis completes Bretz's imaginative theory" (Waitt, 1985, p. 1286) may prematurely divert attention from some of the outstanding problems that remain in interpreting the spectacular features of the Channeled Scabland.

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Basement-cover relations, Sevier orogenic belt, northern Utah

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ABSTRACT

Precambrian crystalline basement and sedimentary cover rocks were deformed within a regional anticline in the Wasatch Range of northern Utah during the Cretaceous to early Tertiary Sevier orogeny. This regional fold had a complex uplift and deformation history and is an imbricated ramp anticline that marks the transition to basement-involved deformation. Large-scale thrusting and folding within the basement-cored anticline produced about 60% shortening, and most of the slip on thrusts within the anticline was transferred into slip on frontal thrusts of the Idaho-Utah-Wyoming thrust belt.

A network of ductile deformation zones (DDZ's) accommodated heterogeneous internal deformation of the basement, and cleavage, minor folds, and vein arrays accommodated internal deformation of the sedimentary cover. Kinematics of internal deformation are consistent between basement and cover rocks in most areas and vary systematically with structural position, recording early layer-parallel shortening, localized layer-parallel shear and minor detachment of the cover, and later fanning of structures during large-scale folding. DDZ sets with opposite shear senses accommodated bulk coaxial flattening perpendicular to cleavage away from major thrust faults, and DDZ's with consistent shear sense produced simple shear near major thrusts. Cleavage intensifies and refracts and minor folds tighten in the cover near major thrusts and glide horizons. Vein arrays record principal extension down the dip of cleavage, minor extension parallel to the regional fold axis, and differential displacements at high angles to fold axes.

Slip on basement DDZ's produced about 5% to 20% bulk shortening subparallel to the basement-cover contact. Shortening in the cover ranges from about 5% to 30% in most areas but reaches 65% near imbricate thrusts and along glide horizons, reflecting heterogeneous simple shear. Principal extension is down the dip of cleavage, and minor extension parallels local fold axes, consistent with the geometry of mesoscopic structures.

INTRODUCTION

Relations between basement and cover deformation have important consequences for the overall evolution of orogenic belts. These relations partly control transfer of shortening from basement within interior parts to cover within frontal parts of orogenic belts, record differential transport and distortion of thrust sheets, and must be understood for balancing and restoring cross sections. Additionally, these relations partly control variations in crustal strength and the overall mechanics of orogenic belts.

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Important similarities and differences in structural style of basement and cover deformation are observed within different orogenic belts, including the Alps (Milnes and Pfiffner, 1977; Ramsay and others, 1983), the Appalachians (Mitra, 1979; Boyer and Mitra, 1988), the Canadian Rockies (McDonough and Simony, 1988), and the Scandinavian Caledonides (Roberts and Gee, 1985). Crystalline basement is generally undeformed in the frontal parts of orogenic belts, but is deformed along with the cover toward the interior parts of these belts (Hatcher and Williams, 1986). Deformation of basement is generally heterogeneous, being localized into complex networks of ductile shear zones (Ramsay and Graham, 1970; Mitra, 1978; Ramsay and Allison, 1979; Choukroune and Gapais, 1983; Lacassin, 1987). The geometry and spatial distribution of these zones partly control bulk deformation of basement and may vary between regions (Gapais and others, 1987). Deformation of the layered cover results in development of penetrative cleavage and folds of various scales (Mitra, 1987; Boyer and Mitra, 1988). The orientation and intensity of cleavage and folds may also vary between different regions and structural levels (Coward and Kim, 1981; Ramsay and others, 1983; Mitra, 1987). Despite recent advances, understanding quantitative relations between basement and cover deformation remains an important area of research.

The geometry and timing relations of regional structures within the Idaho-Utah-Wyoming thrust belt of the North American Cordillera are relatively well constrained (Royse and others, 1975; Wiltschko and Dorr, 1983), making this an excellent area for studying basement and cover deformation. Crystalline basement crops out for a length of >60 km within the culmination of an anticline exposed in the Wasatch Range of northern Utah (Figs. 1 and 2). The anticline is cut by imbricate thrusts, and different structural levels within thrust sheets are preserved.

Detailed studies of basement and cover deformation were conducted of four representative areas that cover a range of positions within the anticline: (1) the Francis Peak area, (2) the Antelope Island area, (3) the Bountiful area, and (4) the Ogden area (Fig. 1). Ductile shear zones that accommodated internal shortening of basement at deeper levels are exposed in the Francis Peak area on the eastern limb of the anticline, and in the Antelope Island area on the western limb. Relations between basement and cover deformation at shallower levels are preserved at the southern and northern ends of the culmination in the Bountiful and Ogden areas.

In this paper, I describe and discuss variations in style and detailed geometry of basement and cover deformation for different positions within the anticline. The following questions are addressed. (1) What is the large-scale structure of the anticline in relation to regional thrusting? (2) How is internal deformation within the anticline produced? (3) What are the relations between basement and cover structures at different positions within the anticline? (4) What are the magnitudes and principal directions of strain, and how are they related to mesoscopic structures? (5) How is displacement transfered from within the basement-cored anticline into faults within the cover toward the foreland?

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