Quaternary Geology and Geomorphology of the Lower Deschutes River Canyon, Oregon

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The morphology of the Deschutes River canyon downstream of the Pelton-Round Butte dam complex is the product of the regional geologic history, the composition of the geologic units that compose the valley walls, and Quaternary processes and events. Geologic units within the valley walls and regional deformation patterns control overall valley morphology. Valley bottom morphology is mostly the result of Quaternary events. These include several large landslides, which have caused retreat of the canyon rims but have also constricted the valley bottom with immense volumes of debris. In at least two instances (as at Whitehorse Rapids), landslides blocked the channel, resulting in ponding, breaching of landslide dams, and downstream floods. Large floods from other mechanisms have also formed many valley-bottom features along the lower Deschutes River. A large Pleistocene lahar resulting from a *circa* 0.07 Ma eruption of Mount Jefferson left bouldery deposits along the valley margins for most of the canyon length. The 15,000-12,700 ^{14}C yr BP Missoula floods backflooded *up* the Deschutes River from the Columbia River and mantled the downstream-most 60 km of Deschutes River valley with bedded silt and clay. A large, possibly meteorologic flood between 6,500 and 3,000 ^{14}C yr BP left abundant boulder bars and high sand and silt deposits that flank the channel in wider valley-bottom locations. In contrast, large historic main-stem floods, such as December 1964 and February 1996, had few effects on channel geomorphology due to the volume and coarseness of valley bottom deposits left by the earlier and larger floods and landslides.

INTRODUCTION

The overall valley morphology of the 160 km of the Deschutes River downstream of the Pelton-Round Butte dam complex reflects the regional geologic history and local processes and events of the past million years. The purpose of this paper is to give a brief description of the morphologic development of the present canyon and valley bottom, bridging the temporal and spatial scales spanned by previous papers in this volume describing the overall geologic and hydrologic setting of the Deschutes River basin [*Gannett et al.*, this volume; *O'Connor et al.*, this volume] and later papers that provide more detailed information on Holocene processes and events that formed and affected the channel and valley bottom downstream of the Pelton-Round Butte dam complex [*Beebee and O'Connor*, this volume; *Curran and O'Connor*, this volume; *Fassnacht et al.*, this volume; *Hosman et al.*, this volume]. Specific locations are referred to by River Mile (RM) as shown on USGS topographic quadrangles and Oregon State Water Resource Board maps; RM 0 is at the Deschutes River confluence with the Columbia River, and values increase upstream to RM 100.1 at the Reregulating Dam, the downstream-most structure within the Pelton-Round Butte dam complex (Figure 1).

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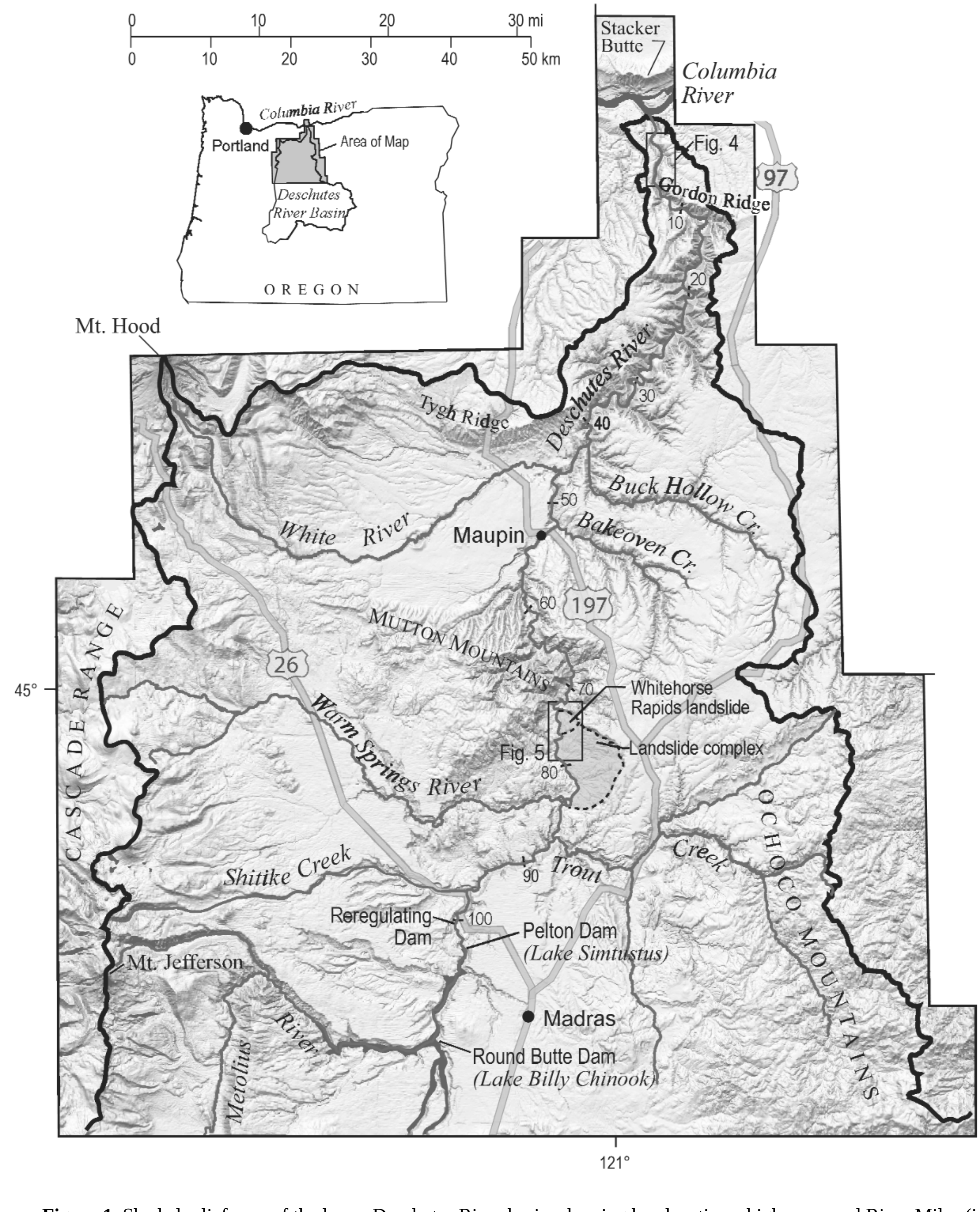

Figure 1. Shaded relief map of the lower Deschutes River basin, showing key locations, highways, and River Miles (in 10-mile increments) along the Deschutes River between the confluence with the Columbia River and the Reregulating Dam at River Mile 100.1. Topography from 30-m resolution digital elevation models produced by the U.S. Geological Survey. River Mile locations from the 1973 map of the Deschutes River drainage basin (map no. 5.6) produced by the State Water Resources Board, Oregon.

GEOLOGIC CONTROLS ON THE GEOMORPHOLOGY OF THE LOWER DESCHUTES RIVER

Canyon Morphology

Downstream of the Pelton-Round Butte dam complex, the Deschutes River flows within a canyon incised up to 600 m into the adjacent terrain (Figures 1, 2, and 3). Canyon depth corresponds to regional tectonic deformation of the once-planar top of the Miocene Columbia River Basalt Group (Figure 3). The deepest portions of the canyon occur where the Deschutes River crosses the trends of the uplifted Mutton Mountains (~RM 70) and the Tygh Ridge anticline (~RM 40). The canyon rims are lowest downstream of the Pelton-Round Butte dam complex between RM 100 and 85, near the town of Maupin (RM 50), and at the Columbia River confluence—all locations where the Deschutes River flows through post-Columbia River Basalt Group basins that have been partly filled with the Miocene-Pliocene Deschutes and Dalles Formations.

Canyon rims are formed of flows of the Columbia River Basalt Group, except where younger basalt flows cap structural basins. Upstream of RM 60, outcrops of lava flows and tuffs in the John Day and Clarno Formations form prominent ledges below the canyon rims (Figure 2a). Canyon walls are bare or talus-covered bedrock, landslide debris, and locally, thick colluvial aprons that descend in long sweeping slopes. Alluvial terraces up to 40 m above river level are preserved in wider portions of the canyon, notably near the Shitike Creek, Trout Creek, and Warm Springs River confluences. A blanket of fine sand, silt, and clay laid down by the late-Pleistocene Missoula floods mantles the lower portions of gentler slopes downstream of RM 40 (Figure 3).

The bedrock forming the canyon margins varies downstream in degree of internal deformation and weathering, susceptibility to landsliding, and grain size of weathering products, and these differences influence canyon and valley bottom morphology. From the Pelton-Round Butte dam complex downstream to about RM 85, the Deschutes River flows within a canyon with exposures of John Day Formation near water level, overlain by up to 150 m of Columbia River Basalt Group, which in turn is covered by as much as 150 m of Deschutes Formation capped by younger basalt flows. Downstream of RM 85, the Deschutes River approaches the uplifted Mutton Mountains and much of the canyon wall is formed of the Eocene-to-Miocene John Day Formation and underlying Clarno Formation. High and distant rims are capped by flows of the Columbia River Basalt Group. By RM 60, the Deschutes River has passed the Mutton Mountains and enters the tectonic basin near Maupin, where the top of the John Day Formation descends to near river level, and the northward-thickening Columbia River Basalt Group composes much of the valley walls. In the structural trough between RM 60 and 40, Dalles Formation and capping basalt flows overlie the Columbia River Basalt Group. Downstream of RM 50, the entire canyon is formed in Columbia River Basalt Group except for less than 100 m of Dalles Formation capping the tablelands downstream of RM 10.

From the Pelton-Round Butte dam complex downstream to about RM 60, the John Day and Clarno Formations are exposed at river level. These formations are susceptible to landsliding, and the canyon margins within this reach have been affected by an almost continuous series of landslide complexes. Landslides are largest where the John Day Formation has been uplifted the highest, such as along the axis of the Mutton Mountains anticline between RM 85 and 65. The largest landslides cover areas as great as 50 km^2, and extend as far as 6 km from canyon rim to river. As described in greater detail in later sections, some of these landslides temporarily blocked the Deschutes River. The overall result in this reach is that the Deschutes River winds tortuously through an irregular, landslide-dominated topography underlain by the alternating soft tuffaceous sediment and much harder volcanic rocks of the John Day and Clarno Formations, as well as by volcanic rocks displaced from canyon walls by slides, and remnant landslide dams of various ages.

Downstream of RM 60, the Deschutes River flows within a canyon largely composed of layered basalts of the Columbia River Basalt Group. The general canyon trend follows regional structures, especially where it parallels the Tygh Ridge anticline between RM 40 and 12, and the Gordon Ridge anticline between RM 12 and 8 (Figure 1). The somewhat regular meandering pattern (Figures 1 and 2b) within these general trends must reflect the course of a low-gradient and wandering proto-Deschutes River, before incision into the Columbia River Basalt Group sometime between 4 and 1 million years ago [*O'Connor, Grant, and Haluska*, this volume]. At present, the river flows within a consistently narrow canyon deeply incised in the resistant Columbia River Basalt Group (Figure 3b), between precipitous valley walls drained by short, steep tributaries.

Valley-Bottom Morphology

The influence of underlying geology on morphology of the valley walls and bottom of the Deschutes River below

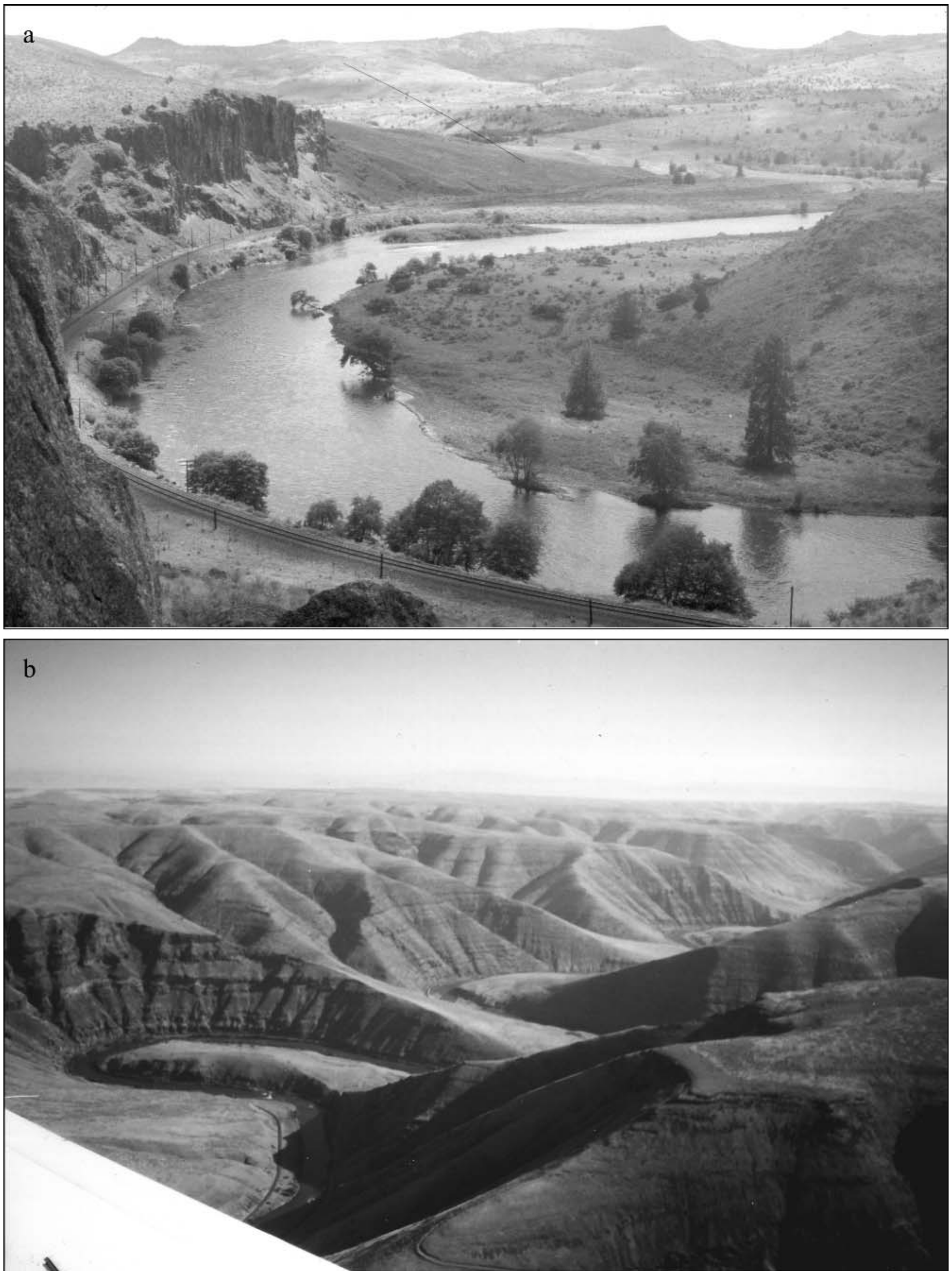

Figure 2. Photographs of the Deschutes River canyon. (a) June 24, 1998, view upstream (southwest) from RM 76.5. Cliffs in left foreground are rhyolite lava flows in the John Day Formation. Photograph by J.E. O'Connor. (b) Oblique aerial view downstream (northeast) from about RM 35, showing meanders incised into the layered flows of the Columbia River Basalt Group. Meander loop in the foreground is known as Beavertail, and the exposure shown in Figure 12 was excavated during construction of a railroad alignment across the inside of the meander loop. July 28, 1998, photograph by J. E. O'Connor.

the Pelton-Round Butte dam complex is illustrated by the transition in canyon morphology that takes place near RM 60, where the river leaves the Mutton Mountains anticline (Figure 3). Upstream of RM 60, the river flows on the soft and landslide-prone John Day and Clarno Formations, whereas the downstream 100 km of valley is mostly within the more resistant Columbia River Basalt Group. This distinction is most clearly shown by the pattern of valley width

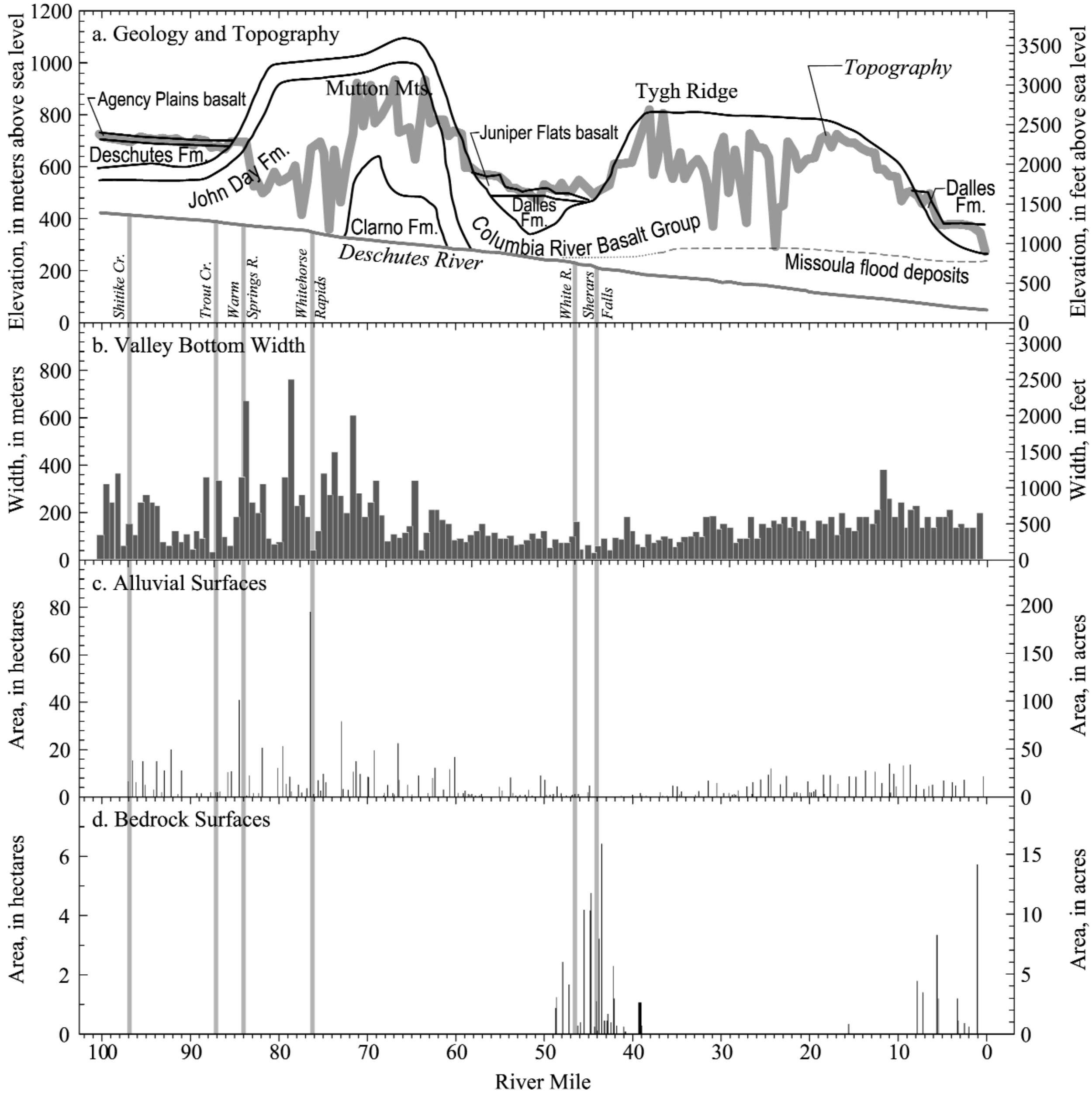

Figure 3. Canyon and valley-bottom characteristics of the lower Deschutes River. (a) Deschutes River profile and generalized geology and topography. River profile from elevation data on USGS 7.5-minute topographic maps. Topography is represented by the highest point within 2 km of the Deschutes River channel, measured from transects placed orthogonal to the channel and spaced at 1 km increments on USGS 7.5-minute topographic maps. Pre-Quaternary geology of the valley walls along the river generalized from position of contacts on 1-km spaced transects oriented perpendicular to the channel. Contact positions, in downstream order, after *Smith* [1987], *Smith and Hayman* [1987], *Waters* [1968a, b], *Bela* [1982], *Sherrod and Scott* [1995], and *Newcomb* [1969]. The distribution of Missoula flood deposits blanketing older rocks was mapped from aerial photographs and field reconnaissance. (b) Width of the valley bottom, as measured from transects oriented perpendicular to the channel and spaced at 1-km increments. (c) Distribution and size of alluvial surfaces within the valley bottom. (d) Distribution and size of bedrock surfaces within the valley bottom.

along the canyon bottom (Figure 3b). Upstream of RM 60, valley width varies greatly, ranging from 35 to 750 m. Most of the narrow reaches occur where relatively recent landslides or mass movements have encroached upon the valley bottom. There is no clear trend in downstream variation of valley width within this reach. Downstream of RM 60, valley width is almost everywhere less than 200 m, has much less spatial variation, and systematically increases downstream. Variation in valley width has in turn affected the sizes and distribution of valley bottom features such as alluvial surfaces (relatively flat surfaces underlain by fluvial deposits; Figure 3c), because there is additional space within the valley bottom upstream of RM 60 to accommodate their formation and preservation. *Curran and O'Connor* [this volume] present more information on the distribution of valley-bottom surfaces and their relation to channel and valley processes.

Another effect of overall canyon geology is the distribution of bedrock in the channel bottom and along channel margins. Between RM 50 and RM 40, bedrock crops out along much of the channel bottom and flanking valley-bottom surfaces (Figure 3d). In this reach, channel and valley width are at their overall minima. This section of the Deschutes River is a bedrock reach [e.g., *Tinkler and Wohl*, 1998], and it corresponds to the canyon entering the rising limb of the Tygh Ridge anticline (Figure 2). This reach is almost completely erosional, with little stored alluvium, due to either ongoing uplift or a more pronounced resistance of the slightly upturned Columbia River Basalt Group flows [e.g., *Bretz*, 1924]. A similar reach of abundant bedrock surfaces occurs in the final 13 km of the river above its confluence with the Columbia River. All of the large bedrock rapids and falls on the lower Deschutes River are within these two reaches [*Curran and O'Connor*, this volume].

QUATERNARY HISTORY OF THE LOWER DESCHUTES RIVER

Within the context of general geologic controls on valley morphology, several processes and events have further shaped the lower Deschutes River. Here we describe evidence and features left by several specific events, including volcanic eruptions (lava flows, lahars and pyroclastic fallout), landslides, and floods of various sources and magnitudes. This overview is not complete, especially for the Pleistocene history of the canyon, for which we have conducted only reconnaissance-level field studies. Description of Holocene and modern events and processes is brief here—further information may be found in the cited companion papers that appear later in this volume.

Volcanic Events

In the last million years, the Deschutes River canyon has been inundated by basalt flows, pyroclastic flows, and lahars that have locally altered canyon morphology and introduced immense quantities of sediment to the channel and valley bottom.

Lava flows. Just upstream of the study area, extensive flows from the Newberry Volcano and nearby vents flowed into and partly filled the Crooked and Deschutes River canyons between <0.4 and 1.2 Ma [Figure 2 of *O'Connor et al.*, this volume; *Russell*, 1905; *Stearns*, 1931; *Smith*, 1986, 1991; *Sherrod and others*, in press]. The vestige of these flows farthest downstream is the bench near RM 109, just upstream of Round Butte Dam. After partial filling of the canyons with lava flows, the Deschutes and Crooked Rivers re-incised through the lava flows and less resistant rocks at the margins of the flows. This incision was likely rapid at first, but slowed as the overall gradient declined and the downcutting river intersected more resistant strata. The pronounced increase in the general gradient of the Deschutes River upstream of the Crooked River confluence (now inundated by Lake Billy Chinook) indicates that the river has not yet completely re-incised to pre-lava flow levels. While these lava flows did not have a direct effect on the Deschutes River valley downstream of the Pelton-Round Butte dam complex, early stages of incision of new canyons likely generated tremendous volumes of sediment that would have been rapidly transported downstream.

Within the study area and close to the Columbia River confluence, a small basaltic lava flow from Gordon Butte entered the Deschutes River canyon from the east at RM 4.5 and flowed down the canyon for at least 2 km (Figure 4). Two remnants are preserved on valley margins on the east side of the river: one where the flow entered between RM 4.5 and 4.0, and another between RM 2.6 and 2.2 [*Newcomb*, 1969]. At both locations, the river apparently incised new canyons and channels west of the previous river corridor, resulting in locally narrower valley bottoms, flanked on the east by basalt cliffs formed from the lava flow remnants. Exposures at the base of the lava flow at RM 2.5 show basalt directly on top of unweathered gravel, indicating that the flow invaded the Deschutes River at an elevation about 17.2 m above the elevation of present low-flow water surface. An adit excavated 30 m into the gravel along the base of the basalt flow shows that the bottom of the basalt flow dips slightly eastward, suggesting that the thalweg of the river at the time of the flow was likely under the present extent of basalt. The basalt flow has not been dated, but a similar flow from Stacker Butte, 21 km to the NNW

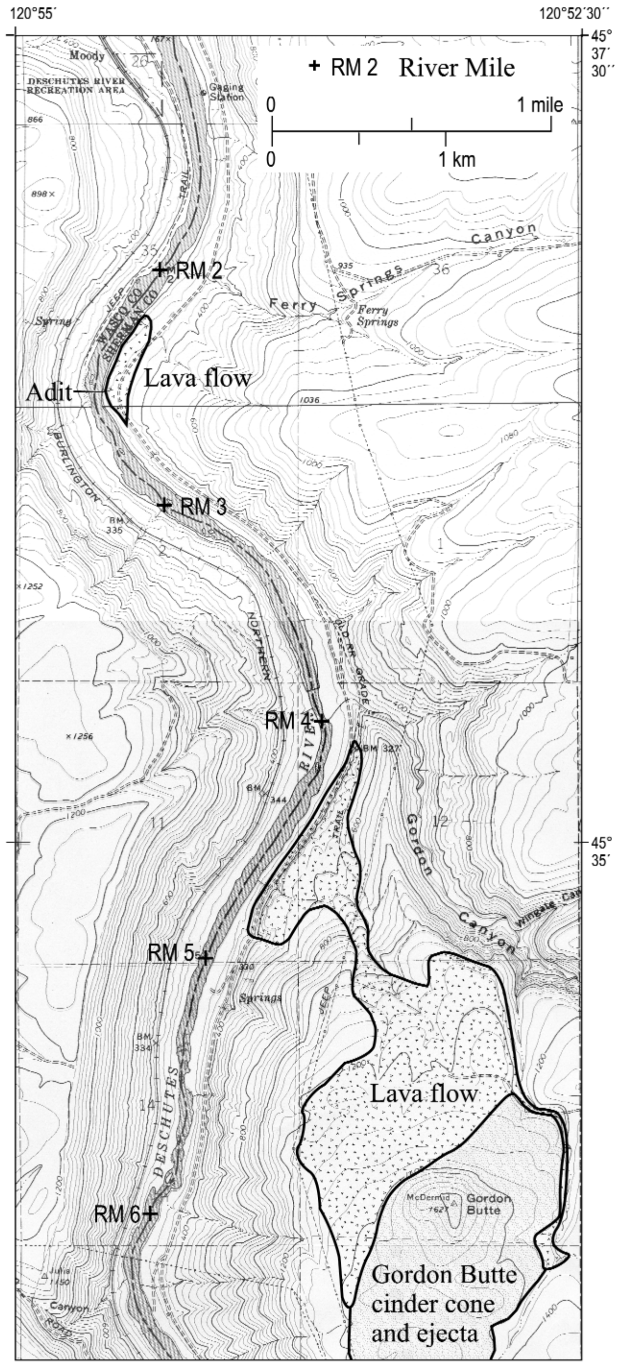

Figure 4. Quaternary cinder cone and lava flow from Gordon Butte. Topographic base from Emerson USGS 7.5-minute topographic quadrangle [contour interval 40 ft (12 m)]. Geology modified from *Newcomb* [1969].

(Figure 1), which flowed into the Columbia River at a similar elevation relative to the historic (pre-impoundment) river elevation, has been dated at 0.9±0.1 million years (Ma) [*Shannon and Wilson Inc.*, 1973, cited in *Bela*, 1982].

Lahars, pyroclastic flows, and tephra falls. Lahars (debris flows of volcanic origin) and other eruptive processes have episodically transported large quantities of sediment to the lower Deschutes River canyon. Concentrations of pumice granules and pebbles geochemically matching the 0.40-0.46 Ma Bend Pumice have been found in fluvial deposits exposed at RM 64 and RM 31 (Tables 1 and 2). Quaternary terraces between Pelton Dam (RM 104) and the Shitike Creek confluence (RM 96.6) contain a white pyroclastic-flow deposit, probably from Mount Jefferson [*Smith and Hayman*, 1987; G.A. Smith, 1998, written communication]. The elevation above the river and degree of soil development on the terrace surfaces suggest that the pyroclastic flow probably occurred at least 100,000 years ago.

Two outcrops along the lower Deschutes River contain deposits of a large lahar from Mount Jefferson that flowed most of the length of the canyon downstream of RM 84 (Figure 5). A 35-m-exposure at a railroad-cut into a terrace along the right side of the river at RM 84.0 shows round-cobble gravel capped by several pumice- and ash-rich beds of sand and gravel, including a 10-m-thick bed of poorly sorted bouldery gravel (Figure 5a). Glass in pumice clasts collected from a 20-cm-thick zone of pumice fragments at the bottom of the basal sandy bed (unit d of Figure 5a) geochemically matches a thick pumice fall from the last major explosive eruption of Mount Jefferson (Tables 1 and 2). At this exposure, the lahar deposits overlie about 5 m of unweathered fluvial gravel exposed to an elevation about 25 m above present river level, indicating that the channel was substantially higher at the time of the Mount Jefferson eruption than at present.

Another outcrop of a Mount Jefferson lahar from the same eruptive episode is exposed in a railroad cut at RM 14 (Figures 5b, 5c). The bottom of the exposure at this site consists of white silt-sized tephra conformably overlying grayish-tan silt, presumed to be loess (Figure 5c). This tephra, which also geochemically matches Mount Jefferson tephra and pumice (Tables 1 and 2), is capped by 50 to 75 cm of coarsening-up pumiceous sand and angular gravel, with diameters of isolated pumice clasts exceeding 5 cm. These pumice clasts also match Mount Jefferson tephra and pumice sampled from dispersed sites in Oregon and Idaho (Tables 1 and 2). Conformably overlying the angular gravel is 3 m of poorly sorted bouldery gravel. The contact between the two gravel units is gradational over about 30 cm, leading to the inference that both units were deposited contemporaneously

Table 1. Lower Deschutes River tephra and pumice correlations.

Site	Sample ID	Location (latitude and longitude, in degrees W and N)	Stratigraphic context	Correlation	Comments regarding stratigraphic relations, age and sources of age information
South Junction terrace RM 84.0	5/14/99-1(1)	44.8582121.0708	1-5 cm diameter pumice clasts in lahar deposit	Mount Jefferson	Correlative tephra exposed in south-central Idaho underlies a Yellowstone tephra that has been dated by fission-track on glass at 76±34 ka [*Pierce*, 1985].
Whitehorse landslide ARM 75.6	6/24/98-2(1)	44.9400121.0653	Silt-sized fallout tephra within closed depression on land-slide	Mount Mazama	6730±40 ^{14}C yr B.P. [*Hallett et al.*, 1997]
Dant debris flow RM 64.2	7/16/98-2(2)	45.0403121.0006	Pumice granules in fluvial and lacustrine deposits	Bend Pumice	Correlative pumice fall near Bend, OR, dated at 400 to 460 ka [*Sarna-Wojcicki et al.,* 1989; *Lanphere et al.,* 1999]
Caretaker Flat RM 61.7	7/16/98 1(17)	45.0670121.1125	Silt-sized fallout tephra interbedded with Holocene fluvial overbank deposits	Mount Mazama	(see above)
Beavertail RM 31.0	9/29/98-1(1)	45.3325120.9422	Pumice granules in Pleistocene fluvial deposit, 25.4 m above present river level	Bend Pumice	(see above)
Lockit siding RM 14.0	10/1/98-2(1)	45.4962120.8345	Silt-size fallout tephra on Pleistocene loess	Mount Jefferson	(see above)
Lockit siding RM 14.0	10/1/98-2(2)	45.4962120.8345	1-2 cm diameter pumice clasts at base (lowermost 5 cm) of lahar deposit overlying fallout tephra	Mount Jefferson	(see above)
Lockit siding RM 14.0	10/1/98-2(3)	45.4962120.8345	2-5 cm pumice clasts from lower 1.0 m of lahar deposit	Mount Jefferson	(see above)

or in close succession by a large lahar or lahars from Mount Jefferson. The lahar deposits are capped by 1.5 m of tan silt, probably Missoula flood deposits and loess. The base of the lahar is 13 m above the present water surface. Although the channel elevation at the time of the lahar at this location is not known, the presence of at least a meter of loess beneath the lahar deposit indicates that the river was several meters lower than the bottom of the lahar deposits.

The lahar-producing Mount Jefferson eruption was large and explosive, generating thick tephra falls in the 20 km surrounding the volcano, and lighter falls detectable as far away as southeast Idaho [*Beget*, 1981; *Conrey*, 1991; *Walder et al.*, 1999]. Pyroclastic flows entered drainages on the east and west sides of the volcano [*Conrey*, 1991]. Judging from the distribution of eruptive products, the eruption was likely concurrent with advanced glaciers (Gary Smith, University of New Mexico, written communication, 2001). Consequently, tremendous meltwater volumes were likely generated during major eruptions, significantly contributing to the size and travel distance of the lahars. The thickness and coarseness of deposits and >225 km travel distance, as well as the superposition of lahar deposits on >5 m of unweathered coarse gravel fill at RM 84 (Figure 5a) and loess that was apparently actively accumulating at RM 14 (Figure 5c), suggest that an eruption occurred during a time of substantial ice volume. Poorly exposed boulder

Table 2. Volcanic glass compositions of tephra in outcrops along the lower Deschutes River and inferred correlative and dated tephra (shaded). Composition values given as oxides, in weight-percent, recalculated to 100 percent on a fluid-free basis. Details of analytical methods provided in Appendix 2 of O'Connor et al., 2001.

Sample[a]	SiO_2	Al_2O_3	Fe_2O_3	MgO	MnO	CaO	TiO_2	Na_2O	K_2O	ΣWt%
					Bend Pumice Correlatives					
7/16/98-2(2)	74.69	13.70	1.85	0.11	0.07	0.80	0.14	5.11	3.55	100.02
9/29/98-1(1)	74.82	13.58	1.81	0.10	0.06	0.76	0.14	4.95	3.78	100.00
BPT-1 (Bend Pumice, near Bend, OR)	74.85	13.72	1.94	0.11	0.07	0.74	0.14	5.09	3.34	100.00
					Mount Jefferson Correlatives					
5/14/99-1(1)	74.58	14.36	1.77	0.34	0.04	1.60	0.25	4.80	2.27	100.01
10/1/98-2(1)	74.87	13.85	1.76	0.31	0.05	1.54	0.24	4.87	2.50	99.99
10/1/98-2(2)	74.56	13.93	1.81	0.31	0.05	1.55	0.26	5.01	2.53	100.01
10/1/98-2(3)	74.60	13.89	1.77	0.32	0.06	1.57	0.27	5.02	2.50	100.00
SD 1490-1 (Pumice at Mount Jefferson)	74.85	14.08	1.79	0.32	0.05	1.54	0.24	4.68	2.44	99.99
ARCO A2-8 (>0.40 Ma tephra in south-central Idaho)	74.98	13.78	1.79	0.31	0.05	1.54	0.26	4.77	2.52	100.00
					Mount Mazama Correlatives					
7/16/98 1(17)	72.54	14.52	2.20	0.47	0.06	1.58	0.43	5.40	2.79	99.99
6/24/98-2(1)	72.55	14.54	2.25	0.48	0.05	1.58	0.44	5.35	2.76	100.00
Mazama Ash average (n=100)	72.79	14.65	2.12	0.46	0.05	1.61	0.42	5.19	2.71	100.00
					Homogenous Glass Standard (18 electron microprobe analyses)[b]					
RLS 132 (±1σ)	75.4±0.1	11.3±0.2	22±0.04	0.06±0.01	0.16±0.01	0.11±0.01	0.19±0.01	4.9±0.01	4.4±0.1	100.00

[a]Sample location and stratigraphic context provided in Table 1.

[b]Replicates of a natural glass standard (RLS 132) provide indication of analytical precision.

deposits flanking the Deschutes River valley bottom at many other locations may also date from this Mount Jefferson lahar. A lahar from the same eruptive period also traveled far enough to enter the Willamette Valley from the east [*O'Connor et al.*, 2001]

The only quantitative age information for this Mount Jefferson eruption is from a trench in eastern Idaho, where stratigraphic relations show that the eruption closely predates a Yellowstone tephra that has a 76±34 ka[1] fission track date [*Pierce*, 1985], thus indicating that the eruption and associated lahars are probably 42-110 ka. A plausible time for the eruptive period is the penultimate ice age of 74-59 ka during marine oxygen isotope stage 4 [*Martinson et al.*, 1987].

A much more recent lahar from Mount Hood flowed down the White River and perhaps into the Deschutes River at the White River confluence at RM 46.4. Downstream of the confluence, deposits of coarse lithic-rich sand flank both sides of the Deschutes River for several kilometers downstream on surfaces just above the limits of historic flooding.

[1] ka = kilo-annum, or thousand years ago

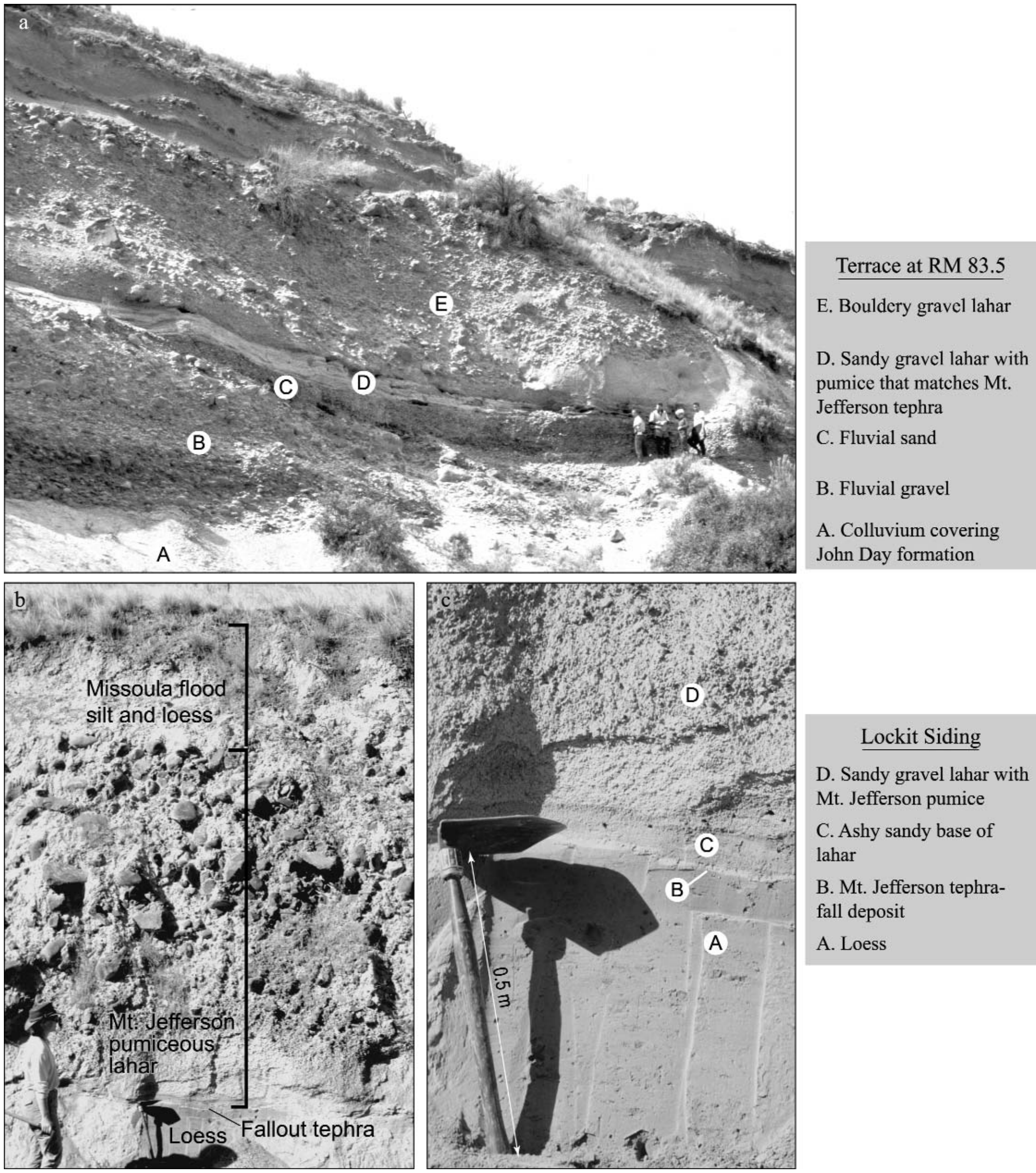

Figure 5. Exposures of lahar deposits from the circa 70 ka eruption of Mount Jefferson. (a) Part of 35-m-high terrace exposure at RM 83.5. The terrace surface is about 50 m above river level. A. colluvium covering tilted strata of the John Day Formation; B. fluvial gravel; C. fluvial sand; D. sandy gravel lahar with abundant pumice and volcanic glass fragments, including pumice clasts (collected near people in photo) that geochemically match Mount Jefferson tephra (Tables 1 and 2); E. poorly sorted bouldery gravel lahar containing abundant pumice clasts. May, 14, 1999, photograph by J.E. O'Connor. (b) Exposure of deposits at Lockit siding on the left valley margin, 12 to 17 m above the level of the Deschutes River and 22.5 km upstream of the Columbia River confluence (R.M. 14.0). The volcanogenic units overlie unweathered micaceous loess and are overlain by 15-12 ka Missoula flood deposits and loess. The fallout tephra at the base of the lahar and pumice clasts within the lahar match the composition of pumice and tephra from a large, *circa* 70 ka, Mount Jefferson eruption (Tables 1 and 2). Shovel handle is 0.5 m long. Photograph by J.E. O'Connor, Oct. 1, 1998. (c) Close-up of b (shovel is in the same location), showing: A. loess; B. Mount Jefferson tephra-fall deposit; C. ashy sandy base of lahar; and D. poorly-sorted sandy gravel lahar, containing abundant pumice clasts from Mount Jefferson. Photograph by J. E. O'Connor, Oct. 1, 1998.

These sands are similar to sandy lahar deposits farther up the White River that were deposited during the Old Maid eruptive period at Mount Hood, tentatively dated at AD 1781 (Patrick Pringle, Washington Dept. Natural Resources, and Thomas Pierson, U.S. Geological Survey, oral communication, 2002). The White River is continuing to entrain these sandy lahar deposits from eroding terraces, bringing much sandier sediment to the lower Deschutes River than the other Cascade Range tributaries [*McClure*, 1998; *Fassnacht et al.*, this volume].

Landslides

Between the Reregulating Dam (at RM 100) and RM 60, the valley margins are an almost continuous series of landslides composed of large and jumbled blocks of John Day and Clarno Formations [*Smith and Hayman*, 1987]. In addition, there have been large landslides within the Columbia River Basalt Group between RM 55 and 53, and smaller mass movements within Missoula flood deposits downstream of RM 8. Some of the largest landslide complexes involved up to 50 km^2 of terrain. Judging from the morphology of the landslides, these mass movements have a wide variety of ages, probably ranging from several hundred thousand years old to perhaps less than 10,000 years old. These large landslides have had several direct and indirect effects on the valley bottom and channel that persist today. Foremost, many have rafted large volumes of debris to the bottom of the canyon and significantly narrowed the valley bottom and channel. The mass movements at Whitehorse Rapids (Figures 6 and 7; RM 76) and near Dant (Figure 8; RM 64) completely blocked the channel, resulting in temporary impoundment of the Deschutes River. Such landslide dams are intrinsically unstable [*Costa and Schuster*, 1988], and boulder deposits downstream of Whitehorse Rapids and Dant suggest that both landslide dams breached rapidly enough to cause large floods. Additional evidence for such flooding is discussed in a later section.

Whitehorse Rapids landslide. The Whitehorse Rapids landslide (Figures 6 and 7) is probably the youngest and best documented of the major mass movements to affect the valley bottom. This landslide, shown on the 1968 geologic map by Waters, covers a semicircular-shaped area of 7 km^2 on the eastern valley slope. The slope slid or flowed westward, completely blocking the valley for a channel-wise distance of about 1 km. The entire landslide is within the John Day Formation, locally composed of rhyolite lava domes and flows, as well as tuffs and tuffaceous claystones [*Waters*, 1968a, 1968b; *Robinson et al.*, 1984]. The surface of the landslide is a chaotic landscape of knobs and closed depressions, bounded to the east by an arcuate scarp 60 to 100 m high. Whitehorse Rapids landslide is a morphologically younger part of an even larger landslide complex involving 50 km^2 of terrain that has moved westward into the Deschutes River Canyon between RM 85 and 74 (Figure 1).

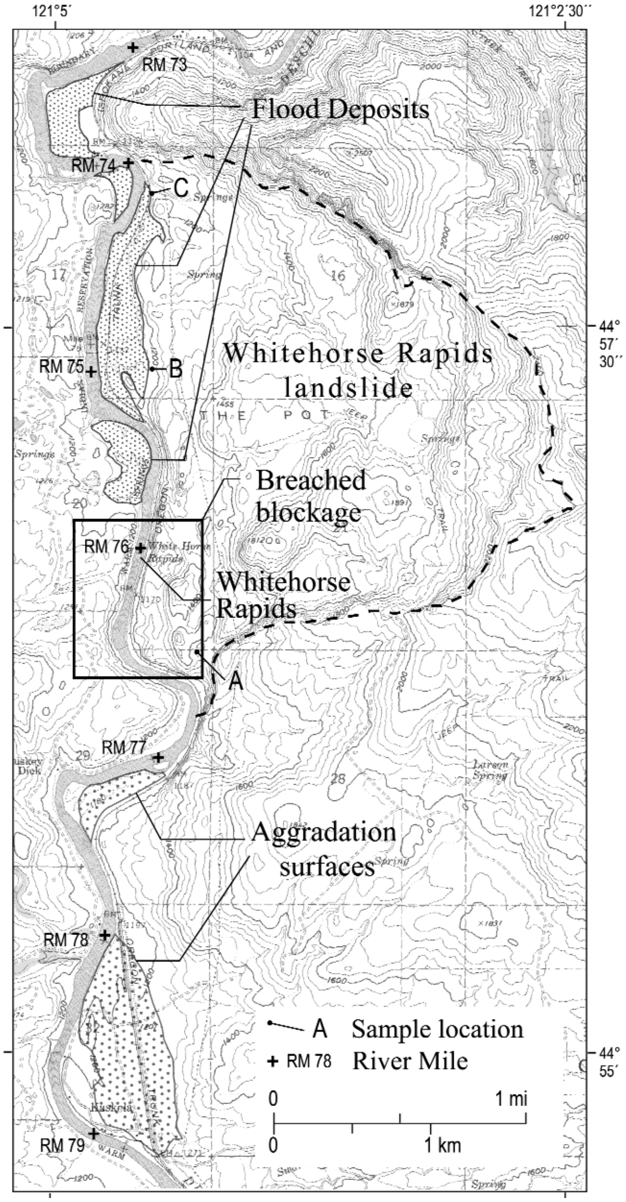

Figure 6. Portion of the Kaskela USGS 7.5-minute topographic quadrangle [contour interval 40 ft (12 m)] showing the Whitehorse Rapids landslide and associated features. Labeled stratigraphic sites are described in text and show locations where radiocarbon and tephra samples were collected and analyzed (Tables 1 and 2).

The present topography near the blockage indicates that the maximum impoundment could have been as high as 35-

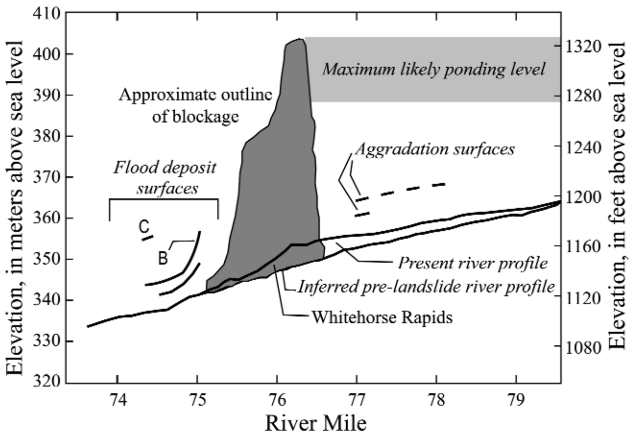

Figure 7. Profile of Deschutes River, and landslide and flood features near Whitehorse Rapids landslide. Deschutes River profile from the 1.5-m-contour (5 ft) river survey of *Henshaw et al.* [1914], conducted in 1911. The river mile locations in that survey are about 0.2 mi downstream of river mile locations shown on current maps. The pre-landslide river profile was based on extrapolation of the surveyed gradient of the Deschutes River from downstream of Whitehorse Rapids. Elevations of flood deposits and aggradation surfaces from site surveys and USGS 7.5-minute topographic quadrangles. 'C' and 'B' indicate locations of stratigraphic sites shown in Figure 6 and discussed in text.

50 m (Figure 7), which would have caused a temporary lake to back up nearly 30 km upstream to the Trout Creek confluence, although there is no evidence that ponding attained that level. Two gravel terraces, 5 m and 10 m above present river level, 1.4 km upstream from the blockage, are inferred to have resulted from aggradation behind the landslide dam. These terraces may indicate that the landslide dam lasted long enough to allow more than 10 m of sand and gravel to accumulate in the valley bottom (relative to the inferred pre-landslide channel profile shown in Figure 7); and that there were at least two periods when the channel was stable at different elevations significantly higher than the present channel elevation. The latter inference suggests that the landslide dam did not fail completely in one breaching episode, but with multiple episodes of incision, and that the time between incision episodes was sufficient for development of distinct aggradation surfaces. The landslide has not yet been completely breached—Whitehorse Rapids, which drops about 12 m in 1 km, is the remains of the landslide dam, and is composed of blocks too large to have been moved during incision. A comparison of the present river profile with the overall channel profile trend indicates that the present channel is as much as 7.5 m higher than the pre-landslide channel elevation (Figure 7).

There is no precise information on the age of the landslide and resulting flood(s). An excavation into the floor of a small closed depression on the landslide exposed Mazama ash 1.1 m below the surface (site A in Figure 6; Tables 1 and 2), indicating that the landslide formed before the 6730±40 ^{14}C yr BP [*Hallet et al.*, 1997] climactic eruption of Mount Mazama. A small charcoal clast in a high flood deposit exposed in a railroad cut at RM 74.4 (site C in Figure 6) had an age of 38,740±540 ^{14}C yr BP (Table 3), providing a maximum bracketing age on both the landslide and the flood that apparently resulted from breaching of the landslide dam (discussed in a subsequent section). A piece of charcoal sampled from sediment that accumulated on top of flood deposits (site B in Figure 6) indicates that the high flood deposits predate 3840 ^{14}C yr BP (Table 3). In summary, these dated samples suggest that the landslide occurred between 38,740±540 ^{14}C yr BP and 6730±40 ^{14}C yr BP, and the resulting flood(s) took place between 38,740 ^{14}C yr BP and 3840±40 ^{14}C yr BP. The thick and coarse accumulation of gravel upstream of the blockage may indicate that the landslide and resulting aggradation occurred during a period of enhanced sediment production in the basin, perhaps coincident with aggradation of the Willamette River (the adjacent basin to the west) during the last glacial maximum of 30,000 to 22,000 years ago [*O'Connor et al.*, 2001].

Dant debris flow. Another mass movement blocked the Deschutes River channel near the present-day vacation community of Dant at RM 64 (Figure 8). This mass movement has a flow-like morphology, with 5- to 10-m-high subparallel ridges aligned with the apparent flow direction for the lowermost kilometer of the deposit. The flow deposit is not traceable to a distinct mass movement or head scarp, but emerges from 1 km^2 of terrain that presumably partly collapsed. Like the Whitehorse Rapids landslide, the source area of the Dant debris flow is entirely within the John Day Formation.

The Dant debris flow apparently blocked the Deschutes River for a length of about 0.5 km. Bedded silt, sand, and gravel exposed in a road cut and underlying a colluvial slope at the upstream margin of the blockage (Figure 8) is inferred to have been deposited in the lake dammed by the deposit, and indicates that the blockage was at least 30 to 40 m high. Fluvially transported pumice granules deposited within these lacustrine deposits are chemically similar to the Bend Pumice [*Taylor*, 1981; Tables 1 and 2], and may indicate that the landslide occurred close to the 0.40-0.46 Ma age [*Sarna-Wojcicki et al.*, 1989; *Lanphere, et al.*, 1999] of the pumice.

Boxcar landslide complex. Three contiguous landslides, all smaller than the Whitehorse and Dant mass movements,

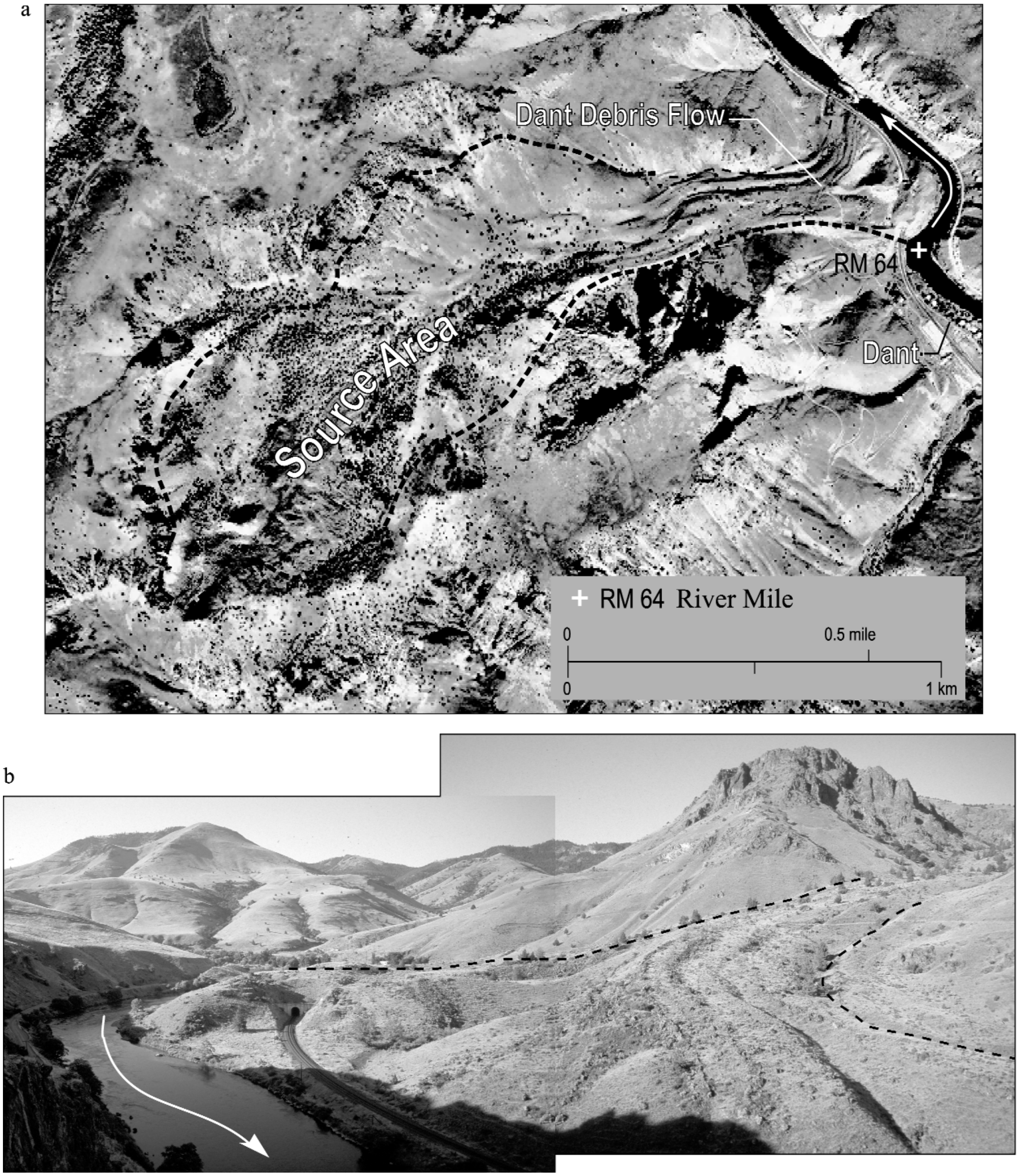

Figure 8. Dant debris flow. (a) Vertical aerial photograph (1995, Portland General Electric) showing flow deposit and inferred source area. (b) Two-photograph panorama directed upstream (south) showing terminus of flow deposit. Railroad and tunnel indicate scale. July 17, 1998, photographs by J. H. Curran.

Table 3. Radiocarbon ages.

Site	Sample ID (field label and laboratory no.)	Location (latitude and longitude, in degrees W and N)	Material	Corrected conventional ^{14}C age BP ±1 sigma[a]	$^{13}C/^{12}C$ ratio (%)
Whitehorse Landslide B	6/25/98-2(1) Beta 122390	44.9559 121.0825	charcoal	3840±140	-23.6
Whitehorse Landslide C	6/25/98-3(3) Beta 121597	44.9666 121.0750	organic fragments	38,740±540	-26.6

[a] Radiocarbon ages (in ^{14}C yr B.P.) are calculated on basis of Libby half-life for ^{14}C (5568 years). The error stated is ±1σ on basis of combined measurements of the sample, background, and modern reference standards. Age referenced to AD 1950.

have partly blocked the lower Deschutes River between RM 55 and RM 53 near Boxcar Rapids (Figure 9). These landslides moved northwestward, and each involved 0.5 to 1.0 km^2 of eastern valley margin, which is formed entirely of Columbia River Basalt Group and overlying loess. A head scarp, up to 20 m high and extending 150 m above the present river level, bounds each of the landslides. Cross-cutting relations among the bounding scarps and distinct differences in the morphology of the landslide surfaces indicate that the middle landslide is the youngest, and the landslide to the northeast is the oldest. The landslides presently constrict the channel and valley bottom between a steep, 150-m-high, canyon slope formed in the Columbia River Basalt Group on the left, and 60-to-70-m-high bouldery slopes along the landslide toes. Some or all of these landslides may have blocked the river, although we have found no evidence of ponding upstream. Coarse bouldery deposits at Boxcar Rapids (Figure 9) and downstream may have resulted from breach of one or more landslide dams at this site.

Other hillslope processes. These previously discussed mass movements are the most dramatic hillslope processes modifying the valley bottom, but not the only ones that have affected valley-bottom and channel morphology [*Curran and O'Connor*, this volume]. Rockfall and talus aprons reach the valley bottom in places, contributing coarse sediment to the valley and channel, especially where Columbia River Basalt Group is exposed at river level. Rills and gullies have locally contributed fine sediment to the valley bottom. Mecca Flat, at RM 95, appears to be composed of three coalescing alluvial fans that emanate from unvegetated gullies in the east canyon walls (Figure 10a). These gullies and several others nearby are incised into the fine sediment of the John Day Formation, possibly triggered by breaching of the protective cover of rubble derived from basaltic lava flows higher in the canyon walls. Rills and gullies have also dissected hillslopes flanking the lower 60 km of the Deschutes River valley, where late Pleistocene Missoula flood deposits and loess have mantled the valley with up to several meters of silt and clay (Figure 10b). Rill and gully formation triggered by local convective storms has likely always occurred, but the frequency and volume of rill and gully erosion have probably increased during the last two centuries due to domestic grazing, rerouting of drainage by rail and road construction, and field runoff.

Floods and Fluvial Processes

Large floods and debris flows generated by a variety of mechanisms have coursed down and into the Deschutes River canyon during the Quaternary. These floods and their transported materials are mainly responsible for the present morphology of the river channel and floodplain features. Mainstem Deschutes River floods of different sizes and sources have formed nearly all of the alluvial surfaces that flank the channel and have carved the bedrock channels in the bedrock reaches [Curran and O'Connor, *this volume*]. Tributary floods and debris flows have built up alluvial fans at tributary junctions. Backflooding of the Deschutes River by the late Pleistocene Missoula floods flowing down the Columbia River also left deposits on the Deschutes River valley bottom and margins.

Landslide breach floods. Large bars of coarse bouldery gravel downstream from landslides are evidence that breaching of temporary landslide dams has been an important mechanism producing large floods on the lower Deschutes River. These gravel bars locally contain subrounded boulders with diameters exceeding 5 m. Immediately downstream from breached landslide dams, flood bars are far above stages achieved by more recent floods. Flood deposits are lower and finer farther down-

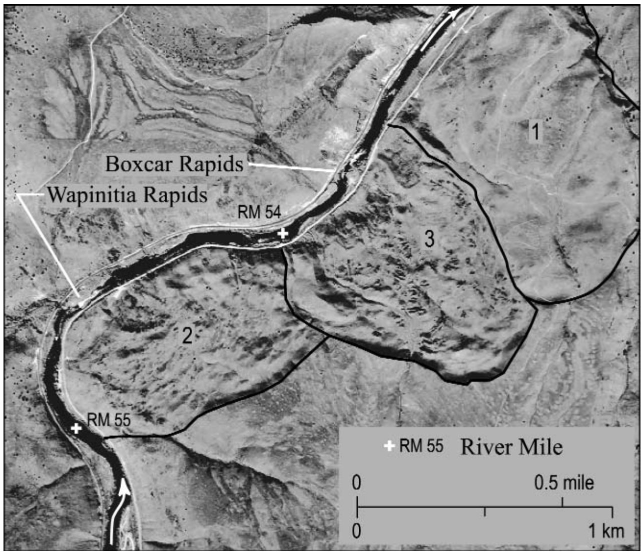

Figure 9. Vertical aerial photograph (1995, Portland General Electric) of landslides near Boxcar Rapids.

stream of breaches, and indistinguishable from flood deposits left by other types of floods. Between RM 100 and RM 50, bouldery gravel deposits can be traced up to 5 km downstream from the landslide breaches at Whitehorse Rapids, Dant, and Boxcar Rapids. The evidence is most compelling for the Whitehorse Rapids landslide, where the landslide, upstream aggradation features, and downstream flood features are all logically linked (Figures 6, 7, and 11). Nevertheless, bouldery deposits downstream of the Dant debris flow and the complex of landslides near Boxcar Rapids are strong evidence that such breaches have been a repeated process.

Based on limited information on the age of the landslides and resulting floods, most of these deposits are likely of Pleistocene age. Nevertheless, the clasts left in deposits by landslide-breach floods are exceptionally large and have been very resistant to subsequent modification. Consequently, landslide-related flood deposits form persistent elements of the Deschutes River valley bottom. These features include flood bars adjacent to the channel downstream of all the identified landslide breaches, and in-channel boulder accumulations that now form rapids or riffles downstream of breaches, such as Trout Creek Rapids (RM 87), the bouldery riffles downstream of Whitehorse Rapids landslide, and the Four Chutes Rapid 0.8 km downstream from the Dant debris flow. Boxcar Rapids may also be part of an accumulation of huge, 5-to-10-m diameter boulders, deposited during the breach of one of the nearby landslides [Figure 9; *Curran and O'Connor*, this volume].

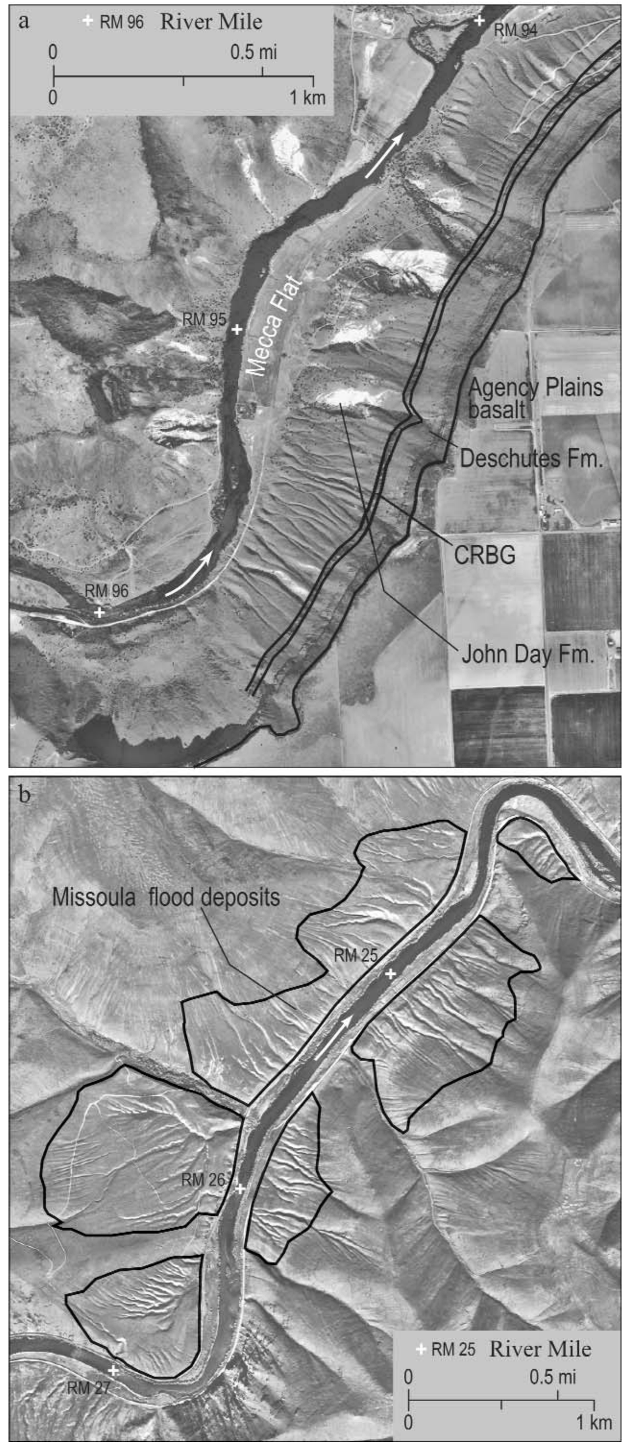

Figure 10. Vertical aerial photographs (1995, Portland General Electric) of gullies and rills. (a) Large gullies and rills in the John Day Formation near Mecca Flat at RM 95. CRBG is Columbia River Basalt Group. (b) Rills developed in Missoula flood sediment near RM 25. Missoula flood sediment mantles valley slopes developed in Columbia River Basalt Group.

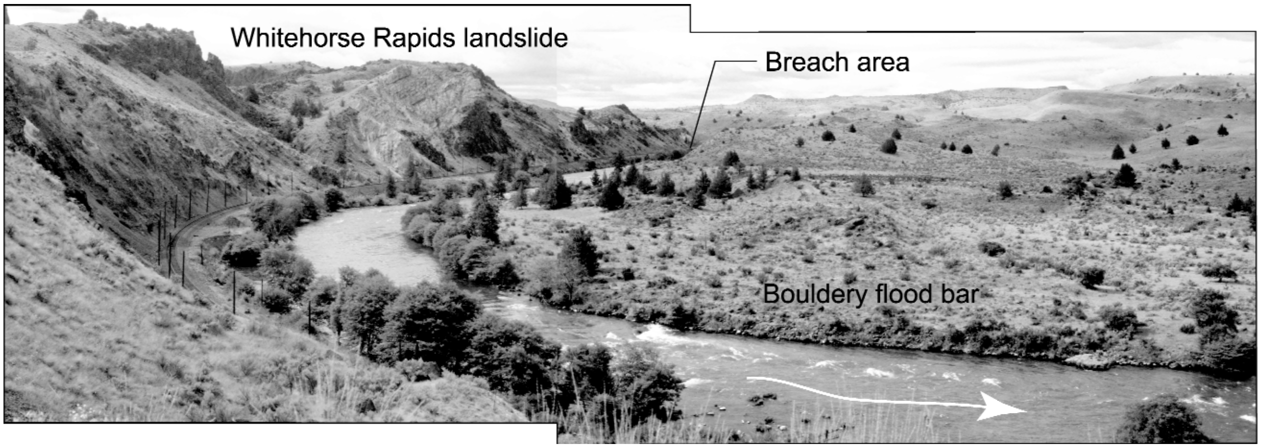

Figure 11. Two-photograph panorama of boulder flood bar downstream from the breach of Whitehorse Rapids landslide. View upstream (south) from right valley slope near RM 75.0. June 25, 1998, photographs by J. E. O'Connor.

Missoula floods. The late Pleistocene Missoula floods resulted from cataclysmic releases of ice-dammed Glacial Lake Missoula [*Bretz*, 1969], sending 60 to 90 large floods [*Atwater*, 1986] down the Columbia River between 15 and 12.7 ka [*Waitt* 1980, 1985; *Atwater*, 1986]. In the Columbia River valley at the confluence of the Deschutes River, the maximum stage of the largest Missoula flood was about 300 m above sea level or about 250 m deep [*O'Connor and Waitt*, 1995] If maintained at steady state, this water would have back-flooded the Deschutes River valley as far upstream as RM 60. Silt, clay, and sand carried in suspension by these floods mantles the valley slopes adjacent to the Deschutes River to an elevation of 275 m above sea level and as far upstream as RM 45 (Figures 3 and 12). Erosion of these fine-grained deposits has been a continuing source of sediment to the lowermost 60 km of the lower Deschutes River (Figure 10b).

The Outhouse flood. At several locations along the Deschutes River canyon downstream of the Pelton-Round Butte dam complex, high cobble and boulder bars are evidence of a Holocene flood much larger than the largest historic floods of 1964 and 1996 (Figure 13). We have termed this flood the "Outhouse flood," because of the apparent preference of the Bureau of Land Management for placing outhouses on bouldery bars deposited by this flood. Features, deposits, and the timing of this flood are discussed in more detail by *Beebee et al*. [this volume], *Curran and O'Connor* [this volume], and *Hosman et al.* [this volume]. In summary, many valley bottom features, including large boulder-gravel point bars and islands that stand several meters above maximum historic flood stages, similarly high stripped bedrock surfaces and adjacent boulder deposits, and large accumulations of sand and silt in backwater areas, are inferred to have been produced by this flood. Hydraulic modeling indicates that the peak discharge of the Outhouse flood was about twice that of the largest historic floods. Evidence for increasing peak discharge in the downstream direction indicates that this flood was likely generated by a meteorological mechanism rather than by some sort of natural dam failure [*Beebee and O'Connor*, this volume]. Radiocarbon and tephra relations constrain the age of the flood to between 7.6 and 3.0 ka.

Like the deposits resulting from the landslide dam failure floods, the cobbly and bouldery Outhouse flood bars are extremely resistant to subsequent erosion. The clasts composing these bars are larger than clasts entrained by modern floods, and large portions of these bars stand above stages achieved by the largest historic floods of 1964 and 1996. Only locally are these large bars eroded where high velocity flows of modern floods attack bar edges, but nowhere does it appear that cumulative erosion has exceeded more than a few percent of their original extent. It is far more common for these Outhouse flood bars to have been mantled by silt and sand by more recent flooding, especially on upstream apices.

Recent mainstem floods. Large floods have historically affected the mainstem of the Deschutes River downstream of the Pelton-Round Butte dam complex. Notable floods include the December 1964 and February 1996 floods, both of which had discharges of about 2000 m^3/s at the mouth and were the largest recorded in nearly 100 years of systematic streamflow measurements. Both of these floods, as well as the large historic flood of December 1861, were regional 'rain-on-snow' events caused by rapid melting of heavy mountain snowpacks by warm and wet storms from the south Pacific. The Willamette River also flooded during

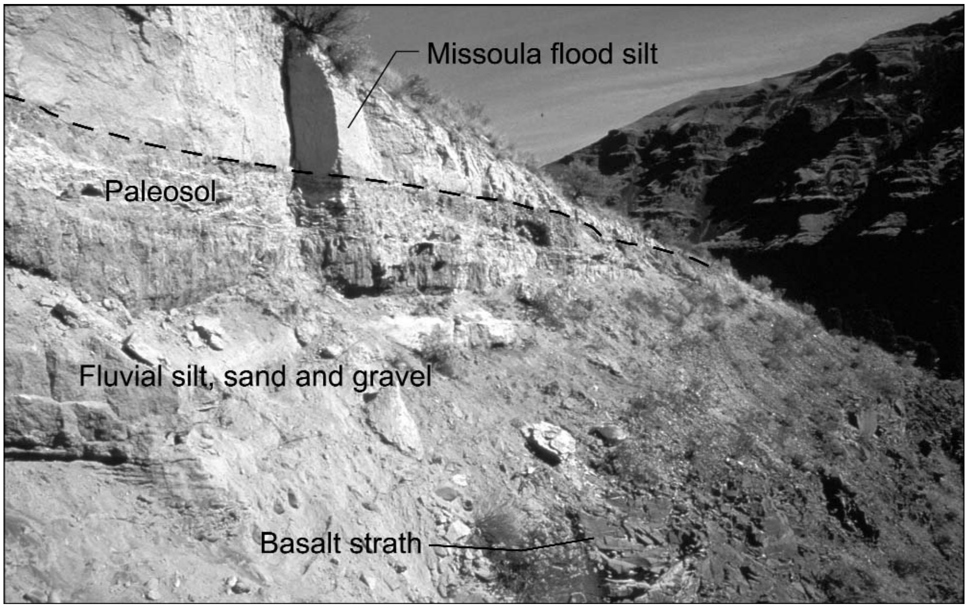

Figure 12. Exposure of 3 m of Missoula flood silt capping a paleosol developed in terrace gravel on right valley margin near Beavertail Campground at RM 31.0. The upper 2 m of terrace sand and gravel contains local concentrations of pumice grains that are geochemically similar to the *circa* 0.4 Ma Bend Pumice (Tables 1 and 2). September 29, 1998, photograph by J. E. O'Connor.

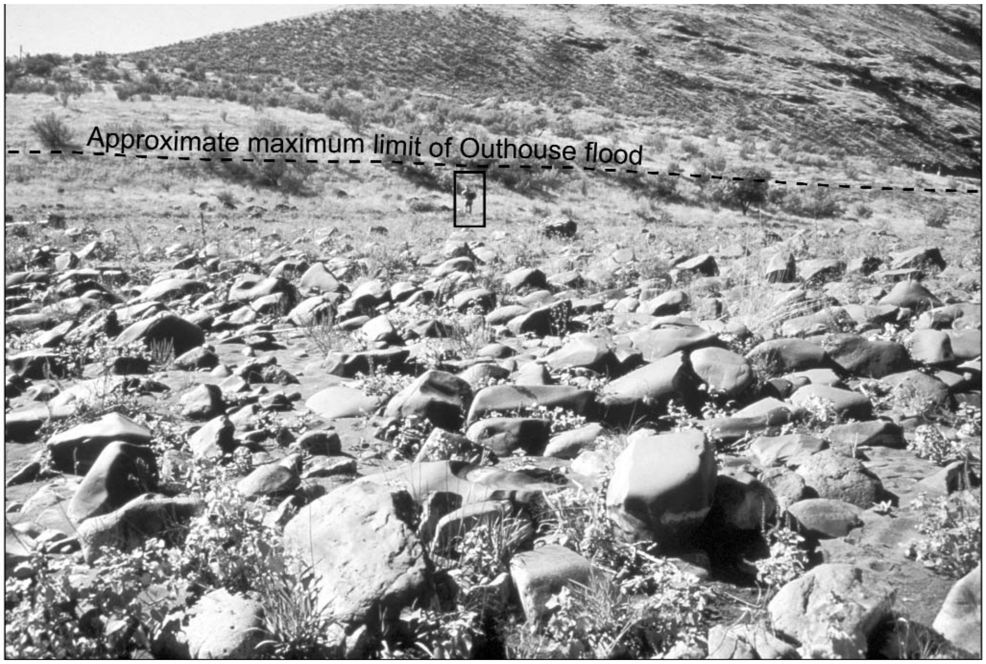

Figure 13. Boulder gravel of an Outhouse flood bar on the left bank at RM 4.2. Intermediate diameters of most boulders between 30 and 50 cm, but some intermediate diameters exceed 1 m. Sand between boulders deposited by the February 1996 flood, which inundated the bar to the elevation at which the person is standing, approximately 4 m lower than the prominent trim-line inferred to have been formed by the Outhouse flood. October 2, 1998 photograph by J. H. Curran.

each of these storms, as did the John Day River in 1964. On the Deschutes River, the 1861 flow was reported to be "higher than was ever known to white man or aboriginal" (Salem Statesman, December 23, 1861). Early settlers in the Willamette Valley vaguely recorded a large flood in the fall of 1813 that achieved stages close to, but probably slightly less than the 1861 flood [*Brands*, 1947]. Given the correspondence between the two basins, there was likely also an exceptionally large flow in the Deschutes River Basin in 1813. Flood frequency analysis of large main-stem floods, based on the historic record of flooding augmented by a 6000-yr stratigraphic record of Deschutes River floods is reported in *Hosman et al.* [this volume].

The February 1996 flood eroded some islands and margins of alluvial surfaces flanking the Deschutes River channel, and deposited gravel, sand, and silt in some overbank areas and island margins [Figure 14; *Curran and O'Connor*, this volume]. But compared to the consequences of large floods on most other alluvial rivers, there were few major or persistent geomorphic effects of the February 1996 flood in terms of bank erosion and deposition, channel scour and fill, and changes in substrate conditions [*McClure*, 1998; *Curran and O'Connor*, this volume, *Fassnacht et al.*, this volume]. The major broad-scale effects on the valley bottom due to flooding of this magnitude are the episodic accumulation and erosion of silt and sand overbank deposits on alluvial surfaces away from the main channel. A common depositional location is on the upstream ends of Outhouse flood bars (Figure 13). In these settings, high flows ramp up and slacken against the ascending bar surface. During the February 1996 flood, silt, sand, gravel, and much of the large wood carried by the flood was stranded on the upstream ends of many of these large bars. Wood accumulations locally impeded or concentrated flow, controlling patterns of scour and deposition of these surfaces. These alluvial surfaces may serve as the major sediment storage sites for sand and silt transported into and down the Deschutes River canyon. The sporadic accumulation and release of fine sediment depends on the sequence of flows, flow duration, suspended sediment and wood abundance, vegetation conditions, and the local hydraulic environment. Only the rare floods of magnitude similar to that of February 1996 are large enough to inundate these surfaces and cause significant sediment mobilization.

Tributary floods and debris flows. U.S. Geological Survey 7.5-minute quadrangle maps show 272 tributaries entering the Deschutes River between the Reregulating Dam at RM 100.1 and the Columbia River confluence. These tributaries contribute 6,904 km^2 of the total 27,460 km^2 drainage area of the Deschutes River basin. Most of the drainage area below the dams belongs to three large western tributaries that drain the eastern slopes of the Cascade Range, and three

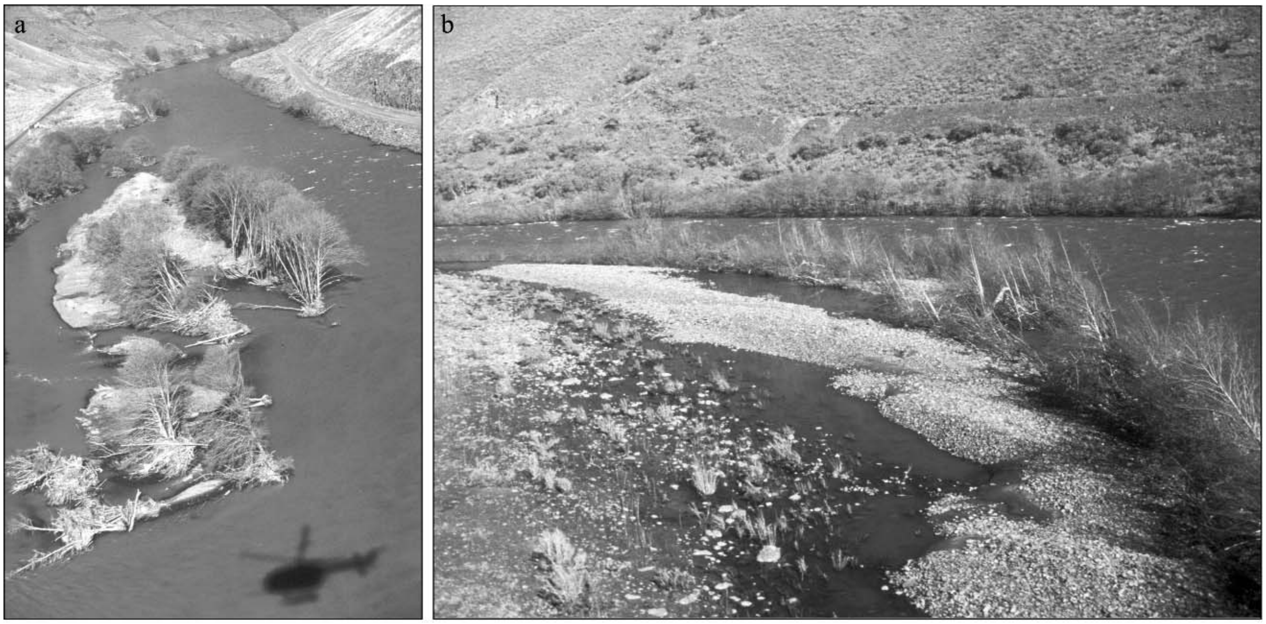

Figure 14. Photographs of effects of the February 1996 flood (February 14, 1996 photographs by H. Fassnacht). (a) View downstream of a partly eroded island with downstream gravel deposits at RM 7.1. (b) View downstream from left bank at RM 8.4 of gravel deposited on point bar in lee of riparian vegetation.

major eastern tributaries that capture flow from the dissected uplands and tablelands of the northern Ochoco Mountains and the Columbia Plain. Shitike Creek (269 km^2), the Warm Springs River (1403 km^2), and the White River (1041 km^2) together drain most of the eastern Cascade Range between Mount Jefferson and Mount Hood. From the east, Trout Creek drains 1789 km^2 of the northwestern Ochoco Mountains, and Buck Hollow (396 km^2) and Bakeoven Creeks (513 km^2) drain tablelands underlain by the Columbia River Basalt Group (Figure 1). These six tributaries account for 78% of the drainage area downstream of RM 100.1.The remaining tributaries are short and steep ephemeral channels that primarily drain the valley sides. Downstream of RM 35, there are no tributaries with drainage areas more than 50 km^2.

Even taken together, these tributaries and the few springs in the lower 60 km of the river contribute little to the mean annual flow. The average annual discharge at the Madras gage at RM 100.1 is 80 percent of the flow at the Moody gage near the Columbia confluence at RM 1.4 [*O'Connor et al.*, this volume], and close to 90 percent of late summer flow is generated upstream of the Madras gage [*Gannett et al.*, this volume]. For large floods, however, the six large tributaries entering downstream of the dams are important contributors [*Beebee and O'Connor*, this volume]. The peak discharges of the 1964 and 1996 floods increased by factors of three to four between Madras and the Columbia River confluence. The major contributors to the February 1996 peak discharge were Shitike Creek (125 m^3/s), the Warm Springs River (640 m^3/s) and the White River (ungaged in 1996, but 375 m^3/s in December 1964). These three tributaries are connected by incised and steep channel networks to high Cascade Range source areas, and during regional rain-on-snow events they rapidly deliver runoff to the mainstem Deschutes. The smaller tributaries generally flow only seasonally and during runoff events. The largest flows on the short, steep tributaries are mostly due to localized storm cells.

In addition to producing high runoff, Shitike Creek, the Warm Springs River, and the White River are situated to deliver abundant coarse sediment to the Deschutes River valley: They generate peak discharges comparable to peak discharges in the mainstem Deschutes River, and they all flow in steep and confined valleys from high-relief source areas composed of unconsolidated volcanic and glacial deposits. Sediment transport by these large tributaries into the mainstem Deschutes River has not been measured, but active islands and gravel bars at the mouth of Shitike Creek, and the large gravel fan at the Warm Springs River confluence demonstrate that these tributaries do indeed contribute significant volumes of coarse sediment into the Deschutes River during large flows. Given the paucity of coarse sediment delivered from the upper Deschutes River basin during the present geologic and hydrologic regime, it is likely that these Cascade Range tributaries contributed as much or more sediment to the lower Deschutes River as did the entire upper Deschutes River basin prior to regulation [*O'Connor et al.*, this volume].

The large eastern tributaries and the smaller tributaries that drain the valley sides also episodically contribute sediment to the valley bottom, although these events generally result from localized storm runoff rather than from regional flooding. Between RM 100 and RM 60, these tributaries primarily drain basins formed in the fine-grained sedimentary rocks of the Deschutes, John Day, and Clarno Formations, and their sediment contributions are likely to be primarily silt and clay, although lava-flow interbeds within these formations and Columbia River Basalt Group lavas contribute some gravel. This is consistent with *McClure's* [1998] observation that the surface grain size distribution of deposits at the confluence of Trout Creek was distinctly finer than most of the thirteen other sampled tributaries. Downstream of RM 60, the small tributaries are primarily formed in rocks of the Columbia River Basalt Group. These rocks produce cobble and pebble-sized clasts that are readily transported by floods and debris flows down channels draining the steep canyon slopes. In July 1995, a number of such flows brought a large volume of cobbly gravel to the valley bottom during an intense convective storm that affected the lower 30 km of the Deschutes River canyon (Steve Pribyl, Oregon Department of Fish and Wildlife, 1998 oral communication).

Tributary floods transport coarse sediment that either directly enters the Deschutes River channel, or is deposited in alluvial fans that build up at tributary mouths. There are 84 recognizable tributary fans between RM 100.1 and the Columbia River confluence, ranging in size from 0.1 hectares (ha) to the 68-ha fan at the Warm Springs River confluence [*Curran and O'Connor*, this volume]. These fans are generally composed of poorly sorted cobbly gravel deposited by debris flows and sediment-laden water flows. Some fans are built out onto alluvial surfaces that flank the Deschutes River and have little direct interaction with the channel, but several fans constrict the channel, narrowing flow and forming rapids or riffles. A recent example is the July 1995 tributary flood at Mud Springs Canyon (RM 8), which built a fan of bouldery debris out into the Deschutes River channel, constricting the channel and forming a new rapid, Washout Rapids [Figure 15; *Curran and O'Connor*, this volume]. The fans that encroach into the channel are commonly eroded by moderate Deschutes River flows, and

Figure 15. View north of the tributary fan at RM 7.8, where a July 1995 debris flow transported bouldery gravel into the Deschutes River channel, forming 'Washout Rapids.' Culverts under railroad have diameters of 4 m. June 2, 1998 photograph by J. E. O'Connor.

they are local sources of readily-mobilized coarse-grained sediment. According to Steve Pribyl (Oregon Dept. of Fish and Wildlife, oral communication, 1998), the fan emanating from the Harris Canyon tributary (RM 12) has repeatedly encroached into the Deschutes River channel over the last decade, only to be trimmed back several times by large mainstem flows. Just downstream, the apex of Harris Island has accumulated angular basalt gravel and cobbles that likely derive from the Harris Canyon fan. *Curran and O'Connor* [this volume] more fully discuss locations, sizes, and distribution of tributary fans.

SUMMARY

The present canyon and valley bottom of the lower Deschutes River reflect a long and complicated history of events and processes, encompassing temporal and spatial scales ranging from regional tectonic forces operating over millions of years to the effects of summer thunderstorms. Late Cenozoic incision of the lower Deschutes River basin, combined with stable discharges, a muted flood generation regime, and exceptionally low sediment production of the upper Deschutes River basin [*O'Connor, Grant and Haluska*, this volume] comprise a setting in which events and processes of long recurrence interval have strong influence on the valley and channel morphology of the lower Deschutes River. Moreover, many features of the present canyon and valley bottom can be attributed to individual events and circumstances. Some aspects of the processes and events that have affected the valley and valley bottom are:

1. Geologic units within canyon walls and regional deformation patterns partly control the overall morphology of the canyon and canyon bottom. The canyon is deeper where it passes through structural highs, and shallower where it crosses structural lows. Where bounded by the softer John Day and Clarno formations, extensive landsliding has widened the canyon, and the width of the valley bottom varies substantially. Where the valley is bounded by Columbia River Basalt Group, the canyon bottom is uniformly narrower and closely sheltered by steep valley walls. A distinct bedrock reach between RM 50 and RM 40 coincides with the Deschutes River flowing toward the rising limb of the Tygh Ridge anticline.
2. Regional volcanism and episodic incursion of lava flows and lahars into the Deschutes River canyon have had localized effects on valley geometry, and these events have probably resulted in rare but immense pulses of sediment into the lower Deschutes River. A large Pleistocene lahar from Mount Jefferson traveled much of the length of the canyon, leaving poorly sorted bouldery gravel deposits along the valley margins. In addition, eruptions of the Bend Pumice, Mount Mazama, and Mount Hood left localized sediment accumulations along the lower Deschutes River.
3. Between RM 100 and RM 60, Quaternary landslides have profoundly influenced valley morphology. Numerous large landslides rafted immense volumes of debris to the valley bottom, constricting the channel and forcing the river into tortuous routes through chaotically deformed terrain. Several landslides temporarily impounded the river, later to fail and send large floods downstream. The breached landslide dams and resulting flood deposits have left a persistent legacy of rapids, such as Whitehorse Rapids and Boxcar Rapids, and immobile flood bars that now constrain the channel.
4. In addition to floods from breached landslides, large floods from within and outside the Deschutes River basin have affected the valley and valley bottom of the lower Deschutes River. Between 15 and 12.7 ka, huge floods from ice-dammed glacial Lake Missoula came down the Columbia River valley and backflooded up the Deschutes River, depositing clay, silt, and sand to an elevation of 275 m on valley slopes 60 km upstream from the Columbia River confluence. Sometime during the late Holocene, probably between 7600 and 3000 yr BP, an exceptionally large flood surged through the Deschutes River canyon. This 'Outhouse flood' likely had a discharge twice that of the 2000 m^3/s discharge of the flood of February 1996. The cobble and boulder bars left by this flood have resisted erosion from subsequent floods, and in many loca-

tions, these features confine the Deschutes River channel to its present position.

5. The largest historic main-stem floods, in December 1861, December 1964, and February 1996, resulted from regional rain-on-snow events. These floods apparently had little effect on the overall morphology of the Deschutes River valley bottom. Their primary effects have been to deposit and erode fine sediment and wood on alluvial surfaces flanking the channel, and to erode edges of alluvial fans that episodically encroached into the Deschutes River channel between mainstem floods.
6. Within the present geologic and hydrologic regime, the major Cascade Range tributaries (Shitike Creek, Warm Springs River, White River) generate a substantial portion of the discharge of large flood flows. Moreover, these tributaries supply abundant coarse sediment, perhaps exceeding the gravel volume delivered from upstream. Smaller tributaries episodically deliver fine and coarse sediment to the Deschutes River and to alluvial fans that have formed on the valley bottom. In some locations, this material is a source of moderately coarse sediment that can be transported by moderate mainstem flows.

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