Earth-Science Reviews 208 (2020) 103181

Contents lists available at ScienceDirect

Earth-Science Reviews

journal homepage: www.elsevier.com/locate/earscirev

The Missoula and Bonneville floods—A review of ice-age megafloods in the Columbia River basin

Jim E. O'Connor [a,*], Victor R. Baker [b], Richard B. Waitt [c], Larry N. Smith [d], Charles M. Cannon [a], David L. George [c], Roger P. Denlinger [c]

[a] U.S. Geological Survey, Portland, OR, USA
[b] University of Arizona, Tucson, AZ, USA
[c] U.S. Geological Survey, Vancouver, WA, USA
[d] Montana Technological University, Butte, MT, USA

ARTICLE INFO

Article history:
Received 7 November 2019
Received in revised form 18 March 2020
Accepted 30 March 2020

Keywords:
Megafloods
Columbia River basin, western North America
Missoula floods
Bonneville flood
Channeled Scabland

ABSTRACT

The Channeled Scabland of eastern Washington State, USA, brought megafloods to the scientific forefront. A 30,000-km^2 landscape of coulees and cataracts carved into the region's loess-covered basalt attests to overwhelming volumes of energetic water. The scarred landscape, garnished by huge boulder bars and far-travelled ice-rafted erratics, spurred J Harlen Bretz's vigorously disputed flood hypothesis in the 1920s. First known as the Spokane flood, it was rebranded the Missoula flood once understood that the water came from glacial Lake Missoula, formed when the Purcell Trench lobe of the last-glacial Cordilleran ice sheet dammed the Clark Fork valley in northwestern Idaho with ice a kilometer thick. Bretz's flood evidence in the then-remote Channeled Scabland, once widely seen and elaborated by the 1950s, eventually swayed consensus for cataclysmic flooding. Missoula flood questions then turned to some that continue today: how many? when? how big? what routes? what processes?

The Missoula floods passed through eastern Washington by a multitude of valleys, coulees and scabland tracts, some contemporaneously, some sequentially. Routings and their timing depended on the positions of various lobes of the multi-pronged Cordilleran ice sheet and the erosional development of the channels themselves. The first floods mostly followed the big bend of Columbia valley looping through north-central Washington. But the south-advancing Okanogan ice lobe soon blocked that path, forming long-lasting glacial Lake Columbia in the impounded Columbia valley. Missoula floods into this lake were diverted south out of the Columbia valley and into eastern Washington coulees and scabland tracts. At least four floods entered Moses Coulee, but then as the Okanogan lobe advanced over and blocked the head of that coulee, more eastern paths took the water, including Grand Coulee and the Telford-Crab-Creek and Cheney-Palouse scabland tracts. Flood routing also depended on the erosion of the coulees. At some point, headward erosion of upper Grand Coulee lowered the divide saddle between the west-running Columbia valley and the deep and wide Grand Coulee heading southwest. Still uncertain is when this happened and the consequences with respect to the stage and extent of glacial Lake Columbia and to flood access to the other, higher, flood routes. Downstream, all flood routes converged into Pasco Basin, flowed through Wallula Gap and the Columbia River Gorge into the Pacific Ocean, following submarine canyons and depositing sediment layers on abyssal plains.

Stratigraphic studies indicate dozens—likely more than a hundred—of separate Missoula floods during the last glacial period. Over the length of the flood route, backwater areas and depositional basins preserve multiple flood beds, many of which are separated by signs of time, including volcanic ash layers and soil development in subaerial environments; and varve-like beds and pelagic mud layers in lacustrine and marine settings. Evidence also comes from the glacial Lake Missoula basin, where stratigraphy indicates dozens of filling and emptying cycles. Varve counts in conjunction with radiocarbon dating and paleomagnetic secular variation show the repeated filling-and-release cycles of glacial Lake Missoula had intervals possibly as long as 100 years early in the lake's history but diminished to just one or two years for the last few floods. This behavior accords with jökulhlaup-style floods released by subglacial drainage from a self-dumping ice-dammed lake. Not yet clear is whether such a mechanism applies to all the floods or if some emptied more cataclysmically as hypothesized by some.

* Corresponding author.
E-mail address: oconnor@usgs.gov. (J.E. O'Connor).

http://dx.doi.org/10.1016/j.earscirev.2020.103181
0012-8252/Published by Elsevier B.V.

Radiocarbon dating of sparse organic materials remains key to defining flood chronology but has been lately bolstered by analyses of terrestrial cosmogenic nuclides and optically stimulated luminescence. Varve counts and paleomagnetic secular variation studies help to define durations and intervals represented by sequences of flood beds. The ~16 ka Mount St. Helens Set S tephra is commonly interbedded within flood deposits, enabling correlation of deposits among sites. Tephra from the 13.7–13.4 ka eruption of Glacier Peak overlies all glacial Lake Missoula and Missoula flood deposits, defining an end time. Overall conclusions are that glacial Lake Missoula was extant and producing floods for at least 3–4 ky during 20–14 ka. At least ~75 floods preceded Mount St Helens Set S, followed by 30 or more after the tephra fall. Most floods entered glacial Lake Columbia, impounded by the Okanogan lobe, for 2–5 ky between about 18.5 and 15 ka. Glacial Lake Columbia outlived Lake Missoula by >200–400 yr but may have been born later since at least one flood came down the Columbia valley before the Okanogan ice lobe blocked the Columbia valley at 18.5–18 ka. The maximum extent of the Okanogan and Purcell Trench lobes, many Missoula floods, substantial erosion of upper Grand Coulee, and the widespread tephra falls from Mount St. Helens eruptions all happened about 17–15 ka. People, in the area since 16.6–15.3 ka, almost certainly witnessed the last of the Missoula floods and later large floods from other ice-dammed lakes in the Columbia River basin.

Quantitative flow analyses give peak discharge estimates and support understanding of erosional and depositional processes. The first flow assessments were simple cross-section calculations but recent assessments employ two-dimensional hydrodynamic models. The general finding is that emplacement of the maximum stage evidence requires about 20 million m^3/s near the Lake Missoula outlet and about 5–15 million m^3/s through Wallula Gap and downstream in the Columbia River Gorge. These hydraulic analyses raise still-unresolved questions regarding canyon erosion and possible additional water sources.

The large Pleistocene Bonneville flood entered the Columbia River system from the southeast from pluvial Lake Bonneville, the Pleistocene predecessor to Great Salt Lake in the eastern Great Basin. During the last glacial, the lake basin filled, covering >50,000 km^2 with 10,400 km^3 of water before reaching its maximum possible stage governed by Red Rock Pass, the lowest divide separating the basin from the Snake River basin to the north. The overtopping lake rapidly incised 108–125 m into the Red Rock Pass outlet, spilling half of its total lake volume. G.K. Gilbert described the essential sequence in the 1870s, but the flood was mostly forgotten until the late 1950s when Harold Malde linked the spectacular scabland topography and bouldery "melon gravel" on the Snake River Plain to the Lake Bonneville overflow. The Bonneville flood appears to have been a singular event at about 18 ka. No evidence of multiple or pre-last-glacial spillovers has yet been found. Its total volume was about twice that of a maximum Lake Missoula flood yet its peak discharge was ~1 million m^3/s, less than a tenth of the largest Missoula floods. Its comparatively simple flow path and much steadier flow make the Bonneville flood ideal for new studies of erosional and depositional processes.

At least two floods seem to have passed down the Columbia valley after the last of the Missoula floods, including a large flood about ~14 ka likely from cataclysmic demise of the thinning Okanogan ice lobe dam impounding glacial Lake Columbia. Floods from earlier glacial ages left scant yet clear evidence in the Channeled Scabland and Columbia valley. But their source, timing, and magnitudes are little understood. Some deposits are paleomagnetically reversed, thus older than ~800 ka. Last-glacial floods and perhaps older ones affected the Snake River Plain, some likely sourced in lakes dammed by alpine glaciers in central Idaho.

1. Introduction

1.1. Ice-age floods of the Columbia

Megaflood science gained early traction in the Columbia River basin of western North America (Fig. 1). The terrain has nourished and stretched thinking about landscape formation and geologic processes for more than 100 years, led by prominent geologists such as G.K. Gilbert, J Harlen Bretz, Richard Foster Flint, and Harold Malde. It has stirred controversy by inspiring new ideas and challenging old ones. It still gives rise to fresh ideas, now being applied to megaflood-carved terrains worldwide and on other planets (Baker, 2009, 2020). Here we review the ice-age megafloods of the U.S. Pacific Northwest, their history of discovery, current understanding, and outstanding questions. The focus is the late Pleistocene Missoula floods and the spectacular Channeled Scabland, but similar in extent and geomorphic record is the Bonneville flood from pluvial Lake Bonneville in the Snake River drainage.

We begin with a short history of flood research in the Columbia River basin. We then describe the two major but overlapping megaflood landscapes: the Bonneville flood chiefly along the Snake River and the Missoula floods and the Channeled Scabland within the Columbia drainage. We summarize their causes, routes, and timing, as well as new questions and opportunities. We conclude with brief comments on other large last-glacial floods in the region, floods from prior ice ages, and the possibility that people witnessed some of the Columbia Basin megafloods.

1.2. A difficult birth

The first written recognition of cataclysmic flooding in the U.S. Pacific Northwest may reside in G.K. Gilbert's field notes for Aug. 16, 1876, inked as he approached the north margin of pluvial Lake Bonneville in northern Utah (Hunt, 1982; O'Connor, 2016). Arriving at the Red Rock Pass outlet, Gilbert wrote "The red rock is only one of a number that are here *exceptionally* bared, in testimony to the stream that has washed them in degrading this pass." His 1890 monograph states "the outpouring was a veritable débâcle" (Gilbert, 1890, p. 175). Yet the Bonneville flood was lost for some 80 years until resurrected by Harold Malde (1960, 1968), in part inspired by the then-recent consensus that the Channeled Scabland of eastern Washington was indeed the product of huge Pleistocene floods.

Better known is the "Great Scablands Debate" featuring J Harlen Bretz (Baker, 1978a, 1981, 2008, 2009; Gould, 1978, 1980). Though not a paradigm shifter like plate tectonics, it exemplifies how science really works. Instead of colorless collecting of facts and observations and unbiased working hypotheses, it involves people—sometimes gangs—with personalities, agendas, and egos. This was, and maybe still is, the case for the Channeled Scabland.

In 1922, Bretz, a 39-year-old assistant professor at the University of Chicago, began a series of eight field seasons in eastern Washington State. His first paper (Bretz, 1923a) described eroded scabland tracts, streamlined loess hills, and stratified gravel, ascribing them to a "great flood" (p. 588). His second paper later that year elevated the flood to a "debacle" (Bretz, 1923b, p. 649). But unlike Gilbert's débâcle 33 years earlier, which

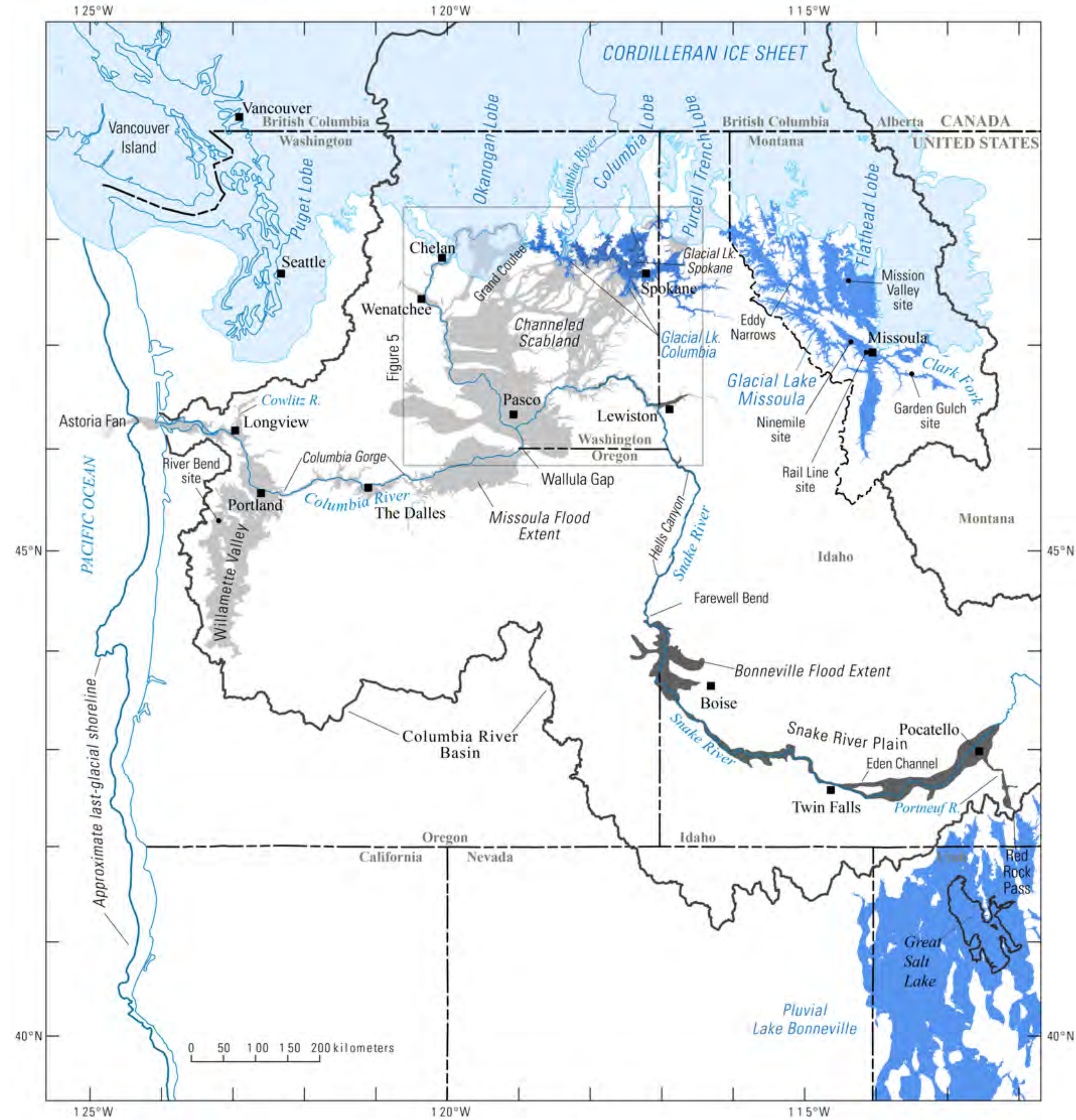

Fig. 1. Overview map of U.S. Pacific Northwest megaflood landscape. Bonneville flood inundation modified from O'Connor (1993); Missoula flood inundation courtesy of Daniel Coe, Washington Geological Survey; maximum extent of Cordilleran ice sheet and adjacent glaciers based on Waitt and Thorson (1983), Booth (1986), Clague (1989), Locke (1995), Carrara et al. (1996), Smith (2004), Riedel (2017), and unpublished mapping by Ralph Haugerud, U.S. Geological Survey. Extent of Bonneville flood downstream of Lewiston is within Missoula flood extent.

remained hardly noticed (O'Connor, 2016), Bretz's advocacy of what he at first called the Spokane Flood provoked weighty resistance.

Most aggressive were the scientists of the U.S. Geological Survey (USGS), culminating in the 1927 meeting of the Geological Society of Washington (Bretz, 1927a; Baker, 2008). Senior USGS scientists such as William C. Alden (head of the Pleistocene Division), James Gilluly, Oscar Meinzer, and Edwin T. McKnight ganged up on Bretz. Objections were many; some minor and easily refuted, but tougher were (1) the yet undefined source of water and (2) the rapid and widespread erosion by a single flood. The first point was valid, though the likely flood source—glacial Lake Missoula—was already pointed out by USGS geologist Joseph T. Pardee (Baker, 1995; Baker, 2008, p. 39). But objections focused on the outsized scale of the flood and its effects were dogmatic rather than scientific. The relative new science of geology, and especially the cautious USGS scientists at the meeting, worked from a notion of uniformitarianism conflating the valid application of modern process studies for understanding the past with a narrower presumption that only *observed* modern processes and rates applied to past events (Baker, 1998). This view denied cataclysms like Bretz's Spokane Flood.

The controversy continued through the 1930s though Bretz had mostly moved to other topics. Richard Foster Flint, prominent on the Yale faculty, was loudest, but others chimed in alternative hypotheses, such as Edwin T. Hodge (Oregon State University and University of Oregon) and Ira S. Allison (Oregon State University). Flint, later known as the "Pope of the Pleistocene" (Goudie, 2008, p. 463), may have had an axe to grind: Bretz had sat on Flint's 1925 University of Chicago Ph.D. examining committee and voted to fail Flint (Baker, 2008, p. 41; J H. Bretz written communication with V.R. Baker). Flint, whatever the motive, went after the catastrophic nature of Bretz's hypothesis, proposing instead that the scabland channels were cut slowly by "leisurely streams with normal discharge" and that Bretz's flood bars were simply unpaired terraces (Flint, 1938, p. 471). Flint (1938, p. 492) also called upon the scabland flanking the Snake River Canyon in southern Idaho, now known eroded by the Bonneville flood (Malde, 1968), as evidence that such terrain did not require "unusual floods." Hodge (1934) offered a complicated story of long-term drainage integration and multiple glaciations such that the "progress of the work was slow and involved many repeated attacks… over hundreds of thousands of years…[by] common and usual methods that one may see at work today…." Allison (1933, p. 676–677) acknowledged evidence of high water stages but also objected to the catastrophic aspect. He suggested a downstream blockage diverting water into upland channels, thereby providing a mechanism that "does not require a short-lived catastrophic flood but explains the scablands, the gravel deposits, diversion channels, and divide crossings as the effects of a moderate flow of water, now here and now there, over an extended period. It thus removes the flood from the "impossible" category." Catastrophes remained the enemy.

Debate was still strong in 1940 when the American Association for the Advancement of Science met in Seattle. A series of papers in a session on the "Quaternary Geology of the Pacific Northwest" piled on to the objections for a catastrophic flood. But late in the session, USGS geologist Joseph Pardee presented a paper titled "Ripple marks (?) in glacial Lake Missoula." He described evidence for extraordinary currents in the basin of glacial Lake Missoula, at its largest, a 2500 km^3 lake dammed 650 m deep by a lobe of the Cordilleran Ice Sheet. These currents he ascribed to rapid drainage following failure of the ice dam. Left unsaid at the meeting and in the ensuing paper (Pardee, 1942) was the fate of all that water. The implications were not lost, however, on participants of a post-meeting day in the scablands with Flint. Much of Flint's evidence for "leisurely streams" was instead interpreted by trip participants as cataclysmic flood features (Baker, 2008; Howard Meyerhoff, trip participant, 1978 written communication with V.R. Baker).

Bretz cemented the cataclysmic flood hypothesis with a season of fieldwork in the Channeled Scabland in 1952 (Bretz et al., 1956). Bretz, with colleagues H.T.U. Smith and George Neff, examined many new exposures, assessing possible explanations. Adding new and compelling support were aerial photographs showing broad fields of giant current dunes on gravel bar surfaces—like those left in the basin by draining glacial Lake Missoula. The rippled bars could not be terraces. With the link to glacial Lake Missoula now established, the floods became the Missoula floods. By this evidence most of the geologic community was swayed. The Missoula floods carved the Channeled Scabland, making it into most textbooks (including a sentence in Flint's 1971 "Glacial and Quaternary Geology"). Bretz won the Geological Society of America's 1979 Penrose Medal. Channeled Scabland debates moved onto other questions such as when? how many? how big?

Why did the scabland debate last 40 years? Not because of lack of evidence. Most pertinent observations were made and reported by 1930—although aerial photography and discovery of the giant current dunes added new information in 1956. Bretz was ornery; Flint was powerful and may have held a grudge. Catastrophic floods were "impossible" to ruling parties of the North American Pleistocene.

Nevertheless a flood cause for the scablands ultimately prevailed. This in turn guided Malde's (1960, 1968) identification of Bonneville flood features downstream of Red Rock Pass. More work in the scablands a few years later (Baker, 1973) at the time of the Mariner 9 photographs of Mars inspired recognition of Martian megafloods (Baker and Milton, 1974; Baker, 1982). Cataclysmic floods are now broadly embraced as a universal shaper of landscapes (Baker, 2013).

2. The Bonneville flood

> For a time the outpouring was a veritable débâcle
>
> [Gilbert (1890, p. 175)]

The first discovered ice-age flood in the U.S. Pacific Northwest was the largest in terms of total flood volume—the Bonneville flood

2.1. Pluvial Lake Bonneville

Pluvial Lake Bonneville was the largest lake in western North America to release a megaflood during the last glacial maximum. Lake Bonneville, Pleistocene predecessor to Great Salt Lake, filled a closed tectonic basin west of the Wasatch Range with more than 10,000 km^3 of water, covering 52,000 km^2 of western Utah (Table 1; Fig. 1). The history of pluvial Lake Bonneville, first comprehensively treated by Gilbert (1890), is thoroughly reviewed in the recent volume edited by Oviatt and Shroder (2016). During its most recent growth cycle beginning about 30 ka, the lake haltingly rose until about 18 ka, when it reached the lowest divide saddle separating it from the Snake River basin to the north. At the crest of an alluvial fan astride the head of Marsh Creek valley just north of Red Rock Pass, the lake stabilized at its maximum Bonneville level of 1552 m above sea level[1] and formed the prominent Bonneville shoreline (Oviatt and Jewell, 2016). Overflow at the outlet initiated erosion and catastrophic outflow (Gilbert, 1890)—the débâcle of the Bonneville flood. Lake level dropped, stabilizing 108–125 m lower upon reaching bedrock at Red Rock Pass, releasing about 5135–5320 km^3 of water (Miller et al., 2013, p. 360; Adams and Bills, 2016, p. 160; Abril-Hernández et al., 2018, p. 7). The lake remained close to this rock-controlled Provo level for about 3 ky before shrinking to closed-basin conditions when climate dried at the end of the Pleistocene.

2.2. The flood route

Effects of the Bonneville flood are traceable for more than 1100 km down the Snake River drainage (Fig. 2). From Red Rock Pass it flowed north down Marsh Creek into the Portneuf River valley (Fig. 3a, b). An immense boulder fan marks the flood's entrance onto the Snake River Plain near Pocatello before gathering into canyons of the Snake River (Trimble and Carr, 1961). The flood followed the vast and incised Snake River Plain west for 500 km, descending steeply through narrow canyon segments, particularly the 100 km from east of Twin Falls to Hagerman

[1] All reported elevations are relative to the National Geodetic Vertical Datum of 1988 and are modern positions. Many Pleistocene landform elevations were affected by isostatic deformation owing to loading by water and ice.

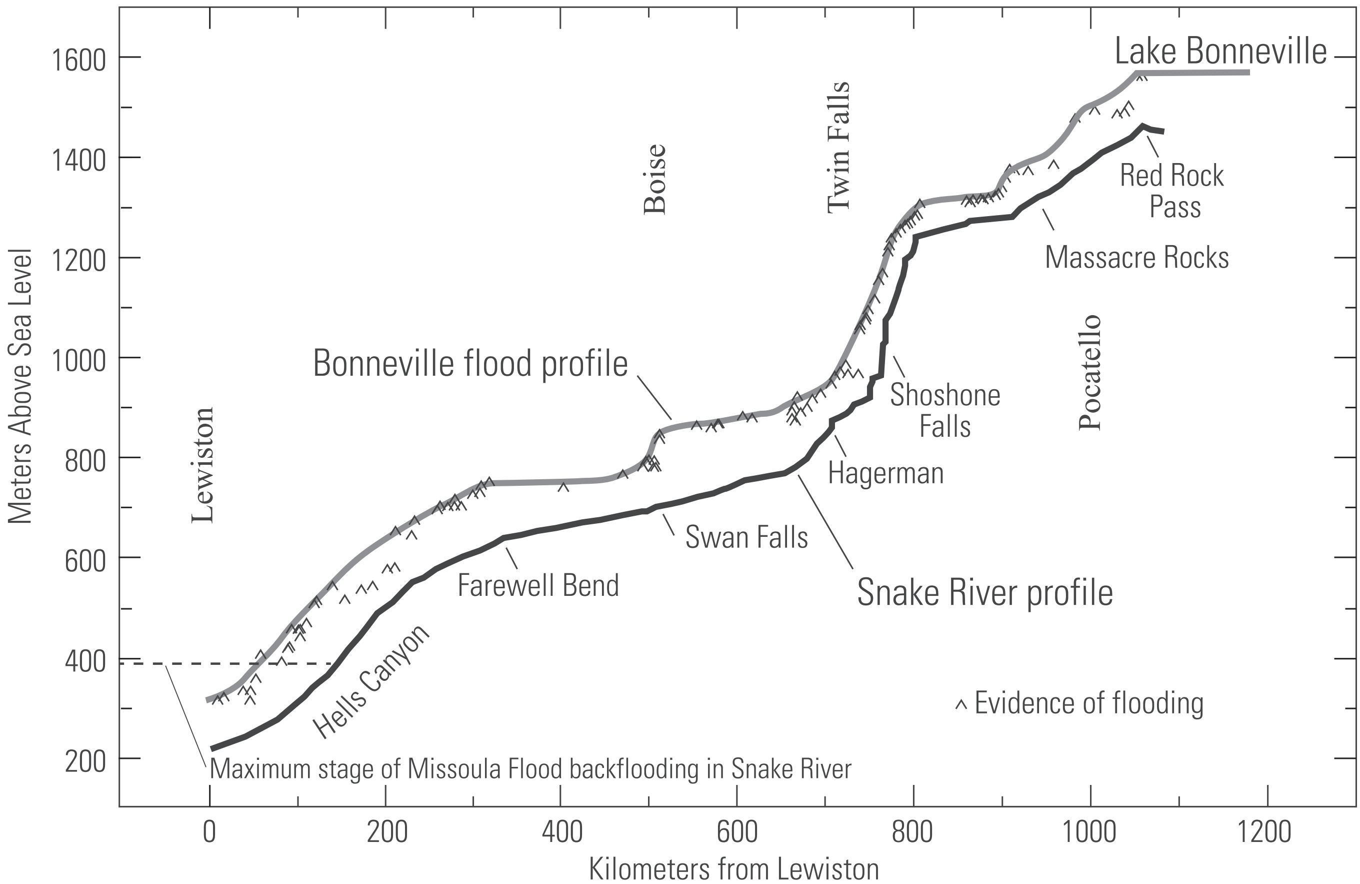

Fig. 2. Profile of the Bonneville flood. Modified from O'Connor (1993, p. 11). Kilometer 0 equivalent to Snake River Mile 140.

(Malde, 1968; O'Connor, 1993). In this steep reach the canyon filled and the flood split, most flow broadly crossing the uplands north of the Snake River, reentering at several spectacular cataracts and blind canyons eroded into the north canyon rim near Twin Falls (Fig. 3c). Deposits of "Melon Gravel" fill canyon expansions and side valleys (Fig. 4; Malde and Powers, 1962, p.1216–1217).

After spreading out at the west end of the Snake River Plain, the flood followed the Snake River north and at Farewell Bend entered confined valleys leading to the deep and constricted Hells Canyon (Figs. 2 and 4c). Here the flood both eroded and deposited gravel more than 175 m above present river level (Fig. 4c; O'Connor, 1993). Bonneville flood deposits extend north to Lewiston (Fig. 4e), and follow the deep Snake River canyon west. The most downstream Bonneville flood deposits yet found are 40 km west of Lewiston at Snake River Mile 103 (Kurt Othberg, Idaho Geological Survey, unpublished mapping). Downstream deposits of the Bonneville flood become progressively buried by Missoula flood beds (Fig. 4e), masking their distribution, particularly below the Palouse River confluence at Snake River Mile 60, where huge bars of Missoula flood gravel partly fill the Snake River canyon.

2.3. A singular flood

The release of the upper 108–125 m of pluvial Lake Bonneville was a one-time event. The distinctly weaker Marsh Creek alluvial sediment compared with the underlying bedrock led to run-away erosion of the outlet until stopped by the harder rock (Gilbert, 1890, p. 175, Abril-Hernández et al., 2018; Garcia-Castellanos and O'Connor, 2018). The lake remained at the Provo level until evaporation exceeded flow into the basin ~3 ky later, starting closed-basin conditions culminating in modern Great Salt Lake. No evidence has yet been found of overflows into the Snake River basin during earlier lake cycles—neither flood evidence in the Snake River Plain nor older high shorelines in the Bonneville basin. Bear River, draining mountain ranges east of Lake Bonneville, diverted into the Bonneville basin at 60–50 ka, drainage that accounts for most inflow into the basin (Pederson et al., 2016); hence the last-glacial lake cycle was likely the highest ever (Bright, 1963).

2.4. Flood timing

The age of the Bonneville flood is most securely known by the radiocarbon chronology of pluvial Lake Bonneville. Oviatt's (2015) analysis shows Lake Bonneville reaching peak level at or after 18.4 ka[2] but dropping to the Provo level before 17.1 ka. He concludes "a reasonable age near the middle of the overlapping range is 18 ka with an uncertainty of several hundred calibrated years." Miller (2016, p. 135–137) interprets recent

[2] All radiocarbon ages are reported in 1000s of calendar years before present, ka, referenced to 1950 AD. Unless otherwise noted, if not calibrated in the source literature we have calibrated specific age results using the IntCal 13 radiocarbon age calibration curve (Reimer et al., 2013) implemented by OxCal 4.3 (Bronk Ramsey, 2009; https://c14.arch.ox.ac.uk/OxCal/OxCal.html) and report the resulting two-standard-deviation confidence range of the calibration of the radiocarbon age and its original counting-error uncertainty.

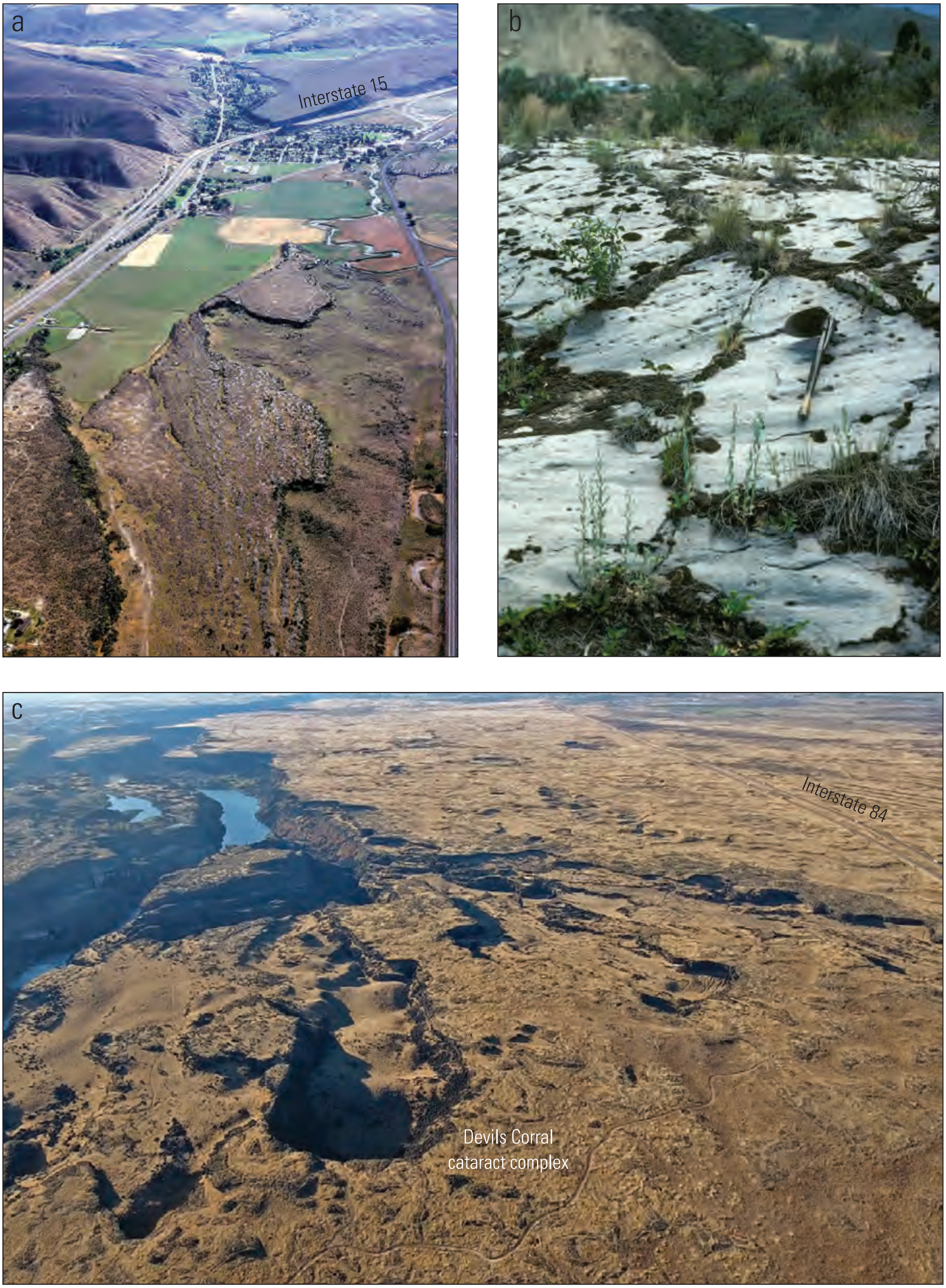

Fig. 3. Erosion by the Bonneville flood. (a) Aerial view east and upstream in the Portneuf River valley, floored by flood-eroded basalt flow. Photograph by Jim E. O'Connor. (b) Ground view of fluted and polished Portneuf Basalt near the scabland in (a). Shovel handle 0.5 m. Photograph by Jim E. O'Connor. (c) Aerial view west and downstream from near Twin Falls, showing scabland and cataracts of the Devils Corral area carved by flow from the Eden Channel reentering the Snake River Canyon near Snake River Mile 617. Photograph by Bruce Bjornstad.

Fig. 4. Deposits of the Bonneville flood. (a) View downstream of immense boulder bar partly filling 1.5 km wide canyon downstream of the Swan Falls constriction near Snake River Mile 457. (b) Close view of bedload on the bar in (a). Many clasts have long diameters exceeding 10 m. (c) View north (downstream) in Hells Canyon near Snake River Mile 227, showing immense point bar (with large Ponderosa Pines along river edge) and flood-swept valley slope to nearly 200 m above. (d) Sand and fine gravel deposited from suspended load in an eddy bar east of Hagerman, Idaho. Pick is 0.7 m long. (e) Horizontally bedded Missoula Flood deposits on top of foreset bedded Bonneville Flood gravel in the Tammany Creek valley near Lewiston, Idaho, Snake River Mile 146. Here about 20 Missoula flood beds were left by that number of floods backing up the Snake River (Waitt, 1985a). The foresets in the Bonneville Flood gravel dip up Tammany Creek valley, indicating up-valley flow as the bar formed. Photographs by Jim E. O'Connor

Provo shoreline age information as indicating an 18.3–18.2 ka age for the Bonneville flood. These ages are consistent with some cosmogenic dates from flood-transported boulders (Amidon and Clark, 2015) and flood-eroded basalt (Lamb et al., 2008, 2014; Amidon and Farley, 2011). For example, cosmogenic ^{3}He and ^{21}Ne ages of five flood-transported basalt clasts 50 km downstream of Red Rock Pass give a combined result of 17.5 ± 0.9 ka (Amidon and Clark, 2015).

2.5. Flood magnitude and duration

Estimates for the size of the Bonneville flood go back to Gilbert's (1890, p. 177) assessment that it was "the flood volume of the Missouri" (Table 2)—about 20,000 m^3/s in Gilbert's time (O'Connor, 2016). Discharge estimates have since escalated, peaking at about 1 million m^3/s by O'Connor's (1993) assessment of peak discharge near the outlet. This estimate, as well as Jarrett and Malde's (1987) determination of 0.7–1.0 million m^3/s farther downstream near Swan Falls are both consistent with recent hydrodynamic modeling of flow through the eroding outlet that indicates peak outflow of 0.85 million m^3/s (Abril-Hernández et al., 2018). This new modeling also indicates outflow at Red Rock Pass exceeded 0.5 million m^3/s for ~17 days. Lapôtre et al. (2016) relate waterfall erosion by block toppling to the geometry of amphitheater-headed canyons—like the cataracts of the Devils Corral complex shown in Fig. 3c—to predict formative flow conditions. For each of two such canyons along the Bonneville flood route they (p. 1249, 1252) estimate canyon-head peak discharges of about 5000 m^3·s^{-1} and a duration of 30 days. Because these canyons carried just a small portion of the overall flood volume, these values may be consistent with the much larger discharges accounting for the total flow volume.

2.6. Bonneville flood questions and opportunities

Because it was a single flood chiefly following a single-channel route, the Bonneville flood provides a superb opportunity for understanding megaflood processes. High-resolution digital topography in conjunction with high-resolution hydrodynamic modeling and fine-scale mapping of flood features could vastly improve understanding of the linkages between flow conditions and erosional and depositional processes. As listed by O'Connor (2016), such analysis could address megaflood questions and issues of broad interest:

- What are the thresholds and processes of bedrock-channel erosion for different rock types and different hydrodynamic environments? This question has been tackled for the Red Rock Pass outlet (Garcia-Castellanos and O'Connor, 2018) and for some of the amphitheater-headed canyons (Lapôtre et al., 2016) but could be assessed at other locations.
- What situations lead to canyon incision instead of widening? And at what rates?
- How do alcoves and cataracts form and develop in the layered basalt flows? The toppling mechanism proposed by Lapôtre et al. (2016) is one plausible explanation, but others might be viable as well.
- What are the main controls on sediment transport and deposition? What is the fate of flood-transported sediment? How does deposit stratigraphy relate to flow dynamics?

Besides such general megaflood process questions, topics more specific to the Bonneville flood merit further study. These include analysis of the processes incising the canyon near Massacre Rocks in the eastern Snake River Plain (Fig. 2), probably formed entirely during the flood (Scott et al., 1982). Similarly, the narrow canyon reach east of Twin Falls, now descending steeply by tall drops and knickpoints like Shoshone Falls, may have largely or entirely formed during the flood. Farther downstream and in a very different geologic environment is steep and narrow Hells Canyon. How much did the Bonneville Flood contribute to its present morphology and depth?

3. The Missoula floods

> To understand the scabland one has to throw away textbook treatments of river work.
>
> [J. Hoover Mackin, as quoted in Bretz et al. (1956, p. 960)]

The Missoula floods may be the best known megafloods on Earth. Their early discovery and ensuing controversy, the exposure and accessibility of spectacular flood features in arid eastern Washington, and the reams of literature—books, scientific journals, park brochures, creationist manifestos—has inspired all types of investigation, exploration, and explanation. The Channeled Scabland continues to be a source of knowledge and amazement regarding planetary landscapes. The flood geography and geology are complex, however, derived from diverse interactions among the Cordilleran ice sheet, ice-dammed lakes, and available flow pathways through the rumpled Pacific Northwest landscape. We first describe the geography, features, and key interactions for the glacial lakes and various flood routes, and then summarize current understanding of timing, the number, and sizes of floods. Along the way, we identify key uncertainties and questions needing more study.

3.1. The Cordilleran ice sheet and ice-dammed lakes

During the last glacial period, ice from the Cordilleran ice sheet advanced south from the Canadian Coast Mountains and Rockies in lobes channeled down mountain valleys. This set the stage for the Missoula floods by deranging, blocking, and augmenting drainage (Figs. 1 and 5; Waitt et al., 2016). The ice lobes of most influence were the Purcell Trench lobe that dammed glacial Lake Missoula, the Columbia lobe descending the Columbia valley and blocking the Spokane valley; and the Okanogan lobe that blocked the Columbia River at its big bend between Grand Coulee and Chelan.

3.1.1. Purcell Trench lobe and glacial Lake Missoula

A thick ice lobe streamed south from the Rocky Mountains, following the Purcell Trench (Figs. 1, 5 and 6). The lobe entered and filled the deep basin now occupied by Lake Pend Oreille, its southern edge abutting the steeply rising Bitterroot Range to the south, blocking the Clark Fork River from its northwestward course. Near where the Clark Fork River enters Lake Pend Oreille, the upper limit of ice was above 1350 m (Waitt et al., 2016), more than 700 m above the modern valley bottom and implying ice a kilometer thick over the 350-m-deep (Fields et al., 1996) basin of Lake Pend Oreille. At its maximum, an ice lobe branched southeast, pushing southeast and up the narrow Clark Fork valley for perhaps 80 km (Breckenridge et al., 1989, p. 15–16).

The resulting ice-dammed glacial Lake Missoula was the source of the Missoula floods. One fourth the volume but twice the depth of pluvial Lake Bonneville (Table 1), Lake Missoula formed (and reformed) in the Clark Fork River basin of northern Idaho and western Montana. Hemmed in by tall valley walls and fed by Cordilleran ice sheet lobes and alpine ice to the east, the water rose behind the ice dam, filling the deep and tortuous valleys of the Clark Fork River and the tributary Flathead River (Pardee, 1910, 1942). Once rising high enough to destabilize the blocking ice, drainage ensued, many times catastrophically. At its maximum elevation of 1295 m, evident from shorelines etched into Mount Jumbo just east of Missoula, the lake covered 10,700 km^2, held about 2500 km^3 of water, and was 650 m or more deep at the ice dam (Figs. 7a and 8). From stratigraphic records in the Missoula lake basin (Chambers, 1971, 1984; Waitt, 1980, 1985a; Smith, 2006; Hanson et al., 2012), and downstream (Glenn, 1965; Waitt, 1980, 1984, 1985a; Atwater, 1984, 1986, 1987;

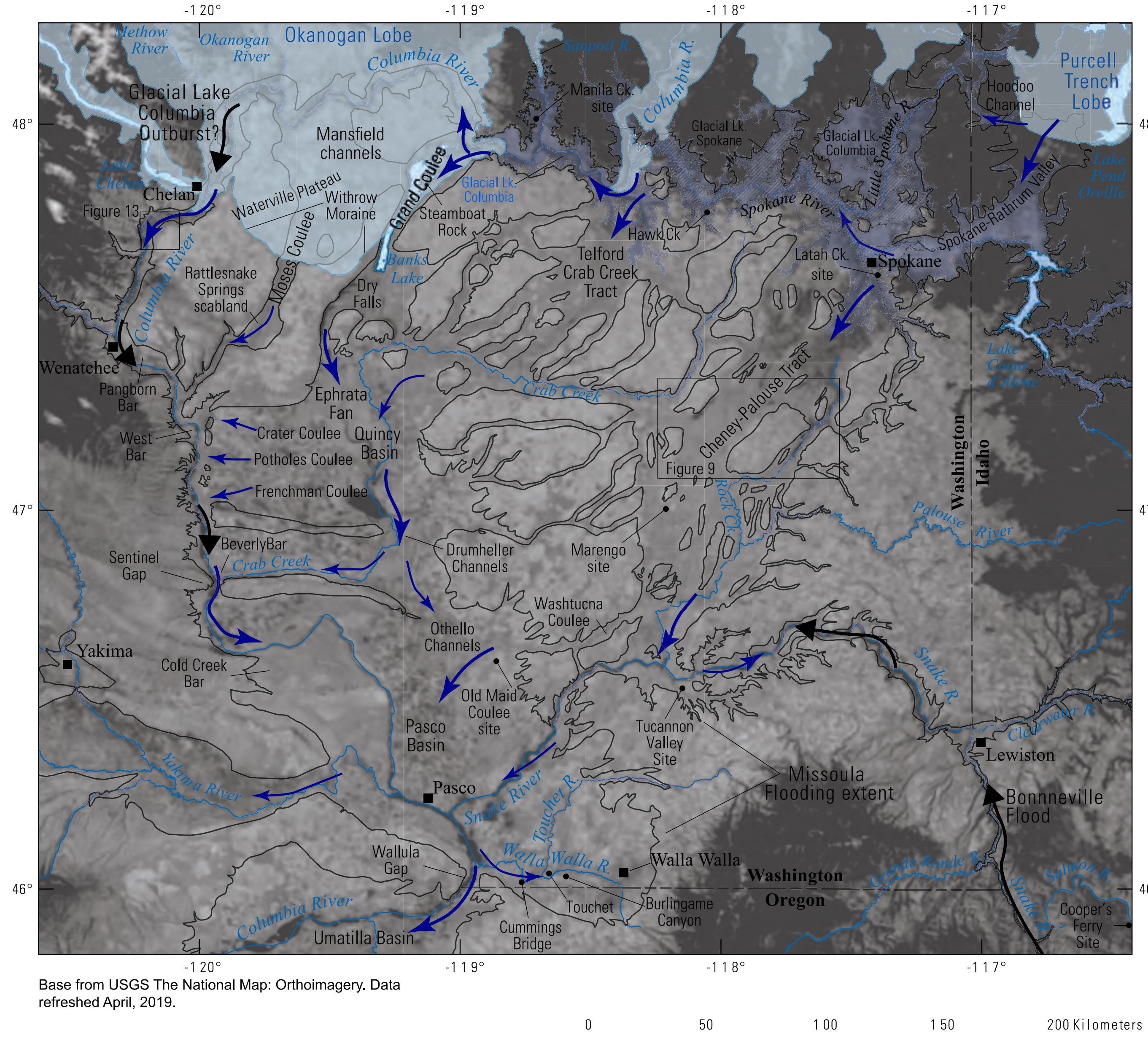

Fig. 5. Map of Channeled Scabland showing approximate last-glacial-maximum extent of Cordilleran Ice Sheet, major flood pathways, and approximate Missoula flooding extent. Glacial Lake Columbia depicted at maximum plausible 730 m elevation; glacial Lake Spokane at 600 m.

Smith, 1993; Clague et al., 2003; Benito and O'Connor, 2003), lake releases apparently recurred dozens of times; each time followed by renewed damming by the south-flowing ice lobe. Not clear is how many times glacial Lake Missoula approached it maximum level, but stratigraphy at Garden Gulch, Montana, shows 7–12 late Pleistocene lake-level fluctuations above 1170 m, representing at least ~65% of its maximum volume (Smith, 2017; Smith et al., 2018). The Purcell Trench lobe apparently reoccupied the Lake Pend Oreille basin after the last catastrophic release of glacial Lake Missoula, rimming the southern end of Lake Pend Oreille with an uneroded moraine (Smyers and Breckenridge, 2003, p. 13).

Calibrated radiocarbon dates[2] showing the growth and retreat of the Purcell Trench lobe of the Cordilleran Ice Sheet put the advancing ice lobe still north of the Canadian border at ~20–22 ka and its melting edge back north of the border by about 13 ka, thus constraining its damming of glacial Lake Missoula to between 22 and 13 ka (Fulton, 1971; Clague, 1980; Waitt, 2016). These dates are consistent with terrestrial cosmogenic nuclide (TCN) dating of five moraine boulders on a prominent near-maximum lateral moraine that give ages between 14.7 ± 1.2 and 16.6 ± 1.5 ka, averaging 15.7 ± 1.3 ka (Breckenridge and Phillips, 2010; age recalculated by Balbas et al., 2017, Table DR2)[3]. A TCN age of 14.3 ± 1.2 ka (FAR04, Balbas et al., 2017, Table DR2) from a boulder in the area of till and outwash (Lewis et al., 2002) bordering the southwest

[3] Individual TCN ^{10}Be ages from Balbas et al. (2017) are reported as ages ± external uncertainty, as listed in their Table DR2; site averages are mean ± standard error as reported in their Figure DR1. Similarly, TCN ^{36}Cl results from Keszthelyi et al. (2009) are reported as ages ± external uncertainty ("Error 2" in their Table 2, p. 861)

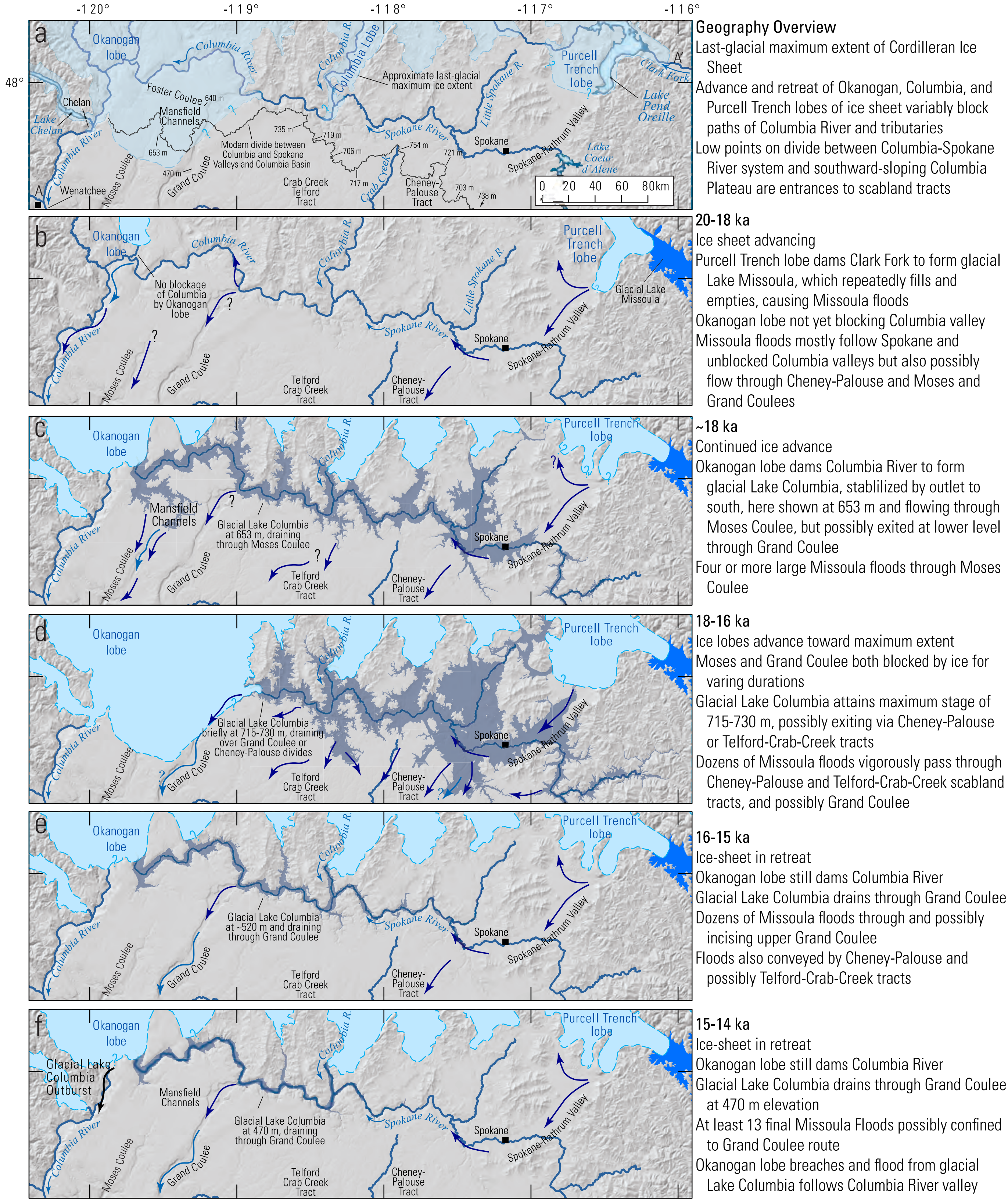

Fig. 6. Hypothesized Missoula-flood scenarios involving interactions with glacial Lake Columbia and the major flood pathways. Modified from Waitt et al. (2016) and Balbas et al. (2017). Timing approximate and summarizes information presented in text. Not shown are possible interactions involving the Columbia lobe blocking the Spokane River and resulting glacial Lake Spokane. (a) Key geographic elements. A and A' show end-points of profile of Fig. 8. (b) Glacial Lake Missoula forms and first Missoula flood(s) pass down Spokane River and Columbia River valleys. (c) Okanogan lobe blocks Columbia Valley, impounding glacial Lake Columbia and shunts at least four Missoula floods south into Moses Coulee (Waitt, 1985, 2016). (d) Maximum Okanogan lobe ice extent, blocking entrance to Moses Coulee and possibly Grand Coulee, and raising glacial Lake Columbia to ~715–730 m. (e) Retreated Okanogan lobe still blocks Columbia River,

maintaining glacial Lake Columbia, but now lowered and stabilized by a Grand Coulee outlet at 510–550 m level (map shows 520 m). (f) Possible last-glacial incision of upper Grand Coulee threshold to modern threshold elevation of 470 m lowers glacial Lake Columbia. Glacial Lake Columbia persists for at least another 200–400 years after the last Missoula flood (Atwater, 1987, p. 185–187) before possibly cataclysmically breaching its ice dam, sending a flood down the Columbia Valley downstream of the Okanogan River confluence.

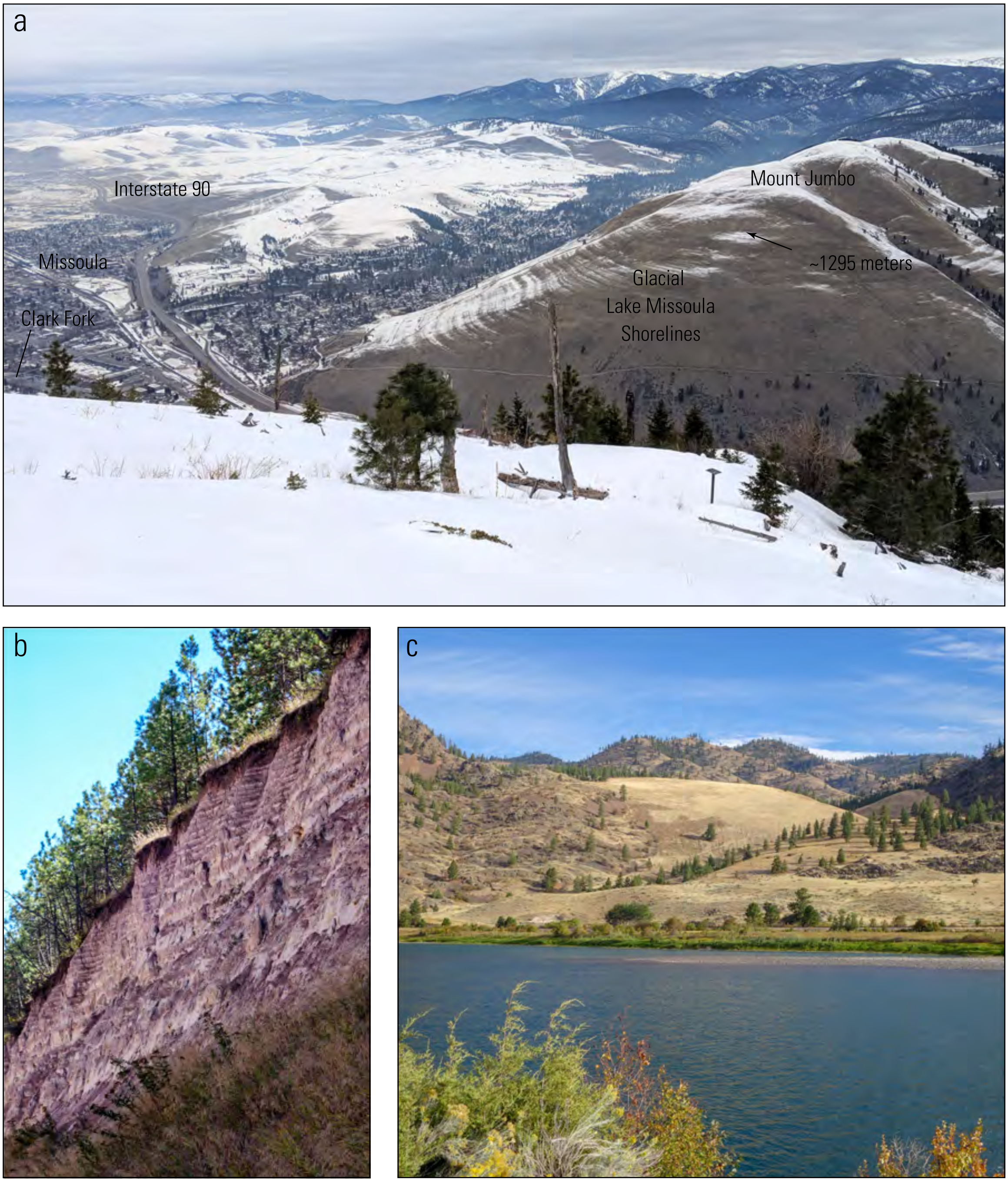

Fig. 7. Glacial Lake Missoula features. (a) View northwest toward Missoula and Clark Fork River Valley from near the summit of 1570 m Mount Sentinel, showing glacial Lake Missoula shorelines etched as high as 1295 m on flanks of Mount Jumbo. Photograph by Brennan D. O'Connor. (b) Ninemile stratigraphic section exposed along eastbound lanes of I-90, where multiple varved sequences separated by sand and silt beds signal dozens of drainings of glacial Lake Missoula (Alt and Chambers, 1970; Chambers, 1971, 1984; Waitt, 1980, 1985a; Smith, 2006; Hanson et al., 2012). The top of the section is at 959 m. Photograph by Jim E. O'Connor. (c) Grass-covered Stout's bar, one of many high eddy bars described by Pardee (1942, p. 1589–1593) filling tributary valleys and gulches flanking the Clark Fork and Flathead River valleys, which became major flow conduits during lake drainage. The top of the deposit is at 1000 m, 230 m above modern river level; flow was from right to left. Photograph by Jim E. O'Connor.

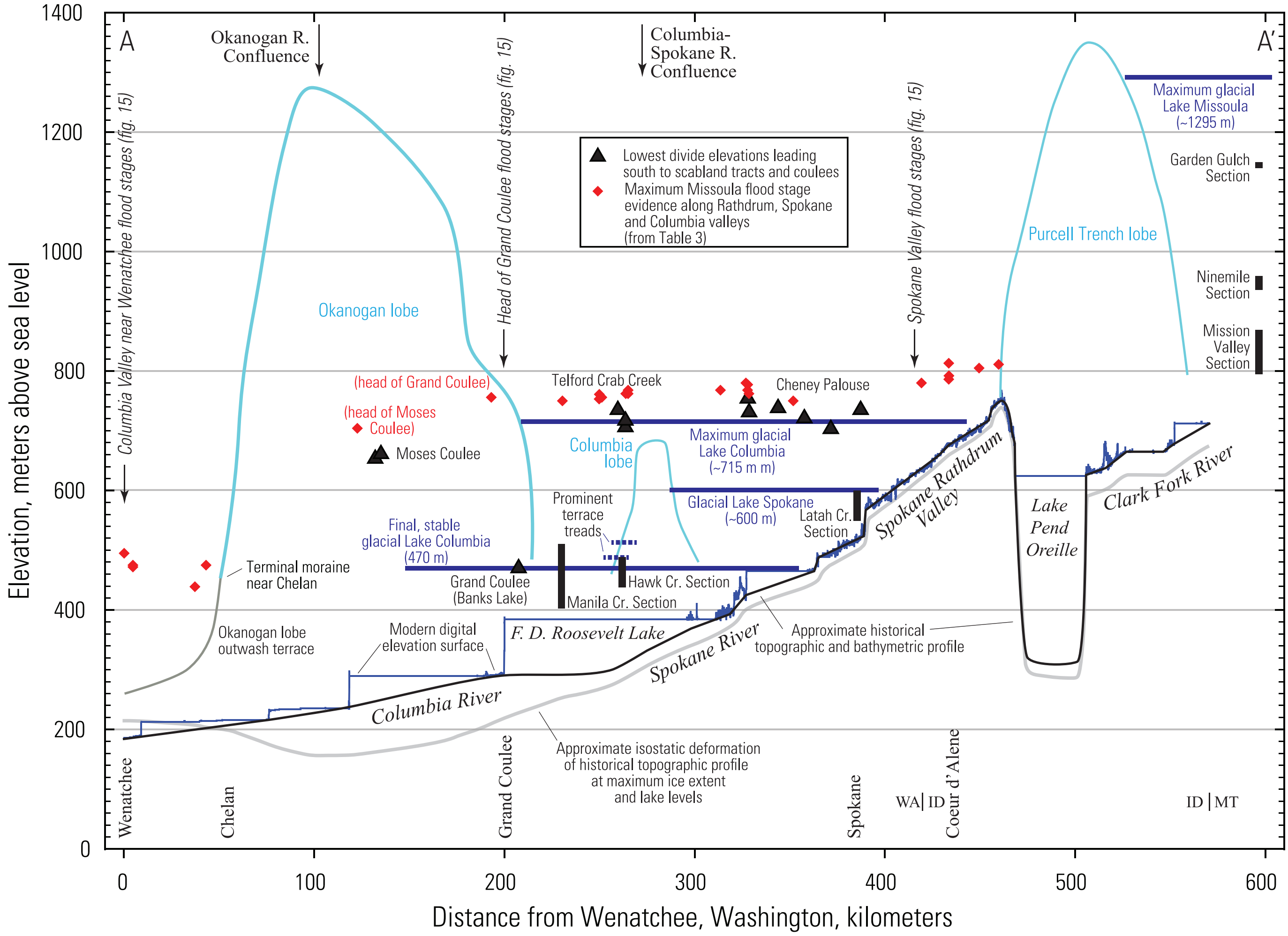

Fig. 8. Profile of key topographic, glacial, and flood features along 600 km of the Missoula flood route between Wenatchee, Washington, and the lower Clark Fork River valley, Idaho and far western Montana. Endpoints shown as A and A' on Fig. 5. Modern digital elevation profile, including modern reservoir surfaces, extracted from U.S. Geological Survey National Map 3D Elevation Program. Approximate historical river and valley profile derived from bathymetric and historical map information. Glacier and terrace profiles estimated from multiple sources listed in fig. 1, supplemented by unpublished observations by O'Connor, Waitt, and Cannon. Estimated equilibrated isostatic response to maximum ice and lake loading (Fig. 6d) is shown as deflection of the historical river and valley profile. Divide elevations are for selected low areas south of Spokane and Columbia valleys leading to Channeled Scabland tracts and coulees cut into southward sloping Columbia Plateau (as depicted in Fig. 6a). Maximum flood-stage evidence from data in Table 3. Key lake levels and stratigraphic section locations and extents from sources cited in text. Glacial Missoula stratigraphic sections are east of the profile. The Mission Valley site is the Landslide Bend section of Levish (1997).

margin of Lake Pend Oreille indicates the timing of final retreat of the Purcell Trench lobe from the Lake Pend Oreille basin after the last cataclysmic release.

Varve counts from lake deposits total 3240–3610 for a low-elevation exposure in Mission Valley, Montana indicating the duration of the lake (Figs. 1 and 8; Levish, 1997, p. 97–98). From optically stimulated luminescence (OSL) dating, Levish estimates an age range for the lake of 19.2–16.0 ka. OSL dating of transgressive sand beneath lacustrine silt and clay at the Garden Gulch section indicates the lake reached 1185 m by 18.7 ± 1.7 to 23.1 ± 1.6 ka, averaging (n = 4) 21.1 ± 1.5 ka (Smith et al., 2018, p. 31)[4]. Hanson et al. (2012) report two OSL ages from the Rail Line section (near Missoula) of glacial Lake Missoula sediments of 12.6 ± 0.6 ka and 14.8 ± 0.7 ka, the younger age contradicted by post-lake deposition of the 13.7–13.4 ka Glacier Peak tephra (Kuehn et al., 2009) across the lake basin (Konizeski et al., 1968; Levish, 1997, p. 131; Smith, 2004) showing the lake gone by then (Levish, 1997, p. 131). Hoffman and Hendrix (2010, p. 71) infer that glacial Lake Missoula had ended prior to 14.6–14.3 ka from a radiocarbon dated sediment sequence deposited in a proglacial lake dammed by a terminal moraine of the Flathead ice lobe after glacial Lake Missoula had left the basin. These age determinations and inferences have some inconsistencies but in general show that glacial Lake Missoula existed for at least 3–4 ky during 20–14 ka.

3.1.2. Columbia Lobe and glacial Lake Spokane

The Columbia River lobe of the Cordilleran Ice Sheet descended the Columbia River valley, blocking the Spokane River near its confluence, creating glacial Lake Spokane (Figs. 5 and 8). The lake attained a stage of at least 671 m and carved benches between 550 and 610 m (Kiver and Stradling, 1995, p. 138, 139; Kiver et al., 1991, p. 238–240), as much as 360 m above river grade at the Spokane-Columbia confluence. Not clear, however, is its timing and role in last-glacial Missoula flooding. Such a blockage and lake, if extant during releases of glacial Lake Missoula, would accumulate flood deposits and divert Missoula floodwater coming down the Spokane valley southwest into the Cheney-Palouse flood tract.

This ambiguity affects interpretation of 15–16 flood deposits interbedded with lacustrine sediment in the Latah Creek valley (also known as Hangman Creek) near Spokane (e.g. Waitt, 1984, p. 52–53; Atwater,

[4] Individual OSL ages are reported as age ± standard error.

1986, p. 37; Kiver et al., 1991, p. 240; Kiver and Stradling, 1995, p. 129; Gaylord et al., 2016, p. 20–21). This exposure, the top of which is at about 595 m (Kiver et al., 1991, p. 241), is 70–90 m higher than similar sequences of deposits formed in glacial Lake Columbia downstream (Fig. 8). This leads us to infer that a higher glacial Lake Spokane was indeed a Missoula flood receptacle for the varve-separated flood deposits, as also inferred by Flint (1936, p. 1858) and Kiver et al. (1991, p. 240). Although it is possible that they were deposited in a high-level glacial Lake Columbia (Waitt and Thorson, 1983) or a smaller lake formed by local impoundment of the Latah Creek valley mouth by aggradation or flood bar (Flint, 1936, p. 1867; Atwater, 1986, p. 37).

Contributing to the ambiguity is poor resolution of the extent and timing of the last-glacial Columbia ice lobe. Waitt et al. (2016, p. 239), from mapping by Kiver and Stradling (Kiver et al., 1989, p.326; Kiver and Stradling, 1989, p. 30–31, 1995, p. 12, 112–118), show the last-glacial ice limit north of the Columbia-Spokane confluence and not blocking the Spokane River. In contrast, our recent field investigations indicate last-glacial ice did indeed extend beyond the confluence (Fig. 1). Last-glacial blockage also may have been by the Enterprise valley sublobe entering the Spokane valley 10 km upstream of the Columbia confluence (Kiver and Stradling, 1995, p. 108, 120). Direct age information on ice position is limited to 100 km north near the Canadian border, where a pre-advance radiocarbon age reported by Clague et al. (1980, p. 323) requires that maximum advance post-dated 21.8–20 ka.

Atwater (1986, p. 36, plate 2) also infers last-glacial blockage of the Spokane valley by the Columbia lobe blockage, denoted by four anomalously thin Missoula flood beds in glacial Lake Columbia deposits downstream. From varve counts tied to a radiocarbon date, he proposes that Columbia lobe blockage lasted "one or two centuries around 15,340 ± 400 [radiocarbon] yr B.P.," which calibrates to approximately 19.6–17.7 ka. Atwater's interpretation, based on varve counts, of only a couple of centuries duration for glacial Lake Spokane possibly conflicts with the duration implied by the 15–16 flood beds and interbedded varves exposed at Latah Creek. The varve counts between Latah Creek flood beds are not fully reported, but range between 10 and 60, with two intervals containing as many as 125 couplets (Kiver et al., 1991, p. 240), although Waitt (1984, p. 53) posits 55 at most. Very approximately, these counts might indicate a last-glacial Lake Spokane duration of ~500 yr.

An alternate interpretation offered for the timing of glacial Lake Spokane is that the lacustrine part of the Latah Creek section was from an older middle Wisconsin lake and flood episode (e.g. Kiver and Stradling, 1985; Kiver et al., 1991, p. 241). Two radiocarbon dates of detrital organic material within floodbeds low within the sequence give ages of >40,000 and 32,450 ± 830 ^{14}C yr BP, possibly consistent with these deposits being older than last-glacial (Gaylord et al., 2016, p. 21). Such ages are largely discounted as too old elsewhere along the flood route because of reworking of the dated materials (e.g. Fryxell, 1962; Benito and O'Connor, 2003). Waitt (1985b) further asserts a middle Wisconsin age is inconsistent with the dated glacial chronology.

Other constraints are offered by geomorphic and stratigraphic relations. Because prominent terraces related to stable levels of downstream glacial Lake Columbia extend far up the Spokane and Columbia valleys (e.g. Jones et al., 1961, p 13, 77–78; Atwater, 1986, p. 36; Kiver and Stradling, 1995, p. 138), the retreat of the Columbia lobe from the Spokane confluence and the demise of glacial Lake Spokane must have preceded the end of glacial Lake Columbia. Hanson and Clague (2016, p. 67) also show stratigraphic evidence that the Columbia lobe both advanced and retreated into an extant glacial Lake Columbia within the Columbia valley north of the Spokane River confluence, requiring that glacial Lake Spokane was both preceded and outlived by glacial Lake Columbia. Coarse flood gravel spilled upstream into the Columbia valley near the Spokane River confluence is inferred by Kiver and Stradling (1995, p. 119) to precede advance of the Columbia lobe as well as glacial Lake Columbia, indicating at least one large flood from glacial Lake Missoula (because it came down the Spokane valley) prior to either the Okanogan or Columbia ice lobes blocking the Columbia.

3.1.3. Okanogan Lobe and glacial Lake Columbia

The Okanogan lobe descended the Okanogan (Okanagan in Canada) River valley, blocking the Columbia River valley near its confluence (Figs. 5 and 6). A sublobe diverted southeast, following the Omak Lake trench (Flint, 1935, p. 175–176), perhaps separately obstructing the Columbia River valley (Hanson, 1970, p. 129). At either location, advancing ice would quickly dam the narrow Columbia valley to depths greater than 760 m, thus creating an ice dam standing at least 530 m above natural river grade. After first filling the east-west trending Columbia valley, the blocking lobes then spilled over to the opposing uplands, coalesced and flowed 50 km south across the Waterville Plateau. At its maximum extent, the Okanogan ice lobe filled 150 km of Columbia valley between Chelan on the west and Grand Coulee on the east, locally with more than 1000 m of ice (Figs. 6d and 8; Kovanen and Slaymaker, 2004). The prominent Withrow moraine marks the broad terminus on the Waterville Plateau (Flint, 1935, p. 177–178; Hanson, 1970, p. 83–88, Kovanen and Slaymaker, 2004). Till of the eastern ice margin rims the west edge of Grand Coulee, and till on Steamboat Rock (Bretz, 1932, p. 35–37), a basalt-floored island stranded by the incision of upper Grand Coulee, shows that the ice sheet once covered at least some of the area of upper Grand Coulee. The western margin of the Okanogan ice lobe sent a short tongue down the Columbia valley to near Chelan, where moraines terminate in prominent outwash terraces 230 m above the natural river grade (Waitt, 1987). This western sublobe was also joined by ice from the more westerly Skagit lobe of the Cordilleran Ice Sheet spilling over passes in the North Cascades and flowing down the Lake Chelan trough (Waitt, 1987).

Blockage of the Columbia valley raised glacial Lake Columbia (Figs. 5, 6 and 8; Table 1). This lake received and diverted ~100 Missoula floods coming from the east, leaving a stratigraphic record of interbedded flood sand and lacustrine varves (e.g. Atwater, 1984, 1986, 1987; Steele, 1991; Kiver and Stradling, 1995; Hanson, 2013; Hanson et al., 2015; Hanson and Clague, 2016). The composite Manila Creek section in the Sanpoil valley records ~89 flood beds (Atwater, 1984, 1986, 1987)—the most complete stratigraphic record of Missoula flooding yet described.

At its end and possibly for much of its existence, glacial Lake Columbia was at a stable stage of ~470 m, spilling through rock-bottomed Grand Coulee (Atwater, 1986, 1987). During rapid influx of Missoula flood waters, Lake Columbia briefly swelled to as high as 750 m near Grand Coulee, marked by ice-rafted erratics at that elevation along the Columbia and Spokane valleys and in the Sanpoil River valley east of Grand Coulee (Fig. 8; Table 3; Atwater, 1986, p. 5–7). Between floods, high short-lived levels of Lake Columbia are traced by patchy strandlines, beach deposits, and possibly fine-grained valley fill—as high as 715–730 m (Pardee, 1918, p. 15; Cochran and Warlow, 1980, p. 40 (725 m); Atwater, 1986, p. 6 (715 m); Kahle and Bartolino, 2007, p. 13–15 (729 m). This highest lake stage is referred to as "Lake Columbia I" by Waitt and Thorson (1983, p. 57, 67) and Kiver and Stradling (1995, p. 121, 129) and "high level Lake Columbia" by Atwater (1986, p. 5). Lower and slightly more prominent strandlines are preserved between 610 and 700 m east of Grand Coulee (Kiver and Stradling, 1995, p. 15, 61, 138), near and within the Sanpoil River valley (Flint, 1935, p. 189; 1936, p. 1857; Cochran and Warlow, 1980, p. 43), and in the Columbia valley upstream of the Spokane River confluence (650–670 m; Flint, 1936, p. 1857). Long-stable lake levels are shown by strongly formed benches between 475 and 535 m that are nearly continuous along the Columbia valley upstream of the Okanogan River confluence. Chiefly composed of bedded silt and sand but commonly capped by fluvial gravel, these benches in part correspond to the "Nespelem silt" of Pardee (1918, plate 1, p. 15, 28–29). They hold the key stratigraphic sections of Manila Creek (Atwater, 1984, 1986) and Hawk Creek (Hanson, 2013; Hanson and Clague, 2016). Kiver and Stradling (1995, p. 17, 67, 124, 140) distinguish two levels among these prominent surfaces, a "Lake Columbia II" stage at 510–535 m that stands distinctly higher—our observations indicate about 30 m near the Spokane-Columbia confluence—than the "Lake Columbia III" terraces at 475–490 m (Fig. 8), which they infer grade to the modern 470-m Grand Coulee threshold.

What controlled the various stages of glacial Lake Columbia and why did it not repeatedly empty like glacial Lake Missoula? Unlike glacial Lake Missoula, deeply confined within Rocky Mountain valleys sloping down toward the ice dam, glacial Lake Columbia was bordered on the south by the uplifted and wrinkled but generally south-sloping basalt of the Columbia Plateau. Along this southern lake border, low spots afforded overflow routes to the south at lake stages well below those that could destabilize the ice dam, especially when ice was thick over the Columbia valley and advanced onto the Waterville Plateau (Figs. 6a and 8, Atwater, 1986, p. 4). The distribution and elevations of these low spots, in conjunction with the advancing Okanogan lobe, dictated lake level. Key elevations from west to east are: (1) 653 m for the divide near Mansfield leading to southward draining Moses Coulee; (2) the 470-m threshold bottom of upper Grand Coulee, the lowest by far and the outlet controlling lake elevation at the end of the last glacial (Fig. 6f; Flint and Irwin, 1939; Waitt and Thorson, 1983; Atwater, 1986, 1987); (3) the 706 m threshold of the Hawk Creek divide separating the Columbia valley and the Telford-Crab-Creek scabland tract; and (4) the 703 m threshold dividing the Spokane River valley from the Cheney-Palouse scabland tract. These outlet elevations are determined from 3 to 6 m contour-interval topographic maps and have uncertainties of similar order. Moreover, these elevations have been almost certainly affected by isostatic adjustments, as discussed in the next section. Despite these uncertainties, the evidence for lake levels ~250 m higher than the 470-m modern Grand Coulee outlet requires that Grand Coulee had not yet fully incised or that it was temporarily filled with ice, as discussed in the "Grand Coulee" section. Similarly, lake levels above the 653 m level of the entrance to Moses Coulee also require ice-blockage or later erosion of the head of that coulee.

Timing of the Okanogan lobe's interaction with the Columbia River is known by the glacial chronology as well as the stratigraphy and chronology of glacial Lake Columbia deposits. A radiocarbon age from deposits preceding ice advance shows that the Okanogan ice margin was at least 250 km north of the Columbia confluence at 23.6–22.5 ka (Clague et al., 1980, p. 323–325). The youngest of five ages reported by Hibbert (1985, p. 98) from carbon-rich sediment and peat below Okanogan lobe till exposed in the Columbia valley calibrate to 36–24 ka. Four ^{10}Be TCN ages of moraine boulders on the Withrow moraine show that the Okanogan lobe pulled back from its maximum extent on the Waterville Plateau after 15.4 ± 1.4 ka (site mean ± standard error), perhaps before the down-Columbia sublobe (which was bolstered by the Skagit ice lobe coming down the Chelan trough) pulled back from its maximum extent near Chelan by 14.5 ± 1.2 ka (Balbas et al., 2017). Within uncertainties, these ages mostly concur with more variable ^{36}Cl TCN ages of 16.3 ± 1.0 ka, 13.2 ± 0.9 ka, and 12.2 ± 1.1 ka obtained by M. Zreda and V.R. Baker on erratics inferred left by Okanogan lobe near its southeastern limit on top of Steamboat Rock in Grand Coulee (Keszthelyi et al., 2009, p. 86). The Balbas et al. (2017) results are significantly younger than previously reported ^{36}Cl TCN ages of 29.4 ± 0.8–22.3 ± 0.8 ka for the Withrow moraine (recalculated by them from results reported by Swanson and Caffee (2001)). The Okanogan lobe had retreated 15–20 km north from the Columbia River confluence by the time 13.7–13.4 ka Glacier Peak tephra fall (Kuehn et al., 2009), judging from its preservation on the floor of the lower Okanogan valley (Porter, 1978, p. 39).

Age estimates from glacial Lake Columbia, impounded by the ice lobe, derive from radiocarbon dates, the presence of the Set S tephra, and varve counts. Two radiocarbon dates from detrital plant material within varved lacustrine sediment give ages of 14,490 ± 290 ^{14}C yr BP (Atwater, 1986, p. 29; Manila Creek section) and 13,400 ± 100 ^{14}C yr BP (Hanson and Clague, 2016, p. 74; Hawk Creek section), showing the lake persisted until at least the 16.4–15.8 ka calibrated age of the younger date. Additional lake duration of several centuries beyond this younger age is implied by 37 flood beds inferred above the 16.4–15.8 ka sample at Hawk Creek (Hanson and Clague, 2016, p. 74). From varve counts at the Manila Creek section in the Sanpoil valley, Atwater (1986, p. 11) estimates a total lake duration of 2,000–3,000 yr. If we accept the 18.35–16.83 ka Manila Creek calibrated radiocarbon age and account for its position with respect Atwater's (1986) varve counts and their uncertainties, the Manila Creek section shows lake onset began by 19.1–17.5 ka and the last Manila Creek varve was laid down sometime within 17.1–14.6 ka. As Atwater (1986, p. 29) notes, these ages for lake onset and demise may be too old if the detrital wood was older than the enclosing deposits. The timing of the last varve at Manila Creek includes an estimated 200–400 varves above the uppermost flood bed (Atwater, 1987, p. 185–187). From this Atwater infers that glacial Lake Columbia outlasted blockage of the Clark Fork by the Purcell trench lobe and impoundment of glacial Lake Missoula by 2–4 centuries, suspecting, however, that this is an underestimate because of poor exposure (Atwater, 2020, written communication).

Hanson's (2013, p. 90, 116, 206) discovery of the Mount St. Helens Set S tephra in the Manila Creek section adds another tie point for age estimates which gives similar results. Atwater (1986, fig. 28, Plate 3) counted 70–100 varves between the position of the Set S found by Hanson, 15–17 flood beds from the top, and the highest flood bed at the Manila Creek section, indicating about a century of Missoula floods in this section after the ~16 ka tephra fall. Varve counts below the tephra total 1655–2525 (Atwater, 1986, fig. 28, Plate 3), indicating glacial Lake Columbia formed by 18.5–17.7 ka, as measured from the ~16 ka age of the Set S, which also has poorly resolved uncertainty of several hundred years (Clynne et al., 2008).

An assessment of the Manila Creek section independent and discordant from the varve counts is Hanson's (2013) analysis of the paleomagnetic secular variation within the uppermost ~46 flood beds and intervening lacustrine deposits. Correlating the measured secular variation to an established reference curve, Hanson (2013, p. 114) estimates that the deposits formed over a period of 2600 years between 14.2 and 11.6 ^{14}C ka BP, equating to 17.3–13.4 ka. Extrapolating based on Atwater's, 1986 description of ~40 additional stratigraphically lower flood beds and the 55 yr recurrence interval estimated from the secular variation within the upper part of the section, Hanson (p. 115) suggests that the base of the lacustrine deposit is ~2200 years older, about 19.5 ka, a result about 1 ky older than the radiocarbon based chronologies. Moreover, in contrast with the 2600-yr age span and the average 55-yr interflood interval inferred from secular variation, the varves interbedded with the upper 46 flood beds at Manila Creek likely total 600–1000 and no individual interflood interval has more than 27–37 varves (Atwater, 1986, his figs. 15–17 and plate 3).

Taken together, the chronologic information indicates the Okanogan lobe blocked the Columbia River for 2–5 ky after 19.5 ka and before 13.5 ka, and during the ~16 ka Mount St. Helens tephra fall. The radiocarbon- and tephra-based chronologies from the Manila Creek section give a tighter approximate range; ~18.5 ka to ~15.5 ka with about 1 ky uncertainty for both beginning and ending.

3.1.4. Isostatic deformation

For all three glacial lakes, their elevations, volumes, and outlets were affected by isostatic depression and rebound as the masses of ice and water came and went. Warping also likely affected flood routes and gradients. The complicated submergence and emergence patterns and their hydrologic consequences in the Puget Sound area demonstrate the importance of these effects (Thorson, 1989). In conjunction with the Okanogan ice lobe, isostatic adjustments are most evident in the elevation of lacustrine silt deposited in glacial Lake Columbia where its outlet arm extended south through upper Grand Coulee to the ~470 m lip of Dry Falls. Here the silt terrace rises north, toward the main masses of ice and water, from 480 m to 500 m elevation over 40 km (Flint, 1935, p. 189; Atwater, 1987, p. 188). This northward tilt of 0.5 m/km approximates the 0.6 m/km of net northward tilt in Puget Sound as it rebounded from the mass of Puget lobe (Dethier et al., 1995, p. 1299).

We have the estimated regional isostatic response to ice and lake loading by finite element modeling of the maximum ice and lake extents shown in Figs. 6d and 8. These calculations assume a fully equilibrated flexural response governed by the thickness and rigidity of the crustal structure estimated by Buehler and Shearer (2010) from seismic interferometry.

Results are shown in Fig. 8 in relation to the historical river and valley profile along the flood route between Wenatchee and the Clark Fork River valley. Importantly, the actual magnitude of isostatic elevation change for any specific feature depends on when the feature formed in relation to loading and the dynamic crustal response, which is mostly unknown. The portrayed deflection indicates maximum plausible elevation adjustments for last glacial features along the river and valley profile. For example, where the Okanogan lobe was thick and broad over the Columbia valley near the Okanogan River confluence, we estimate nearly ~90 m of equilibrated depression at maximum loading. The magnitude diminishes to the south and east in a manner approximating the observed northward rise of lacustrine silt in upper Grand Coulee. The results also indicate ~40 m of differential post-glacial uplift along the trace of the southern divide and outlets of glacial Lake Columbia. This variation brings the last-glacial maximum elevations of the less-uplifted divides entering the Cheney-Palouse and Telford-Crab-Creek scabland tracts (now at 703 and 706 m) to 20–30 m above the 715-m-high shoreline evidence in the Sanpoil valley described by Atwater (1986, p. 6–7), which our calculations show depressed to about 680 m elevation at the time of maximum ice and lake extent.

Isostatic loading to the west as the Okanogan lobe advanced may have facilitated progressively greater flood flows into Moses Coulee and Grand Coulee as the entrances to those tracts were weighted down (but prior to being blocked) by the advancing ice and resulting lake. Differential uplift also complicates correlation of terraces related to glacial lakes Columbia and Spokane (Kiver and Stradling, 1995, p. 141–142). Broadscale tilting also affects flow-path gradients, adding uncertainty to the flow modeling results based on modern topography. Isostatic deformation beneath glacial Lake Missoula may also require adjustment of lake volume estimates based on modern topography, much like it does for pluvial Lake Bonneville (e.g. Abril-Hernández et al., 2018, p. 2–3). For many reasons, better understanding of the rates and patterns of regional isostatic deformation in consequence of the various ice lobes and resulting glacial lakes would help overall understanding of the lakes and floods.

3.2. Missoula flood routes

All floods from outrushing glacial Lake Missoula exited the lake basin through the Clark Fork valley in western Montana and entered the eastern Pacific Ocean by way of the Columbia River through the Columbia River Gorge (Fig. 1). But between leaving the lake and entering Wallula Gap, the plexus of flood routes depended on conditions at the breaching ice dam, the position of the Okanogan ice lobe, the status of glacial lakes Columbia and Spokane, progressive erosion of the scabland tracts, and release size. After entering the Pacific Ocean, flood waters may have taken different submarine routes depending on ocean conditions.

3.2.1. Clark Fork and Flathead River valleys

At maximum ice and lake extent, the 150-km-long Clark Fork valley between Lake Pend Oreille and the confluence of the Flathead River was occupied by ice and the deep and narrow downstream arm of glacial Lake Missoula, here confined by the Cabinet and Bitterroot Mountains of western Montana and northern Idaho. The narrowest section is "Eddy Narrows" (Fig. 1; Pardee, 1942, p. 1574), where at its maximum 1295 m elevation (Table 3), glacial Lake Missoula spanned only 2.25 km across. When the ice dam failed, however, the Clark Fork valley, as well as the lower Flathead River valley, became raging flood conduits for water funneling out of the voluminous basins upstream in the upper Clark Fork and Flathead River drainages, constrained by bottlenecks like Eddy Narrows (Pardee, 1942, p. 1594–1597). The outrushing floodwater left immense dune-covered boulder gravel bars in the valley bottom and high eddy bars in tributary mouths (Fig. 7c) and intensely eroded the Proterozoic metasedimentary rocks forming the valley walls (Pardee, 1942; Smith, 2006). Inset against and locally capping the coarse flood deposits are silt, clay, and lesser amounts of sand and gravel in sequences up to 25 m thick (Fig. 7b). These beds are inferred to be lacustrine sediment deposited in reformed glacial Lake Missoula (Pardee, 1942; Smith, 2006), and have been studied as records of the number and timing of lake drainages (Alt and Chambers, 1970; Chambers, 1971, 1984; Curry et al., 1977; Waitt, 1980, 1985a; Fritz and Smith, 1993; Levish, 1997; Shaw et al., 1999, 2000; Atwater et al., 2000; Hanson et al., 2012; Hanson, 2013; Smith and Hanson, 2014).

3.2.2. Spokane-Rathdrum valleys

Floods exited the breached ice dam near the south shore of Lake Pend Oreille. Some water diverted northwest into the Pend Oreille River valley and then circled back south to the Spokane River valley northeast of Spokane by way of the Little Spokane River valley (Fig. 5; Kiver and Stradling, 1982; Waitt, 1984; O'Connor and Baker, 1992; Waitt et al., 2016). Most water, however, coursed southwest through Rathdrum and Spokane-valleys (Bretz et al., 1956; Baker, 1973). This 75-km long trench narrows from about 12 km wide at its head to 5 km wide just east of Spokane. In most places the valley bottom is filled with 200–250 m of Pleistocene glaciofluvial and flood deposits documented by well logs and seismic data (Bolke and Washington, 1980; Gerstel and Palmer, 1994; Kahle and Bartolino, 2007, p. 17–18, plate 2).

Floods speeding through this route eroded midvalley knobs and bedrock valley slopes, deposited massive foreset-bedded gravel bars, some mantled with giant current dunes, and left valley-blocking eddy bars in tributary mouths (Flint, 1936; Richmond et al., 1965; Weis and Richmond, 1965; Weis, 1968; Baker, 1973; O'Connor and Baker, 1992; Breckenridge and Othberg, 1998a, 1998b; Lewis et al., 2002; Waitt et al., 2016). High eddy bars, divide crossings, scabland, and ice-rafted erratics define maximum-flood stages declining from 810–820 m near the southern tip of Lake Pend Oreille to ~780 m near Spokane and downstream (Table 3; Fig. 8; Baker, 1973, p.17; O'Connor and Baker, 1992, p. 274). These stages are as much as 170 m above the present valley floor. Side-valley lakes in tributary valleys now blocked by eddy bars, like 64-m-deep Coeur d'Alene Lake (Woods and Berenbrock, 1994), suggest a pre-flood main-valley bottom 60–70 m lower than present and flood depths as great as ~250 m.

Floods entering the Spokane-Rathdrum valley at times encountered glacial Lake Columbia or possibly glacial Lake Spokane whenever lake stages exceeded ~600 m elevation. At its maximum stage of 715–730 m, glacial Lake Columbia would have reached the terminal moraine now impounding the southern arm of Lake Pend Oreille (Figs. 6d and 8). And the 750 m flood swollen stage of glacial Lake Columbia (Atwater, 1986, p. 5–6) would impede and elevate Missoula flood stages through the valley (O'Connor and Baker, 1992, p. 275). At glacial Lake Columbia's minimum stable level, draining over the present Grand Coulee threshold at 470 m, the valley would be lake-free (Fig. 6f). It is also likely that glacial Lake Spokane at its maximum level ~670 m (Kiver and Stradling, 1995, p. 137–139) backflooded the valley near Spokane and would be engorged by Missoula floods down the Rathdrum valley. One or both of the glacial lakes seem recorded in the Rathdrum valley by fine-grained layers in the valley fill, ranging between 504–729 m in Rathdrum Prairie and a prominent layer, 49–81 m thick and topping at elevations of 505–524 m in the western Spokane valley north of the main floodway (Kahle and Bartolino, 2007, p. 13–15).

This reach has had a long history of Quaternary investigations. Some of what are now interpreted as Missoula flood deposits were originally thought glacial, leading to inferred last-glacial ice margins south and west of Spokane (Bretz, 1928b; Flint, 1936, 1937), much beyond current mapping of the last-glacial Cordilleran ice sheet in northeast Washington and Idaho (Waitt et al., 2016). Balbas et al. (2017, Table DR2) report a ^{10}Be exposure ages ranging from 14.3 ± 1.2 to 17.2 ± 1.4 for three boulders deposited on one of the broad expansion bars at the head of the Rathdrum valley. The southwest edge of Lake Pend Oreille is bordered by till, dated by the 14 ± 1.2 ^{10}Be exposure age on a protruding boulder (Balbas et al., 2017, Table DR2, FAR04), showing by its position that the Purcell Trench lobe blocked the Clark Fork valley after the last substantial lake outflow (Breckenridge et al., 1989, p. 19; Lewis et al., 2002; Smyers and Breckenridge, 2003, p. 13). Final outwash and minor flooding (if any) flowed northwest through the small Hoodoo Channel to the Pend Oreille River valley.

The Latah Creek section near Spokane has been the major stratigraphic focus in this reach (Rigby, 1982; Kiver and Stradling, 1982, 1985; Waitt, 1984, 1985b; Kiver et al., 1991; Meyer, 1999; Gaylord et al., 2016, p. 18–26), although Waitt (1984) describes additional sections in the area. At Latah Creek, 15–16 floods left turbidite-like beds in glacial Lake Spokane or possibly a high level of glacial Lake Columbia, overlain by another 12 flood beds subsequent to lake withdrawal and local incision (Waitt, 1984; Gaylord et al., 2016, p. 18–26). The lacustrine deposits in the valley reach ~600 m elevation and may correlate with the fine-grained layers in the valley fill described by Kahle and Bartolino, 2007, p. 13–15). As noted above, the age of this section is uncertain. Waitt (1984, 1985b) asserts that it is entirely last glacial, yet Kiver et al. (1991, p. 241) suggest the lower, lacustrine, part may be middle Wisconsin, in part based on several radiocarbon dates from nearby flood sediment giving ages of 30–40 ka (p. 238).

3.2.3. Cheney-Palouse tract

Floods exiting the Spokane and Rathdrum valleys entered the deep but narrow Spokane River valley on its 120-km-way west to meeting the Columbia River valley coming in from the north. The immense flood volumes overwhelmed the tortuous drainage, at times already partly filled by glacial lakes Spokane and Columbia. Spilling west and south, the floods overtopped the northern edge of the basalt plateau and its cover of old alluvium and loess. The diverted water carved the Cheney-Palouse scabland tract, a plexus of basalt-floored channels extending 150 km southwest from the Spokane area, mainly toward the Snake River and Pasco basin. It is the longest and easternmost of the scabland tracts (Fig. 5). Ragged, sharp-edged channels starkly contrast with the rolling loess hills of the Palouse, commonly left as streamlined 'loess islands' between floodways (Fig. 9; Washington Department of Natural Resources, Division of Geology and Earth Resources, 2016). This contrast prompted Bretz's (1923a, p. 573) first paper advocating flows of "great vigor over large tracts."

Maximum flood levels reached 770–780 m spilling out of the Spokane River valley into the Cheney-Palouse tract (Table 3; Fig. 8). The now-eroded cols of those divides are 703 m for the large eastern tract chiefly following Rock Creek and the Palouse River, ultimately leading to the Pasco basin, and 721 m for the smaller western tract which shunted some flow into the Crab Creek drainage and Quincy basin (Figs. 5 and 6a). These divides stand 240–255 m above the present Spokane River valley bottom. Flow in the eastern scabland tract sloped steeply southwest, guided by the south dipping surface of the Columbia River Basalt Group. At the divide between the former Palouse River course of Washtucna Coulee and the Snake River canyon, maximum stages were about 460 m, descending to about 410 m at the confluence of the Snake and Palouse Rivers (Bretz, 1928a, p. 646). Floods poured into the Snake River Canyon, all water eventually heading west to Pasco basin but some at first backflooding 250 km *up* the Snake River leaving pebbly silt and ice-rafted erratics as high as 395 m near Lewiston and slackwater deposits far up the Snake River in Hells Canyon (Bretz, 1929a, p. 405–427; Schmidt et al., 2009; Kurt Othberg, Idaho Geological Survey, written commun., 2017).

Bretz's (1923a) first report on the Cheney-Palouse tract identified (and named) the eroded "scabland" (p. 577–579). He described channels excavated into ~60 m of loess (p. 588) and then as deep as 60 m into the basalt, anastomosing between "isolated linear groups of Palouse Hills, their marginal slopes steepened notably" (p. 588). Within the channels are "great bars and terraces" (p. 584), chiefly in the lee of basalt outcrops. Bretz returned (1928a, p. 646–664), mainly describing the bars but also the flood-caused diversion of the Palouse River from its former Washtucna Coulee path west southward to the Snake River (Fig. 5). The Cheney-Palouse tract also attracted flood doubters; especially Flint (1938) who proposed that the bars were instead terraces recording regional aggradation. Allison (1941) soon after noted the implausibility of several aspects of Flint's fill hypothesis, instead suggesting many of the features in the Cheney-Palouse were caused by ice jams downstream that impounded a temporary lake. Patton and Baker (1978a) later mapped the Cheney-

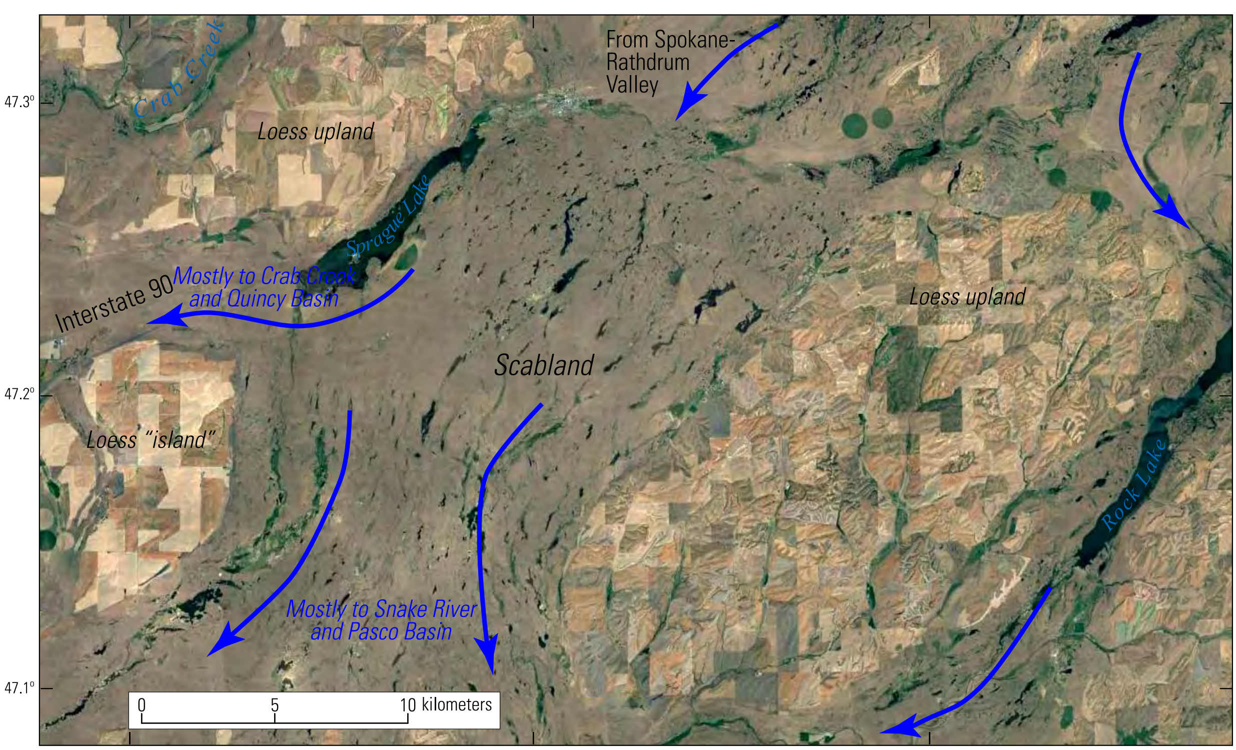

Fig. 9. GoogleEarth image of a part of the Cheney-Palouse tract showing plexus of scabland eroded into the basalt of the Columbia River Basalt Group separated by streamlined "islands" of uneroded loess uplands. Area of figure shown in Fig. 5.

Palouse tract, describing the morphology of scabland flood forms and relating them to erosional and depositional processes.

The scabland and giant floods bars of the main Cheney-Palouse scabland tract are spectacular, but the backflooded valleys entering from the east best preserve stratigraphic records. These valleys were systematically described by Bretz in 1929a where he documents "pebbly silt" (e.g. p. 402), commonly containing erratic clasts, inferred ice-rafted, as well as current structures locally showing *up valley* transport. The maximum elevations of these backflood deposits are similar within individual drainages but descends among the drainages, north to south, following the overall scabland gradient (Table 3). These deposits are most studied along the Snake valley upstream of the Palouse River confluence, where energetic water from the Cheney-Palouse tract left bedded silt and sand up the Snake River and its tributary valleys. Exposures in the Tucannon valley, which joins the Snake valley 5 km upstream of the Palouse confluence (Fig. 5), record about 25 floods (Smith, 1993, p. 95) that passed over the Palouse-Snake divide and up the Snake valley, including seven that post-dated the ~16 ka Mount St. Helens tephra (p. 98). The number accords with the 21 fine-grained Missoula flood beds 130 km farther up the Snake valley overlying bouldery gravel left by the 18.3–18 ka Bonneville flood coming down the Snake River (Fig. 4e; Waitt, 1985a, p. 1277). Foley (1976, p. 33–34) reports that slackwater deposits up the Snake River near the Idaho border overlie alluvium containing charcoal dated 16.2–15 ka.

The Cheney-Palouse tract also contains evidence of earlier flood episodes (Patton and Baker, 1978b; Baker et al., 1991; Bjornstad et al., 2001). Most compelling are gravel bars capped by thick petrocalcic horizons appended to the downstream tails of loess islands, such as the Marengo locality (Baker et al., 2016, p. 51) where foreset-bedded flood gravel is overlain by paleomagnetically reversed loess and a petrocalcic horizon giving a Th/U radiometric age of 800 ka (Bjornstad et al., 2001). The locations of these deposits show that at least some of the channels of the Cheney-Palouse tract had incised earlier.

3.2.4. Telford-Crab-Creek tract

Some floodwater entering Crab Creek and Quincy basin came by way of the Telford-Crab-Creek scabland tract, 60 km east of the Cheney-Palouse (Fig. 5). This plexus of shallow basalt-floored channels formed by flood-flow overtopping in several places along a 25-km-wide swath south of the Spokane-Columbia River confluence. The lowest divide saddle is at 706 m, ~380 m above the Columbia River valley bottom (Figs. 6a and 8). Most flow followed the preflood southwestward drainages of Crab Creek, Coal Creek, Duck Creek, Lake Creek, Marlin Hollow, Canniwai Creek, and Wilson Creek; carving ever-deepening coulees into the Columbia River Basalt Group before spreading into the broad 2000-km^2 Quincy basin. Maximum flood stages descended steeply from 760–770 m where they first spilled into the tract from the Columbia Valley, to 410–420 m in Quincy basin (Table 3).

Despite the pronounced scabland tract left by flood water exiting the Columbia valley, preliminary two-dimensional flow modeling (discussed below) shows that this tract only conveys substantial water if boosted by glacial Lake Columbia standing at a high level, requiring blockage of the Columbia River by the Okanogan lobe and that upper Grand Coulee is also blocked (Baker et al., 2016, p. 15–17; Denlinger et al., in press). We know of no geochronology or stratigraphic studies for this little-studied tract. And no evidence of older flooding has yet been reported.

3.2.5. Grand Coulee

Grand Coulee is an 80-km-long trench extending southwestward from the Columbia River valley to Quincy basin (Figs. 5 and 10). It is 1.5–8 km wide and cuts as much as 340 m into adjacent basalt uplands and underlying granitic rocks. The coulee and its genesis were beautifully described by Bretz (1932), Bretz et al. (1956, p. 967–974) and Bretz (1969, p. 522–524). As inferred by 19th-century observers, Grand Coulee held a diverted Columbia River when blocked from its big bend route around the northern margin of the Columbia River Basalt Group by ice of the Okanogan lobe

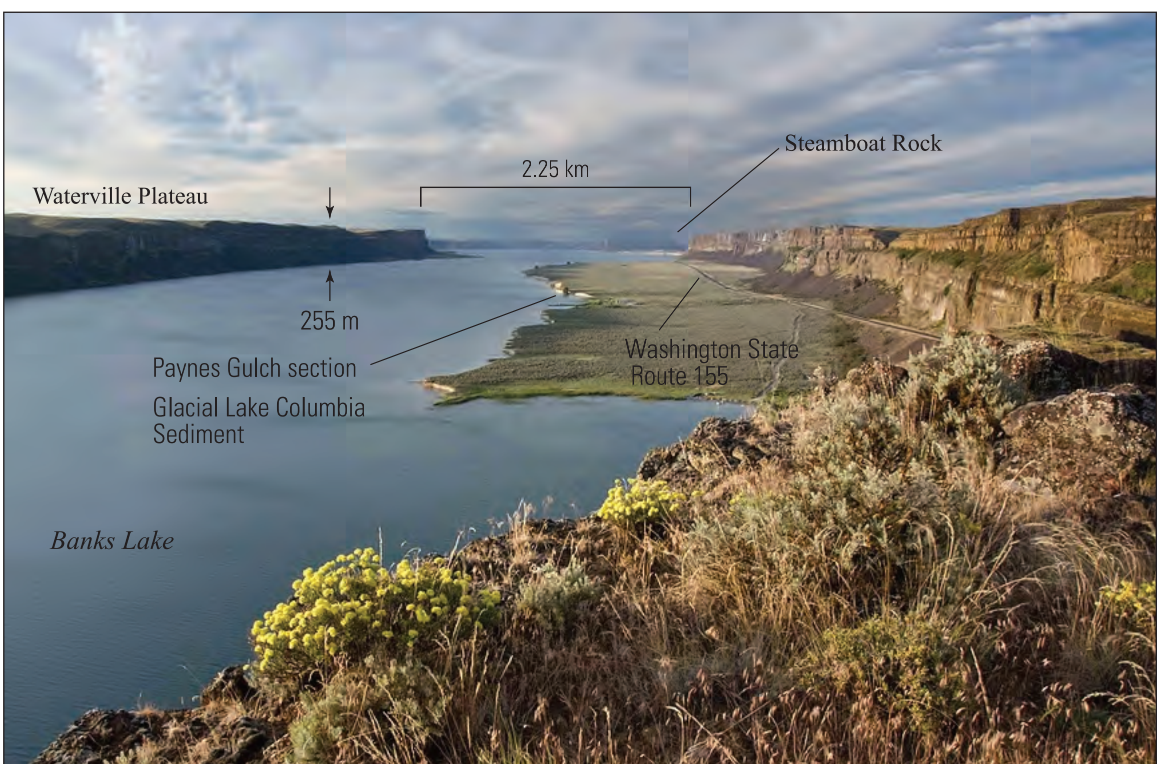

Fig. 10. View north (upstream) of upper Grand Coulee. Steamboat Rock mid-coulee in the obscure distance. Banks Lake, hemmed in by low dams built at both ends of the coulee now floods the irregular coulee bottom to ~479 m, ~10 m above the historical 470–m threshold at the south end of the lake. Glacial Lake Columbia sediments flanking Banks Lake include the Paynes Gulch section described by Atwater (1987, p. 192–195), as high as 486 m, and probably graded (but isostatically uplifted) to the 470–m outlet elevation (Atwater, 1987, p. 188). Photograph by Bruce Bjornstad.

(Fig. 6d; Parker, 1844, p. 307; Symons, 1882, p. 109, 119–120; Russell, 1893; Dawson, 1898). Bretz (1932) clarified the role of huge floods in carving the coulee (though Symons, 1882, p. 120, postulated formation by "a flood of water or ice"). Advancing Meinzer's (1918) assessment, Bretz (1932) attributed erosion of the coulee to headward erosion by giant cataract complexes; the lower Grand Coulee (p. 61) by a set of cataracts headcutting 30 km north along the southwest trending Coulee monocline, stalled and preserved as the spectacular 120-m-high Dry Falls complex. Above Dry Falls, upper Grand Coulee (p. 47–61) formed by another cataract system retreating nearly 40 km north from where the Coulee monocline bends northeast. The upper cataract completed its erosional work all the way to the Columbia River valley, consuming itself as it opened a 2.7-km-wide, 250-m-deep entrance to upper Grand Coulee floored 175 m above the Columbia.

Grand Coulee and its flanking uplands are marked by flood and glacial features, and its bottom hosts flood bars coated by lacustrine deposits left late in glacial Lake Columbia's occupancy (Fig. 10). Most spectacular are the scabland coulees and cataracts. The lower coulee, from the Dry Falls cataract complex to Quincy basin, is a wild maze of dry canyons, vertical plunges, and deep holes carved into the basalt (Bretz, 1932, p. 8–32). The upper coulee is chiefly a vertical walled-trench, 2 km across and 250 m deep, but it widens to 8 km across where it is joined by sets of cataract-headed (and beheaded) canyons from flood flow on the flanking uplands spilling into upper Grand Coulee. This widening is possibly where retreat of the upper cataract slowed or stalled upon the rising granitic rocks outcropping under the Columbia River Basalt Group. Here also the mid-coulee prominence of Steamboat Rock preserves an island of original flood- and ice-scoured basalt upland. In the coulee bottom bouldery flood bars lie in protected side-canyon mouths, in the lee of Steamboat Rock and other knobs and mesas (Bretz, 1928a, p. 670), mounded in wider coulee segments, and in the broad pre-flood depression of Hartline basin north of Dry Falls between the upper and lower coulee segments (Bretz, 1932, p. 23–26, p. 44–45, 52–53, 68, 72–73).

Upper Grand Coulee closely traces the eastern limit of the maximum last-glacial extent of the Okanogan lobe on the Waterville Plateau (Figs. 5 and 7d; Bretz, 1932, p. 34–38, plate 1; Waters, 1933, p. 785–787; Flint, 1935; Hanson, 1970, p. 128–129; Waitt and Thorson, 1983). Steamboat Rock, the remnant island of upland in the upper Coulee, is partly mantled with glacial till (Bretz, 1932, p. 35–37), as is the upland west of the upper coulee. Bretz (1932, p. 36–38) and Flint (1935, p. 179) describe features possibly glacial on the coulee floor, leading Bretz to infer that the upper cataract retreat was complete and the coulee was fully entrenched before final occupation by the Okanogan lobe. This interpretation is supported by Bruce Bjornstad photographs (2019 written communication) of glacial striae etched into granite along the shore of Banks Lake in the coulee floor 2 km northeast of Steamboat Rock.

Geochronology for last-glacial Grand Coulee flooding is chiefly by TCN dating of granitic surfaces exposed by erosion of the overlying basalt. Keszthelyi et al. (2009, p. 861) report a ^{36}Cl exposure age of 17.2 ± 2.7 ka for an exhumed inselberg near the head of Grand Coulee. Balbas et al. (2017; Table DR2) report exposure ages of 14.9 ± 1.2, 15.8 ± 1.3 and 16.0 ± 1.3 from ^{10}Be analyses of exhumed granitic rocks within the Northrup Canyon cataract complex, indicating the timing of the most recent floods across the upland scabland east of the coulee and erosion of the capping basalt within the abandoned cataracts. Two OSL ages, both greater than 36 ka and seemingly erroneously old, are reported by Keszthelyi et al. (2009, p. 861) for lacustrine deposits near Steamboat Rock. No pre-last-glacial flood deposits have been found in Grand Coulee.

The present Grand Coulee outlet threshold at the south end of Banks Lake near the brink of Dry Falls held glacial Lake Columbia at ~470 m during the late phase of Missoula flooding (Table 1; Figs. 6f and 8). Varved lacustrine silt and clay covering the floor of upper Grand Coulee, the outlet arm of the lake, and in the adjacent Columbia valley up to elevations of 504 m (Fig. 10; Bretz, 1932, p. 76, 78; Atwater, 1987) suggest a persistent lake graded to the outlet, but now tilting up to the north because of differential isostatic rebound. Moreover, rippled sand beds within these lacustrine deposits in the lee of Steamboat Rock indicates at least 14 current-emplaced sediment incursions—inferred as late and small Missoula floods—into this low-level glacial Lake Columbia, some separated by just three or four varves (Atwater, 1987, p. 188; Waitt, 1994).

Was this 470-m outlet elevation in Grand Coulee the controlling level for glacial Lake Columbia for the entire last-glacial period? Or was there significant lowering during the period of Missoula flooding? And when did the retreating cataract complex of upper Grand Coulee break through to Columbia valley and lower the coulee entrance? These related questions arise because stratigraphy and geomorphology indicate a fully deepened coulee early in the last-glacial period. Yet flow modeling results suggest that upper Grand Coulee was not incised prior to last-glacial flooding of Moses Coulee and the Columbia valley to the west around its big bend. The timing and magnitude of incision of upper Grand Coulee has been a persistent question, addressed by Bretz (1932, 1935, p. 35–38), Richmond et al. (1965, p. 237–239), Hanson (1970, p. 64, 69), Waitt and Thorson, 1983, p. 57, 64), Atwater (1986, p. 36, 57), and Balbas et al. (2017) among others. It has implications relating to the various levels of glacial Lake Columbia, flood magnitudes in Grand Coulee and other scabland tracts, the temporal evolution of scabland tract erosion, and possibly routing of pre-last glacial floods.

Our renewed interest in upper Grand Coulee incision owes to recent flow modeling results (Denlinger et al., in press, discussed in more detail below) suggesting that a still-blocked upper Grand Coulee is required to funnel sufficient flow into the two flood tracts farther west—Moses Coulee, and the Columbia valley following the big bend. Both routes, as described below, passed large last-glacial floods, probably early in the overall Missoula flood sequence (Waitt, 2016). If, however, upper Grand Coulee is open to its present depth and width, too much floodwater diverts south, leaving insufficient flow to produce the observed flood evidence in the western pathways. This conclusion implies that the cataract retreat creating upper Grand Coulee had not yet breached the Columbia valley or otherwise significantly lowered or widened the coulee entrance. This was similarly inferred by Hanson (1970, p. 64, 69) in concluding that large Moses Coulee floods preceded incision of upper Grand Coulee. Additional possible support of last-glacial cataract retreat in upper Grand Coulee are the cosmogenic exposure dates of flood-exhumed granite in upper Grand Coulee and downstream granitic flood boulders in Quincy Basin. Both indicate last-glacial erosion of the basaltic and granitic rocks at the head Grand Coulee at ~17–15 ka (Baker et al., 2016, p. 36–37; Balbas et al., 2017, p. 585).

Geomorphic and stratigraphic evidence, by contrast, suggests that upper Grand Coulee was open to near its present 470-m elevation near the beginning of the last glacial period, at least for the duration of glacial Lake Columbia (Atwater, 1986, p. 37). The lacustrine silt and clay on the floor of upper Grand Coulee (Fig. 10) and prominent terraces at 475–500 m extending up both the Columbia and Spokane River valleys show a long-lived last-glacial lake graded to near the present 470 m-outlet elevation in Grand Coulee while the Columbia River was still blocked by the Okanogan lobe (Fig. 6f; Bretz, 1932, p. 76, 78; Flint, 1936, p. 1868, plate 6; Jones et al., 1961, p 13, 77–78; Atwater, 1987; Kiver and Stradling, 1995, p. 128, 139–140). The evidence for short-lived glacial Lake Columbia levels as much as 250 m higher than the present 470-m Grand Coulee threshold is explained by the Okanogan lobe filling and blocking an already-incised Grand Coulee, signaled low in the Manila Creek section of glacial Lake Columbia deposits by four anomalously thin flood beds interpreted to indicate attenuated deposition in a deepened lake (Atwater, 1986, p. 22–23, 30–31, 34–35; 1987, p. 187). This scenario is consistent with evidence that the Okanogan lobe at its maximum extent did fill an already-deepened coulee.

One scenario possibly resolving the opposing interpretations is a hybrid one similar to that proposed by Waitt and Thorson (1983, p. 57): The earliest last-glacial floods did indeed erode the upper cataract complex back to the Columbia valley, but after or during the first few large floods that first passed through the Columbia valley and Moses Coulee to the west. Prior to the cataract complex reaching the Columbia valley, glacial Lake Columbia, once formed by the advancing Okanogan lobe, may have briefly stabilized at the ~653 m head of Moses Coulee until it was overrun by ice, as shown in Fig. 6c. Once that outlet was covered and prior to full cataract retreat, any glacial Lake Columbia outlet would have been further east, perhaps a saddle where Grand Coulee is now, or one the divides just at 700–710 m leading to the Cheney-Palouse or Telford-Crab-Creek scabland tracts. The absence of strongly formed shorelines at high elevations indicates that any such stability was short-lived (and perhaps also affected by outlet erosion and the growing ice load). The lake and flood dynamics changed markedly, however, when the upper Grand Coulee cataracts reached the Columbia valley, lowering and widening the opening to Grand Coulee. This later condition was likely the predominant circumstance—scores of Missoula floods entering a low-level glacial Lake Columbia and building stratigraphic records like that of the Manila Creek section over 2–3 ky, temporarily perturbed by the Okanogan lobe reaching its maximum extent and filling the fully open Grand Coulee and deepening the lake.

Additionally, Kiver and Stradling (1995, p. 15–16, 126–130, 139–140), propose at least some last-glacial incision of upper Grand Coulee to explain the ~30 drop from an apparently stable "Lake Columbia II" stage at 510–535 m to the "Lake Columbia III" terraces at 475–490 m, which they correlate to the final 470 m Grand Coulee threshold (Kiver and Stradling, 1995, p. 17, 67, 124, 140), thus suggesting "significant changes in base level" within upper Grand Coulee (p. 127). Such lowering, in part premised on distinct terrace levels we have observed near Hawk Creek as shown in Fig. 8, is represented by the different glacial Lake Columbia levels of Fig. 6e and f. Correlation of these terrace levels, however, is complicated by incomplete mapping, uncertain isostatic deformation, and absent geochronology. Moreover, it is uncertain if this minor amount of lowering could change flood flows sufficiently to produce the observed flood levels in Moses Coulee and Columbia valley.

Considering the conflicting flow modeling results and stratigraphic interpretations, the timing, magnitude, and effects of upper Grand Coulee incision remain as unresolved questions for understanding last-glacial Missoula flooding. Nevertheless it is one possibly addressed with lidar-aided geologic and geomorphic mapping and stratigraphic and geochronologic studies focused on the history glacial Lake Columbia and Grand Coulee.

3.2.6. Quincy Basin

All floods passing through Grand Coulee entered Quincy Basin, which also received floodwater from the east from the Telford-Crab-Creek tract and western strands of the Cheney-Palouse tract (Fig. 5). Floods from these varied source areas covered Quincy Basin with bouldery fans and thick gravel and sand deposits (Bretz, 1923a, 1928a, p. 670–673; Bretz et al., 1956, p. 969–974). The immense foreset-bedded deltaic gravel bar complex of Ephrata fan, formed chiefly of basalt excavated from Grand Coulee, is 500 km^2 broad and up to 40 m thick (Bretz et al., 1956, p. 969–974; Baker, 1973, p. 9, 15, 39–42). This deposit includes immense boulders, including an $18.5 \times 11 \times 8$ m fragment of basalt entablature (Fig. 11; Baker, 1973, p. 23–29, 40–41).

Maximum floods exited Quincy Basin by four separate cataract-floored scabland channels—Crater Coulee, The Potholes, and Frenchman Coulee all went west into the Columbia valley; and Drumheller Channels south and then west to the Columbia by way of Lower Crab Creek or south into Pasco Basin through Othello Channels (Fig. 5). The entrances to these four Quincy Basin overflow points all show very similar maximum flood stages of about 410 m (Table 3). Bretz (1928a, p. 670–673) made the case, restated in Bretz et al. (1956, p. 985), that this similar flood stage is implausible by sequential outlet erosion, but instead suggests simultaneous overtopping of Quincy Basin through all four exit spillways. This scenario requires rapid filling of Quincy Basin, compelling support for a huge flood (Baker, 2008, p. 39). Final smaller floods, or at least waning flow, as well as the diverted Columbia River through Grand Coulee between floods, incised through flood gravels on their way through Drumheller Channels into Lower Crab Creek (Bretz et al., 1956, p. 971–974), rejoining the Columbia River valley upstream of Sentinel Gap.

Balbas et al. (2017; Table DR2) report an average age of 15.6 ± 1.3 ka (mean ± standard error) from seven granitic boulders in the Ephrata fan, tightly clustered between 15.0 ± 1.2 and 16.2 ± 1.3 ka. These ages are similar to the exposure ages of unroofed granitic outcroppings in upper Grand Coulee. Flood gravel from pre-last-glacial flooding are also exposed in western Quincy Basin, where two old floods left carbonate-cemented

Fig. 11. Immense fragment of basalt entablature eroded from Grand Coulee and transported to the Ephrata fan where lower Grand Coulee opens into Quincy Basin. Photograph by Jim E. O'Connor.

east-dipping foreset gravels derived from floods spilling east into the basin from the Columbia valley, opposite the direction of most if not all last-glacial floods (Bretz et al., 1956, p. 985; Bretz, 1969, p. 524; Baker, 1973, p. 8–9).

3.2.7. Moses Coulee

To the west, Moses Coulee is a slightly smaller version of Grand Coulee. It leads south from the glaciated Waterville Plateau, entering the Columbia River valley south of Wenatchee (Fig. 5). Bretz (1923a, p. 600–602) described its overall form and setting, later elaborating some of the large flood bars and coulee features (Bretz, 1928a, p. 673–675; 1930, p. 386–396). It was mapped thoroughly and described in the Ph.D. thesis by Larry G. Hanson (1970).

Moses Coulee encompasses three distinct geomorphic settings over its 70-km path—two coulee segments separated by a broad and shallow scabland tract (Bretz, 1923a p. 600). The upper coulee emanates from the poorly defined Mansfield channels (Hanson, 1970, p. 52–53) into a distinct canyon near the Withrow moraine terminus of the Okanogan lobe. Here the coulee is 1–2 km wide and 150 m deep. The canyon walls diminish and lose definition over 15 km southward as it enters a structural trough. Here a plexus of scabland channels and cataracts—the Rattlesnake Springs (or "Three Devils") complex (Fig. 12)—gathers into a deep gash cut across the Badger Mountain anticline, 800 m deep and 2 km across, and the coulee floor drops 140 m in 10 km. Similar to Grand Coulee, Bretz (1923a, p. 600) inferred these cataracts formed by headward retreat from knickpoints initiated near the Badger Mountain anticline. Downstream of this cataract complex, Moses Coulee is a straight flat-bottomed trench for 21 km before joining the Columbia River valley 30 km downvalley of Wenatchee and 80 km below the maximum extent of ice in the Columbia valley near Chelan.

Like Grand Coulee, Moses Coulee shows both flood and glacial features. Scabland basalt emerges from the obscuring glacial margin at the head of the coulee, as well as downstream where flow escaped the coulee confines within the structural trough leading into the Rattlesnake Springs cataract complex at the head of the lower coulee segment. These erosional features indicate the maximum flood stage descended from about 704 m near the Withrow moraine to 580 m where entering the cataract complex (Table 3). Flood bars near the coulee mouth top off at 280 m but maximum stages were undoubtedly much higher. The floods left large convex flood bars studded with boulders in the coulee bottom (Bretz, 1928a, p. 673–675), systematically mapped by Hanson (1970, p. 43–47). One covers 3.8 km^2 and stands 75 m above the valley floor. At Moses Coulee's mouth, a massive foreset-bedded boulder bar protrudes into the Columbia River valley (Bretz, 1930, p. 390–393; Bretz et al., 1956, p. 989–990; Hanson, 1970, p. 48–51; Waitt, 2016, p. 434).

The Withrow moraine defines the maximum Okanogan ice lobe extent on the uplands east and west of the coulee. A prominent flood bar at the head of Moses Coulee is distinctly overlain by the terminal moraine where it drops into the coulee, showing that the flood bars and the coulee itself preceded the Okanogan lobe reaching its maximum extent. Braided outwash gravel surfaces descend down-valley from the terminal moraine, inset against and partly burying flood bars and locally filling the upper coulee segment with >60 m of gravel (Hanson, 1970, p. 40). The lower coulee has fill up to 100 m thick (Bretz, 1930, p. 395; Hanson, 1970, p. 41); most is likely outwash and concealed flood bars, but the upper several meters in the lowermost ~10 km of coulee are bedded silts deposited by Missoula floods backflooding *up* Moses Coulee from the Columbia valley.

The most complete stratigraphic description pertinent to Moses Coulee flooding is for an up-Columbia remnant of the large Moses Coulee bar formed at the coulee mouth (Waitt, 1985a, p. 1275). Here, four up-Columbia-dipping basalt-gravel flood beds are separated by varved sequences containing as many as 37 silt-clay couplets. The basalt gravel beds must owe to Moses Coulee floods since sediment coming down the Columbia is basalt-poor (Bretz, 1930, p. 392). The varves suggest impoundment of the Columbia River valley, probably by the Moses Coulee flood-bar itself. The sequence of Moses Coulee flood beds and lacustrine sediment is capped by 11 m of fine sand and silt beds similar to Missoula flood slackwater deposits ubiquitous to downstream backflooded areas. The beds are uncounted here, but another exposure we recently investigated at the north end of the Moses Coulee bar shows two basaltic gravel beds separated and overlain by sand-silt varve-like couplets. These couplets are overlain by at least 19 visible sand and silt beds and probably

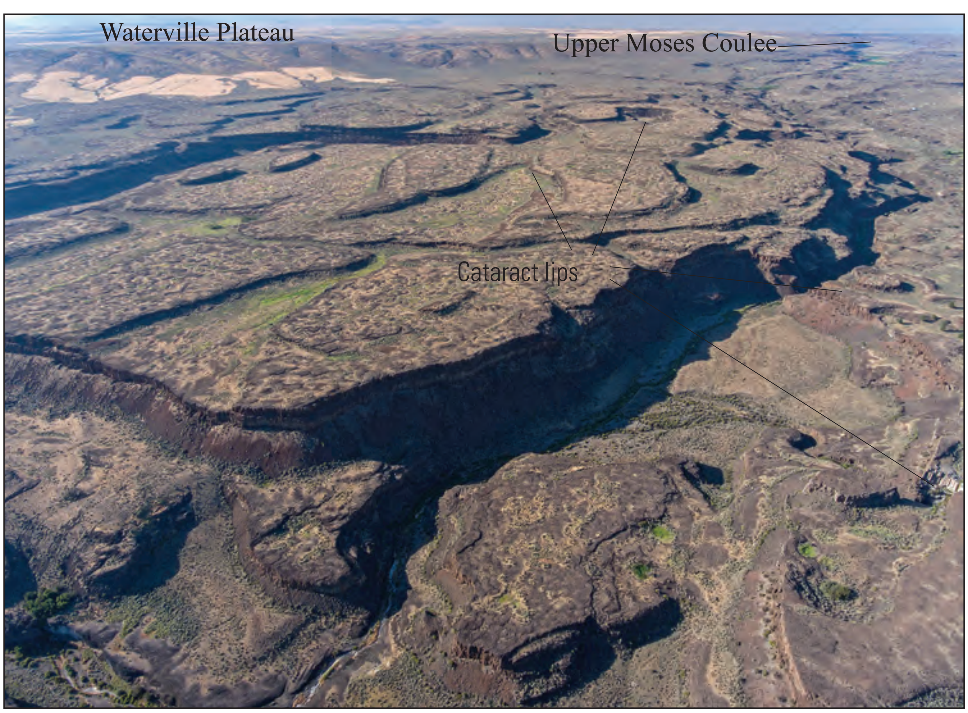

Fig. 12. Oblique aerial view up Moses Coulee showing a portion of the Three Devils scabland complex. Bretz (1923a, p. 600; 1930, p. 394–395) inferred these channels eroded by upstream cataract recession, as suggested by cataracts at the head of several channels. Photograph by Bruce Bjornstad.

another ~15 under cover. The stratified basalt-gravel beds and varves show the last two Moses Coulee floods into a lake. The ~34 overlying sand and silt beds probably owe to far travelled flood floodwater following the Grand Coulee and Telford-Crab-Creek pathways farther east, entering the Columbia via Crab Creek and Quincy Basin spillovers and backflooding ~60 km up the Columbia valley, a scenario shown clearly by recent flow modeling (Denlinger et al., in press).

This stratigraphy in conjunction with TCN dating aids understanding of the last glacial flood sequence involving Moses Coulee. We infer that Moses Coulee conveyed substantial Missoula-flood discharges only during blockage of Columbia River by the Okanogan lobe, enabling flood stages to overtop the ≥653-m entrance divides, but prior to burial of the entrance paths at the head of the coulee by the advancing ice lobe (Waitt, 2016). The stratigraphy at the mouth shows that at least four floods passed down Moses Coulee during this window. A ^{36}Cl TCN age of 15.5 ± 2.9 ka on a basalt boulder on the surface of the flood bar near the head of the coulee may date the last large Moses Coulee flood (Keszthelyi et al., 2009, p. 861). Four ^{10}Be TCN ages of nearby granitic boulders on the Withrow moraine range from 17.1 ± 1.4 to 13.5 ± 1.1 ka, mean 15.4 ± 1.4 ka (Balbas et al., 2017, Table DR2), consistent with the clear geomorphic evidence that the maximum Okanogan lobe followed the last Moses Coulee flood. The fully advanced Okanogan lobe shunted floods to the more eastern flood paths (Fig. 6c; Hanson, 1970, p. 83; Waitt, 2016), but then circuitously backflooding up the Columbia valley and depositing the ~34 fine-grained floodbeds in the lower coulee and overlying the bar at the coulee mouth.

The longer history of Moses Coulee and its relation to Grand Coulee remain uncertain. As noted above, because the cols of the divides between the Columbia valley and Moses Coulee are all above 653 m, Grand Coulee, its entrance upstream and 180-m lower at 470 m, would have diverted much incoming Missoula floodwater. Consequently, Hanson (1970, p. 64, 79) concluded that Grand Coulee could not have been incised to its present depth when Moses Coulee was conveying substantial flood water, a finding consistent with recent flow modeling discussed below showing little water entering Moses Coulee if upper Grand Coulee is fully open (Denlinger et al., in press).

Also uncertain is the timing of the overall formation of Moses Coulee. All flood bars have last-glacial soils (Hanson, 1970, p. 76) and no older flood deposits are known. The Rattlesnake Springs cataract complex appears fresh and probably was the major source of the immense flood bar at the coulee mouth. Yet flood flows within the upper and lower coulee segments were fully contained, indicating a pre-existing drainageway. And the margins of the lower coulee are deeply notched by large tributary drainages, some closely at grade, giving the appearance of a long-established drainageway (Bretz, 1923a, p. 601; Bretz, 1930, p. 395–396; Hanson, 1970, p. 26–27). It likely that Moses Coulee was carved during previous advances of the Okanogan lobe and blockage of the Columbia valley, but the specifics and timing remain unknown.

3.2.8. Columbia River sans Okanogan Lobe

At least one last-glacial Missoula flood came down the mainstem Columbia River valley around the big bend, emplacing large bars, high eddy deposits, and ice-rafted erratics along the 90-km reach between Chelan and the entrance of Moses Coulee (Fig. 5). Although its distinctly more felsic and sandy deposits were recognized by Bretz et al. (1956, p. 986–987), they were not attributed to Missoula flooding until Quaternary geologic mapping in the Chelan and Wenatchee areas in the 1970s by Richard Waitt (1982, 1987), reported in a 1977a abstract titled "Missoula flood *sans* Okanogan lobe."

High eddy bars and tractive bars can be traced from near Chelan at the down-Columbia maximum extent of the Okanogan ice lobe near Chelan to beyond the mouth of Moses Coulee (Waitt, 2016, 2017; Hendrick et al., 2017). They stand above the prominent outwash terrace descending south from the maximum ice extent (Figs. 8 and 13). Two huge bars at Brays Landing and Pangborn are covered with giant current dunes. Pangborn bar, just downstream of Wenatchee and covering nearly 40 km^2, is one of the largest individual Missoula flood bars. The bars have distinctly sandy and quartzo-feldspathic compositions in contrast to the basaltic gravel deposits within and downstream of the eastern scabland channels (Bretz et al., 1956, p. 986–987). These flood deposits may correlate to bouldery foreset gravels and coarse sands underlying till of the Okanogan lobe exposed along the Columbia valley upstream of the Okanogan River confluence, also indicating a pre-Okanogan lobe flood (Hibbert, 1985, p. 96, 107), as well as the coarse flood deposits at the Spokane and Columbia river confluence inferred by Kiver and Stradling (1995, p. 119) to precede glacial Lake Columbia. Maximum flood stages for this down-Columbia flood were 250–300 m above river level, indicated by ice-rafted erratics as high as 495 m near Wenatchee (Fig. 8; Waitt et al., 2019), descending to stripped basalt, erratics, and an eddy bar as high as 420 m on Babcock Bench, 45 km downstream (Table 3).

This flood (or floods) preceded the maximum advance and long-lasting blockage of the Columbia River valley by the Okanogan lobe. Weakly developed capping soils on flood deposits show it to be last glacial (Waitt, 1982, 1987; Waitt et al., 2019). Three TCN ^{10}Be ages of high ice-rafted erratics gave ages of 18.1 ± 1.6, 18.2 ± 1.6, and 18.5 ± 1.6 (Balbas et al., 2017, Table DR2), results consistent with the flood preceding the Okanogan lobe blockage of the Columbia valley and formation of glacial Lake Columbia. A single analysis of a high erratic downstream of Moses Coulee, likely left by the down-Columbia flood, gave an age of 23.0 ± 1.9 ka, possibly affected by prior exposure.

This reach, like the Cheney-Palouse tract, contains evidence of pre-last-glacial floods. A foreset bedded gravel capped with a 4-m-thick petrocalcic horizon underlies a small pendant bar outboard and above Pangborn bar. At least two floods spilled out of the Columbia valley and east *into* Quincy basin by way of Potholes Coulee (Bretz et al., 1956, p. 985; Bretz, 1969, p. 524; Baker, 1973, p. 8–9). Additionally, as discussed below in the "Other Last-Glacial Floods" section, at least two floods in this reach postdated the Okanogan lobe blockage of the Columbia (Waitt, 2016).

3.2.9. Pasco Basin and Wallula Gap

All Missoula floods passed through broad Pasco basin before funneling through Wallula Gap, the tight valley constriction at its outlet (Fig. 5). The Pasco basin is a structural and topographic trough of about 4000 km^2 in south-central Washington. The Columbia River enters from the northwest through Sentinel Gap, cut through the Saddle Mountain anticline. The river continues southeast into the basin, joined by the Yakima River from the west and by Snake River and Walla Walla Rivers from the east. Thus bolstered, the Columbia River exits the basin southward through Wallula Gap, incised through the Horse Heaven Hills—another narrow anticline. From here the river turns west towards Umatilla basin and the Columbia River Gorge, and the Portland Basin.

Missoula floodwater entered the Pasco basin by way of the Columbia River valley (including flow entering from Quincy basin and Crab Creek), from the Snake River valley (mainly supplied by the Cheney-Palouse scabland tract), from Washtucna Coulee and Equatzel Coulee from the northeast (also mostly from the Cheney-Palouse tract), and the Othello Channels from the north (mainly from Quincy basin). All flow exited through Wallula Gap, 2 km wide at its narrowest and 275 m deep. This narrow gap, and others downstream in the Columbia River Gorge, impeded flood flow out of the Pasco basin. This hydraulic ponding backflooded Pasco Basin tributary valleys like Yakima valley to the west (Bretz, 1930, p. 412–420) and Walla Walla valley to the east (Bretz, 1929b, p. 516–536). This passing wave of ponded flood water is sometimes confusedly referred to as "glacial Lake Lewis" (after Captain Merriweather Lewis), a name proposed by Symons (1882, p. 108) to explain the ubiquitous ice-rafted boulders, gravel deposits, and bedded silt.

High ponded flood stages in Pasco basin are marked by thousands of ice-rafted erratics beached on flanking hillslopes (e.g. Bretz, 1930, p. 409–410), the highest yet found is 366 m (Bjornstad, 2014, p. 54). This elevation accords with the highest divide crossing (347 m) and loess scarps (up to 360 m) at the Wallula Gap outlet, 270 m above the Columbia River's natural level (Table 3). A transient flood stage of 360–366 m would

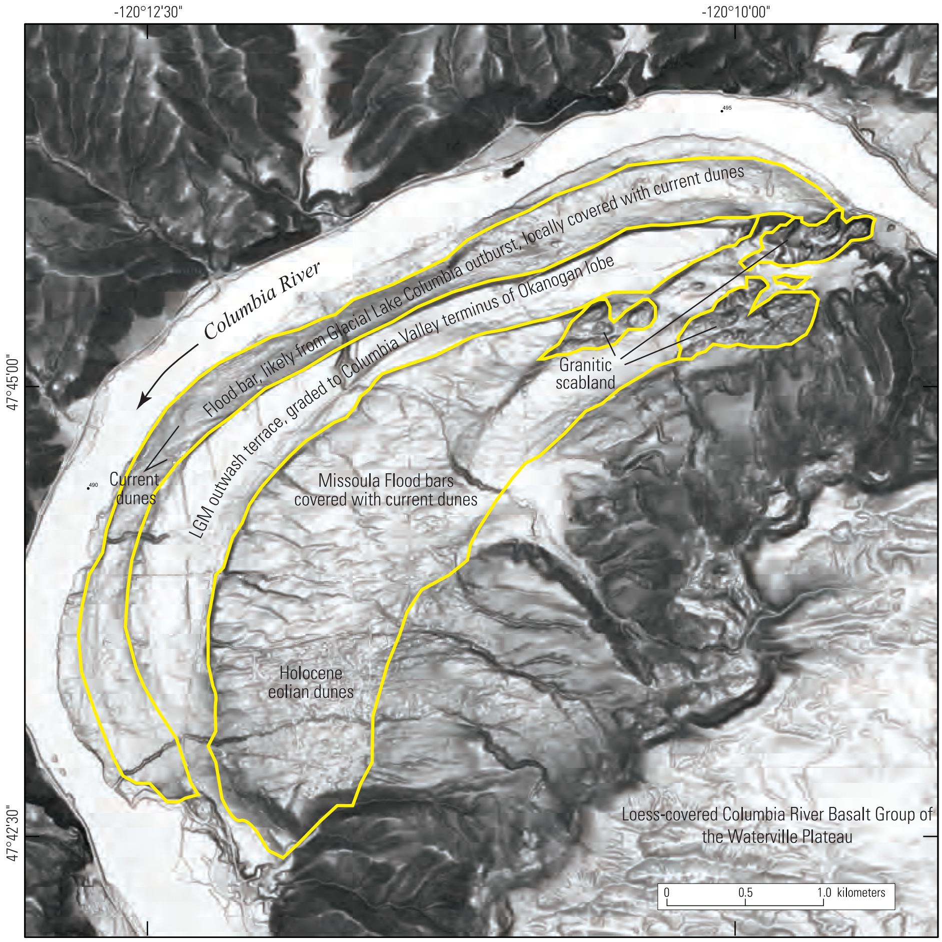

Fig. 13. Last-glacial flood and outwash features in the Brays Landing area mapped onto hillshade digital elevation model derived from 1-m resolution lidar. Immense, bouldery, current-dune covered Missoula flood bar in point-bar position extends six km downstream from eroded granite outcrops and climbs to 260 m above historical river level. This bar was deposited by Missoula flood(s) preceding blockage of the Columbia River by the Okanogan lobe. Inset against the bar complex is the mostly planar, downstream sloping cobble-gravel terrace graded to the maximum extent of a down-Columbia finger of the Okanogan lobe that terminated about 7 km upstream from the right figure edge. Inset against the outwash terrace is another boulder-armored flood bar, its surface mostly disturbed by housing, but still preserves a few obvious giant current dunes. This lower bar probably owes to the outburst of glacial Lake Columbia at about 14.5 ka. Associated flooding may have channeled the outwash terrace above. Figure outline shown on Fig. 5. Lidar obtained from the Washington Lidar Portal (https://lidarportal.dnr.wa.gov).

inundate 13,000–14,000 km^2 of eastern Washington with 1300–1400 km^3 of water (Table 1), backflooding far up the Columbia and Snake valleys. The transiently impounded volume is more than half the total volume of a maximum glacial Lake Missoula.

Floods slackened as they entered the expansive Pasco basin, depositing broad bars of sand and gravel along major flow routes. The largest, Cold Creek bar, covers more than 30 km^2 (Bretz, 1928a, p. 679–681; Bretz et al., 1956, p. 1009–1015; Bjornstad et al., 1991). Away from main flow routes, at higher elevations, and in back flooded tributaries, finely bedded sand and silt drape pre-flood topography (Bretz, 1925a, 1929a, 1929b, 1930). These were named the "Touchet Beds" by Flint (1938, p. 493–495, 503–504).

The Touchet Beds, commonly termed rhythmites because of their repetitive couplets of sand and silt, have long been a focus of Missoula flood controversies (Fig. 14). Bretz (1925a, 1929a, 1929b, 1930) mapped the extent of a "widespread mantle of silt" (Bretz, 1929, p. 393), which he inferred to be slackwater flood deposits composed of reworked loess eroded from scabland channels, though puzzled by the "varvelike arrangement of silt and fine sand" (Bretz, 1929b, p. 533). Flint (1938) and Allison (1941) cited the bedded silts to invoke ponding under alternative explanations for scabland features, inferred formed in the aggrading and then incising fill of a long-lived Lake Lewis dammed by downstream blockages. In Bretz's final (1969) Missoula flood paper, he affirms the rhythmites as loess-derived slackwater deposits laid by water hydraulically ponded behind Wallula Gap, in part based on evidence of deposition by upvalley currents (Bretz et al., 1956, p. 1034), but he wavered on their significance: "There are altogether too many in one section, and they are too thin to assign each couple to a separate flood influx. Thicknesses of the total deposit seem too great, however, to assign to one flood."

Such a "separate flood influx" interpretation was soon put forward. Waitt coming across the spectacular Burlingame Canyon exposure in 1977, concluded from it and similar rhythmite sections in the Yakima and Walla Walla valleys that each layer was indeed from a separate deluge, each "followed by decades of normal subaerial environments" (Fig. 14; Waitt, 1980, p. 655). Thirty-nine layers at Burlingame Canyon suggested about 40 floods, matching counts of similar beds far downstream at the River Bend section in the Willamette Valley (Fig. 1; Glenn, 1965) and 40 rapid drainages inferred at the Ninemile exposure of glacial Lake Missoula bottom sediment (Alt and Chambers, 1970; Chambers, 1971). This evidence came together as the controversial jökulhlaup concept of scores of Missoula floods from a self-dumping glacial Lake Missoula (Waitt, 1980, 1984, 1985a, 1985b).

The question of whether the Touchet Beds represented just a few or several dozen Missoula floods lived on for 40-plus years (e.g. Bretz, 1969; Bjornstad, 1980; Waitt, 1980, 1984, 1985a; Bunker, 1982; Baker and Bunker, 1985; Moody, 1987; Smith, 1993; Shaw et al., 1999, 2000;

Fig. 14. Nearly 40 beds of Missoula flood slackwater deposits exposed in 30-m-deep Burlingame Canyon within the Walla Walla River valley. This section has been analyzed by Bjornstad (1980), Waitt (1980), Moody (1987), and Clague et al. (2003), and was the key site supporting Waitt's (1980) hypothesis that each bed was laid down by individual floods decades apart, thus indicating dozens of Missoula floods. Photograph by Bruce Bjornstad.

Atwater et al., 2000; Clague et al., 2003; Hanson, 2013). Nevertheless, abundant and diverse evidence—tephra falls, paleomagnetic secular variation, mammal and insect recolonization—shows that that most individual rhythmites in Pasco Basin and tributary valleys were indeed emplaced by individual Missoula floods years to decades apart (e.g. Waitt, 1980, 1985a; Spencer, 1989; Smith, 1993; Clague et al., 2003; MacEachern and Roberts, 2013). Still, questions remain, such how do the extensive rhythmite sections correlate with the stratigraphic records in glacial Lake Columbia, Lake Spokane, and Lake Missoula, and even those offshore? And what was the genesis of unconformities in some of the Walla Walla valley sections, particularly evident at the Cummings Bridge section (Moody, 1987, p. 307–308; Bjornstad et al., 1991, p. 706–707), but also noted elsewhere (Spencer and Jaffee, 2002)? Can clues in the slackwater sediment stratigraphy, mineralogy, and geochemistry help unravel the sequence of the various upstream flood routes and their changes with time?

In addition to addressing the number of floods, the extensive deposits in the Pasco Basin attract efforts to determine the timing of last-glacial flooding. Many results are variously tabulated in Bjornstad et al. (1991), p. 230); Hanson (2013, p. 199–203), Baker et al. (2016, p. 11), and Waitt (2016, p. 439). Most dating has been by radiocarbon but OSL and TCN techniques are now being applied (Baker et al., 2016, p. 11; Bjornstad, 2014; Balbas et al., 2017). Key radiocarbon ages (calibrated to calendar years) from within deposits include a 17.9–17.6 ka mammoth bone from within high-elevation rhythmites in the Yakima valley (Lillquist et al., 2005), 18.3–15.9 ka and 16.6–14.6 ka shells within lower Yakima valley rhythmites (Waitt, 1985a; Baker and Bunker, 1985). High-elevation rhythmites in the Walla Walla valley are overlain by a soil containing molluscs dated at 15–14.3 ka, calibrated from ages reported in Spencer and Knapp (2010).

Clague et al. (2003) examined the stratigraphy and paleomagnetic secular variation of three flood-rhythmite sequences in the Yakima and Walla Walla valleys, including the Burlingame section (Fig. 14) previously analyzed by Waitt (1980) and Bjornstad (1980). From the magnetic variation and presence of Mount St Helens Set S (So-Sg) tephra couplet, Clague et al. (2003) developed a correlated stratigraphy among the three sites encompassing 31 flood beds below the tephra couplet and 22 beds above. They concluded from the secular variation that the individual flood beds were laid down decades apart and the entire sequence deposited over a few thousand years. Moreover, by comparing the variation to a reference curve derived from dated lacustrine sections in Oregon and California, they propose two possible correlations: (1) For the 10 flood beds below the tephra and the 13 above (for which they had measurements at multiple sites), the statistically best match indicates an interval of 60 yr between couplets and a tephra age of 13,350 ^{14}C yr BP, equivalent to 16.2–15.9 ka. (2) Considering the entire section and possibly varying intervals between floods, the best correlation predicts a 14,400 ^{14}C yr BP age estimate for the tephra, equivalent to 17.7–17.4 ka, and a ~20.5–15.5 ka range for all 53 beds. Both scenarios are plausible but the first better fits the ~16 ka age for the Set S tephra couplet suggested by Clynne et al. (2008, p. 619).

TCN ages from the Pasco Basin are broadly consistent with the radiocarbon and magnetic variation results. Keszthelyi et al. (2009, p. 861) report four ^{36}Cl TCN ages for ice-rafted erratics on slopes surrounding Pasco Basin. Three for erratics mapped by Bjornstad (2014) on Rattlesnake Mountain give ages ranging from 16.9 ± 3.4 to 16.2 ± 1.3 ka. The fourth age from a large erratic at Badger Coulee is 35.6 ± 2.1 ka, possibly affected by prior exposure. Two erratics on the high western rim of Wallula Gap have ^{10}Be TCN exposure ages of 18.2 ± 0.8 and 66.0 ± 5.8 ka, the older also discounted because of likely inheritance (Balbas et al., 2017, Table DR2), the younger has an age consistent with it being floated on a down-Columbia flood before the Okanogan lobe blocked the Columbia valley.

OSL age ranges of Pasco Basin flood deposits range wider. Keszthelyi et al. (2009, p. 856–857) report ages between 21 ± 2 and 12 ± 2 ka for a sequence of low-elevation flood beds near the center of Pasco Basin. Hanson (2013, p. 99) gives ages of 14.1 ± 0.8 ka and 5.7 ± 1.6 ka just above and below the Set S tephra, respectively, from a sequence of flood beds in Yakima valley.

Flow exited Pasco Basin through Wallula Gap, a narrow constriction flowing 270-m deep during maximum floods. It has been a focal point for discharge estimates (Table 2), starting with a calculation by D.F. Higgins in Bretz (1925b, p. 257–258). The maximum flow stage entering the constriction was about 360 m (O'Connor and Baker, 1992), descending to 340 m near The Dalles (Table 3). Also important is evidence that the constriction was nearly as wide and as deep as it is now during the time of peak flood stage, judging from a gravel bar emanating from a high divide crossing the west shoulder of the gap and descending to the canyon bottom just downstream of the narrowest part of the constriction (Bretz, 1969, p. 535). This evidence indicates the present geometry of the constriction has not enlarged substantially since peak flow, thereby reducing the uncertainty in discharge estimates owing to canyon widening and deepening after emplacement of maximum-stage evidence.

3.2.10. Columbia River Gorge and Portland Basin

Downstream of Wallula Gap, the Columbia River flows through a series of basins and canyons, including the 1000-m-deep Columbia River Gorge where the river crosses the volcanic arc of the Cascade Range. West of the Cascade Range, floods entered the broad Portland basin, hydraulically ponded here behind constrictions farther downstream.

Where the floods coursed through this long-established valley, the erosional features are not as dramatic as in the Channeled Scabland. Nevertheless, large bouldery gravel bars, high eddy deposits and divide crossings, scabland tracts, ice-rafted erratics, and rhythmically bedded slackwater deposits are common for the 300 km between Wallula Gap and the Portland basin (Bretz, 1924, 1925b, 1928a; Hodge, 1931, 1938; Allison, 1933, 1941; Newcomb, 1969; Allen et al., 2009; O'Connor and Waitt, 1995; Benito, 1997; Benito and O'Connor, 2003; O'Connor and Burns, 2009). Floods entering the broad Portland Basin deposited a ~400 km^2 complex of bars, now under the cities of Portland, Oregon, and Vancouver, Washington (Bretz, 1925b, 1928a; Trimble, 1963; Allison, 1978; Evarts and O'Connor, 2008; Evarts et al., 2009). Impounded by downstream constrictions, the floods backflooded 200 km *up* the Willamette Valley (Fig. 1) to the present site of Eugene, Oregon (Allison, 1935; Glenn, 1965; O'Connor et al., 2001; Minervini et al., 2003).

As marked by erratics and divide crossings, the maximum flood stage descended from about 360 m upstream of Wallula Gap to ~120 m in Portland basin (Table 3; Benito and O'Connor, 2003, p. 634). Most of this fall was in the 130-km-long Columbia River Gorge between The Dalles and Portland, where maximum flood stages descended from 340 to 120 m.

The barely detectable descent of maximum stages—only ~20 m over 200 km between Pasco basin, through Wallula Gap, and to the head of the Columbia River Gorge near The Dalles—prompted Ira Allison's hypothesis "that the ponding was produced by a blockade of ice in the Columbia River [Gorge]" instead of a "large, short-lived catastrophic flood" (Allison, 1933, p. 676–677). Hydraulic modeling by Benito and O'Connor (2003) and Denlinger and O'Connell (2010) shows, however, that tight constrictions in the Columbia River Gorge hydraulically ponded floodwater upstream to Wallula Gap and beyond, further raising flow stages in Pasco basin. For the largest floods, flow was likely hydraulically "critical" through some of the narrow constrictions of the Columbia River Gorge.

Stratigraphic studies in the Columbia River Gorge by Benito and O'Connor (2003) were motivated by the relatively simple hydraulic conditions—a single flow route through an existing minimally eroded valley hydraulically controlled downstream by tight constrictions. This situation enables evaluation of the geologic effects and deposits of the floods relative to local hydraulic conditions (Benito, 1997). Evaluating evidence of multiple floods in combination with 25 radiocarbon analyses (mostly detrital organic materials), Benito and O'Connor (2003), p. 637) concluded that at least 25 floods had discharges exceeding 1 million m^3/s, and at least six or seven topped 5–6 million m^3/s, two postdating the Set S tephra. At least ten floods had discharges less than 3 million m^3/s, all likely younger than Set S. The largest floods(s) of about 10 million m^3/s post-date humates from a soil clast that gave a calibrated age of 23.4–22.5 ka. Dates from detrital organic materials in gravel flood bars are as young as 16.9–16.2 ka. A solitary ^{10}Be TCN age of 34.1 ± 2.6 ka for an ice-rafted erratic in the Columbia River Gorge is inferred too old because of inheritance (Balbas et al., 2017, Table DR2).

Stratigraphy in the Portland Basin and adjacent backflooded valleys is like that in the Pasco Basin but correlation is handicapped by being outside the Set S tephra plume. Within the 1200-km^2 Portland Basin the distribution of flood deposits is summarized by Evarts et al. (2009) and shown on the geologic map compilation of Wells et al. (2020). The distribution of ice-rafted erratics, which reach heights of 120 m, is shown by Allison (1935) and Minervini et al. (2003). Deposits include broad channelized sheets and bars of foreset-bedded gravel in areas of main currents, and bedded silt and sand in areas of slacker currents. The most complete stratigraphic records obtained so far in this area are from the backflooded Willamette valley. In the northern Willamette valley, deposits as thick as 30 m underlie the valley bottom at elevation 50–60 m except where eroded by post-flood rivers, pinching out on adjacent slopes as high as 120 m (O'Connor et al., 2001, p. 20). Glenn (1965, p. 88) counted 40 flood beds at the River Bend section, but the number of beds diminishes to the south and up-valley (O'Connor et al., 2001, p. 22). An uncounted sequence of later Missoula flood deposits appears inset into the main valley fill (O'Connor et al., 2001, p. 25).

Here too geochronologic information comes mostly from radiocarbon dating supplemented by tephra stratigraphy, magnetostratigraphy and a few OSL determinations. In the northern Willamette valley the thick sequence of flood beds overlies Willamette River gravel and capping soil dated 26.6–26.0 ka (O'Connor et al., 2001, p. 25). Megafauna bones collected from postflood bogs are as old as 14.5–14.1 ka (Gilmour et al., 2015, Table 1, UCIAMS78132 X). They also report an older post-flood megafauna bone age of 15.1–14.4 ka from the subsidiary Tualatin valley (UCIAMS78124 X). These Willamette valley post-flood ages are consistent with a less-precise 16.6–14.6 ka age reported by Mullineaux et al. (1978, p. 178) from peat, as well as ages of 14.0–13.8 ka and 13.8–13.6 ka reported by Archaeological Investigations Northwest (2015) for post-flood deposits in the northern Portland Basin. Three OSL ages from a sequence of slackwater flood beds in the Tualatin valley give ages ranging from 19.7 ± 2.5 ka to 16.1 ± 1.3 ka (Wells et al., 2020).

Hanson (2013, p. 79–129) examined the magnetostratigraphy of the lowermost 19 beds of the 40-bed River Bend section studied by Glenn (1965) and a better exposed section 750 m north. Adopting a similar procedure and reference curve as Clague et al. (2003), the best-fit correlation "indicates that the 19 floods span approximately 600 years between 13.7 and 13.1 ^{14}C ka BP," which calibrates to 16.7–16.3 ka to 15.9–15.6 ka. Because the base of the flood beds is exposed here (Glenn, 1965, p. 65–99; O'Connor et al., 2001, p. 17), and the magnetostratigraphy is from the base of the section, this age range should encompass the earliest floods of the last-glacial flood episode.

Overall, the geochronology for the Columbia River Gorge and the Portland Basin is like that upstream, especially as estimated from the radiocarbon dates. All or most floods were subsequent to 23 ka and flooding ceased by about 14.5 ka. Floods both preceded and post-dated the ~16 ka Set S tephra. Larger floods preceded smaller floods. No deposits of older cataclysmic flooding have been found in the Columbia River Gorge or Portland basin, though Cordero (1997) attributes unconformities in a loess sequence in eastern Columbia River Gorge to pre-last-glacial flooding (also described in Medley, 2012, p. 109–112; 115–116).

3.2.11. The eastern Pacific Ocean

Down-Columbia floods jetted out of the western Columbia River Gorge, ponded in the Portland Basin, and exited northwest through Kalama narrows (Evarts et al., 2009, p. 8). They then continued down the Columbia valley to the Pacific Ocean (Fig. 1). Missoula flood deposits are scarce downstream of Portland but traceable to Longview at Columbia River Mile 67, where backflooding up the Cowlitz valley deposited rhythmites and erratics as high as 60 m (Table 3). Downstream of Longview, flood deposits are likely below the modern estuary level, buried by Holocene valley fill accumulated during post-glacial sea-level rise. Such fill extends 130 m below sea level at the present Columbia mouth (Baker et al., 2010), and is about 70 m deep 170 km upstream near Portland (Peterson et al., 2011).

In the eastern Pacific Ocean, however, extensive areas of Columbia-basin flood deposits are revealed by ocean-floor cores (Griggs et al., 1970; Brunner et al., 1999; Zuffa et al., 2000; Normark and Reid, 2003; Hendy, 2009; Cosma and Hendy, 2008; and Gombiner et al., 2016). Floods and resulting turbidity currents entered the eastern Pacific by way of the submarine Willapa and Astoria Canyons (Beeson et al., 2017), built up the Astoria Fan (Prytulak et al., 2006), and continued ~1500 km down the Cascadia Channel to the abyssal plains of the eastern Pacific Ocean (Brunner et al., 1999; Normark and Reid, 2003). These deposits—sandy turbidites from sediment-laden density flows—are also composed of multiple fining upward sequences; ~20 late Pleistocene turbidite megabeds in the Escanaba trough decanted off the Cascadia Channel, including one bed 57 m thick (Zuffa et al., 2000, p. 257, 264; Normark and Reid, 2003, p. 621). About 400 km northwest of the Columbia mouth, core MD02-2496 on the continental slope off the coast of Vancouver Island also records many flood beds (Cosma and Hendy, 2008). Counting and provenance assessment by Gombiner et al. (2016) show 44 beds to be from the Columbia River basin, probably hyperpycnal Missoula flood sediment swept north from the Columbia River mouth by the California Current.

Columbia Basin megaflood chronology in the eastern Pacific chiefly derives from radiocarbon-dated cores ODPH 1037B (Zuffa et al., 2000, p. 256) and MD02-2496 (Cosma et al., 2008; Gombiner et al., 2016). Within the ODPH 1037B core in the Escanaba Trough (encompassing units B and C of Zuffa et al., 2000), a radiocarbon date of 25,700 ± 100 (planktonic foraminifera) that calibrates to 29.3–28.5 ka underlies all 20 flood beds by about 80 m. Within the flood-bed sequence are three radiocarbon ages that give calibrated ages, bottom to top, of 18.8–18.4 (planktonic foraminifera), 19.2–19.0 (pinecone fragment), and 12.9–12.5 ka (planktonic foraminifera)[5]. The three dates from planktonic foraminifera have an additional uncertainty of a specified reservoir correction factor that has likely varied over time (e.g. Cosma et al., 2008, p. 943). The 57-m thick bed is between the two older ages, thus about 19 ka if the ages are valid. About three flood beds overlie the 12.9–12.5 ka date. Chronology for the MOD02-2496 core off Vancouver Island is detailed in Cosma et al. (2008). From this chronology and the core stratigraphy, Gombiner et al. (2016, p. 135) counts 44 flood beds emplaced over ~4400 yr between 19.3 and 14.9 ka and having a periodicity of 50–80 yr. Both sites likely lack deposits of smaller Missoula floods from late in the flood sequence.

Late Pleistocene freshwater pulses—perhaps owing to megafloods—are also detected in eastern Pacific Ocean cores by proxy measures of salinity. From a pair of cores, Lopes and Mix (2009, p. 78) detect peak episodes of freshwater influence at 17.5 ka, 20 ka, 23 ka, 27 ka, and 30.5 ka. They attribute the youngest freshwater pulse at ~17.5 ka to Missoula flooding, but it could also owe to the ~18 ka Bonneville flood, which released twice the volume of the largest Missoula flood. The earlier freshwater peaks, like the large 23 ka pulse, likely relate to increased runoff or outburst floods during expansion of a still far-north Cordilleran ice sheet. More recently, Praetorius et al. (2020) summarize records from ten cores in the northeastern Pacific, finding sustained freshening of surface waters between 19.0 and 16.5 ka and accompanying increases in deepwater radiocarbon age, indicating suppressed ventilation of abyssal waters. Moreover, their ocean-circulation modeling of megaflood and ice-sheet runoff into the Pacific during this period suggests far-ranging oceanic cooling from freshwater input from the Columbia River basin, possibly triggering oceanic circulation changes.

Normark and Reid (2003, p. 634) estimated a total volume of last-glacial flood deposits in the eastern Pacific Ocean as 1450 km^3, including about 700 km^3 from a single flood leaving the 57-m thick bed in the Escanaba Trough (but not including the deposits assessed by Gombiner et al., 2016). Assuming a 50% porosity for this sandy unit implies that the entrained sediment volume for this large early flood was about 15% of the maximum plausible total water volume of 2500 km^3, the approximate maximum released volume of glacial Lake Missoula. Thus, this flood (and likely dozens of others) was exceptionally turbid.

3.3. Missoula flood timing synopsis

Distillation of the broad and varied work on the timing of events associated with the Missoula floods leads to some general conclusions as well as continuing uncertainties (e.g. Baker et al., 2016, p. 10–13; Waitt, 2016, p. 438–440). Glacial Lake Missoula was extant and producing floods for at least 3–4 ky during 20–14 ka, as evident from the glacial chronologies, the lacustrine records of glacial Lakes Missoula and Columbia, and from dating of flood deposits along the flood route. The lake was gone by the time of the 13.7–13.4 Glacier Peak tephra fall. Dozens of floods preceded, and dozens post-dated the ~16 ka Mount St Helens Set S tephra fall. Most floods entered glacial Lake Columbia, impounded by the Okanogan lobe for 2–5 ky, most likely between about 18.5 and 15 ka. Glacial Lake Columbia outlived the last flood from glacial Lake Missoula by more than 200–400 yr. At least one flood came down the Columbia valley before Okanogan ice blockage at about 18.5–18 ka. This flood (or floods) and the four or more Moses Coulee floods that soon followed were the earliest floods of the last glacial sequence. At least 20 floods down the Cheney-Palouse tract were after the 18.3–18 ka Bonneville flood and at least seven recorded in the Tucannon valley were after the ~16 ka Set S tephra. For a few centuries during the time of glacial Lake Columbia, glacial Lake Spokane was apparently impounded by the upstream Columbia ice lobe, long enough to receive about 15 flood beds in backflooded Latah Creek valley. The period 17–15 ka was especially dynamic, coincided with the maximum extent of the Okanogan and Purcell Trench lobes, many Missoula floods (particularly in the Cheney-Palouse tract and Grand Coulee), substantial erosion and possibly deepening of upper Grand Coulee, and widespread tephra falls from Mount St. Helens eruptions.

Some timing inconsistencies, issues, and opportunities include the following:

- The local evidence of mid-Wisconsin floods (e.g. Kiver et al., 1991, p. 241–243) appears inconsistent with the established glacial chronology and formation of glacial Lake Missoula. Because this evidence is at least partly based on radiocarbon dating of detrital organic materials, it may be a consequence of old carbon. Such reworking was called upon to explain the very first reported radiocarbon age for scabland flooding—a 32,700 ± 900 ^{14}C yr age from a clast of peat that Fryxell (1962, p. 118) concluded was "reworked deposits from an interstadial bog… preceding the last Wisconsin advance of the Okanogan lobe". Nevertheless, some of the discounted TCN ages (Balbas et al., 2017, Table DR2) as well as unconformities within loess stratigraphy attributed to flood episodes also date to the middle Wisconsin (McDonald et al., 2012), keeping the question open.
- The youngest 12.9–12.5 ka age from radiocarbon-dated offshore flood deposits reported by Zuffa et al. (2000, p. 258) is younger than terrestrial ages showing that all Missoula floods preceded the 13.7–13.5 ka Glacier Peak tephra. The marine ages may be affected by uncertainties in marine ^{14}C reservoir corrections.
- Some of the OSL ages from the region are outside the ranges just summarized, most not by much (e.g. Smith et al., 2018; Hanson et al., 2012; Wells et al., 2020). OSL dating offers great potential for understanding the floods, but discrepancies suggest more work is needed in understanding the application of OSL to Missoula flood deposits.
- The average TCN ages for the maximum extent of the Okanogan lobe at 15.4 ± 1.4 ka and Purcell Trench lobe at 15.7 ± 1.3 ka (Breckenridge and Phillips, 2010; Balbas et al., 2017, Table DR2) seem young relative to the radiocarbon-derived chronology. This is discrepancy is most evident relative to stratigraphic position of the ~16 k Set S tephra near

[5] The three radiocarbon ages of planktonic foraminifera reported here from Zuffa et al. (2000, p. 258) are calibrated using CALIB 7.1 (Stuiver et al., 2020) using the MARINE13 calibration curve and a marine reservoir correction factor delta R of 437 ± 50, as indicated by the closest determination included in the 14CHRONO database at http://calib.org/marine/.

the top of the glacial Lake Columbia section at Manila Creek (Hanson, 2013, p. 90, 116, 206), suggesting the demise of glacial Lake Columbia (and a Okanogan lobe terminus north of the Columbia valley) 300–500 yr after the tephra fall—about the same time as the TCN ages indicate ice pull-back from maximum positions. Post-deposition erosion of the TCN-dated surfaces or uncertain production rates could bias TCN ages to younger values. Another possibility is that the Set S tephra is younger than 16 ka, or possibly glacial Lake Columbia lasted longer than implied by the varve counts and tephra position in the Manila Creek section.

- Interbedded within Missoula flood rhythmites are at least four, and possibly five Mount St. Helens tephras (Mullineaux et al., 1978; Moody, 1978, p. 43–44; Moody, 1987, p. 43–44; Waitt, 1980, p. 664; Waitt, 1985a, p. 1274–1275; Bjornstad et al., 1991, p. 231; Bunker, 1982; Baker and Bunker, 1985). These tephras include the prominent So-Sg couplet of the Swift Creek Stage eruptions of Mount St. Helens. This couplet of Set S ashes is a prominent stratigraphic marker enabling correlation among flood slackwater sediments filling backflooded basins through much of eastern Washington and northeastern Oregon. It offers potential to improve counts of floods (e.g. Clague et al., 2003) and to improve understanding of flood magnitude (e.g. Waitt, 1985a, p. 1284; Benito and O'Connor, 2003) and the timing of various flood routes. Clynne et al. (2008, p. 619) summarize available age information and ambiguities for the Set S, concluding it dates to ~16 ka but note discrepancies among reported ages of up to ~500 yr in both directions. A better age determination for this important and widespread stratigraphic marker would refine Missoula flood timing. Finding traces of this tephra in sections within the glacial Lake Missoula basin could enable more precise correlation of the lake history to flood and lake deposits along the flood route.
- Interflood intervals indicated by varve counts between flood deposits and evidence of lake level fluctuations, paleomagnetic secular variation, and the offshore record summarized by Gombiner et al. (2016) suggest periods of a one to two years up to 100+ yr between floods or lake releases. Most varve records show a general up-section count increase followed by a more prolonged overall decrease in counts. The decreasing trend implies progressively shorter lake-formation times, smaller volumes, and hence smaller floods in conjunction with a thinning Purcell Trench lobe ice dam (e.g. Atwater, 1986, p. 18, 22). In general, the maximum varve counts are 40–60 in glacial Lake Columbia (Atwater, 1986, p. 11), glacial Lake Spokane (Waitt, 1984, p. 53), and at the Ninemile section of glacial Lake Missoula (Fig. 7b; Chambers, 1971, appendix III); although Kiver et al. (1991, p. 240) report as many as 125 varves between flood beds in the Latah Creek section likely formed in glacial Lake Spokane. The 40–60 yr maximum interval is less than those implied by other records, including: (1) The duration and varve counts in the Mission Valley section of Lake Missoula deposits where Levish (1997, p. 97–108) infers 66 lake cycles recorded by thickness variations in a sequence of ~2670 varves. Here, the average cycle is about 40 varves, but several have more than 60 and one has 107; (2) Paleomagnetic secular variation; Hanson (2013, p. 76–77, 114) suggests a best-match correlation giving a 55-yr flood recurrence interval within the uppermost ~46 flood beds and intervening lacustrine deposits at the Manila Creek section, which is close to Atwater's varve counts for the lower part of the section but double or triple his counts for the upper part of the section measured by Hanson. Hanson (2013, p. 110) similarly determines a 55–80 yr refilling interval at the Rail Line section of glacial Lake Missoula; (3) The marine record described by Cosma and Hendy (2008, p. 52) and Gombiner et al. (2016, p. 135) shows a "stable" periodicity of ~80 yr separating the lowermost 22 flood beds deposited between 19.3 and 17.6 ka. Some of these discrepancies among the sites may owe to erosion of varves by floods (e.g Atwater, 1986, p. 11; Hanson, 2013, p. 39), or possibly incomplete flood records offshore (Gombiner et al., 2016, p. 35).

3.4. How many floods?

Bretz first focused on making the case for a single debacle. By his final papers, though, he counted seven or so from geomorphic relations, including pre-last-glacial floods (Bretz et al., 1956, p. 967, 1045–1048; Bretz, 1969, p. 527–530). Work in the 1970s by Baker (1973, p. 42; 1978b, p. 33–35) and Waitt (1977b) countered some of Bretz's evidence for multiple floods, whittling the number down to three or less last-glacial Missoula floods (Waitt, 1980, p. 653).

These early interpretations, chiefly based on geomorphology, have been greatly augmented by stratigraphic analyses of flood and lacustrine deposits. Jerry Glenn's unpublished 1965 Willamette Valley dissertation and Waitt's (1980) analysis of the Burlingame Canyon section in the Walla Walla River valley (Fig. 14) were the first to infer dozen of floods, about forty in each case, from tall exposures of layered slackwater deposits. Waitt's then-controversial finding (e.g. Bjornstad, 1980; Baker and Bunker, 1985; Waitt, 1985a; Smith, 1993; Clague et al., 2003) was soon bolstered by the stratigraphy of glacial Lake Columbia (and possibly Spokane) showing dozens of flood beds separated by varves (Rigby, 1982; Kiver and Stradling, 1982, 1985; Atwater, 1984, 1986, 1987; Waitt, 1984, 1985b; Steele, 1991). The glacial Lake Columbia sections studied by Atwater (1986) indicate 89 separate flood beds sequenced among 2,000–3,000 varves, including 15–17 post-dating the Mount St. Helens Set S tephra (Hanson, 2013, p. 90, 116). Other high counts include 62 beds counted by Bjornstad (1980) in the Walla Walla valley near Touchet and the 44 beds in the marine core off Vancouver Island (Gombiner et al., 2016).

The position of Set S tephra within stratigraphic sections allows us to add to the 89 flood beds counted at Manila Creek. In the backflooded Walla Walla valley, more than 28 rhythmites preceded the Mount St. Helens Set S tephra at Burlingame Canyon (Waitt, 1980, p. 659) and at least 31 rhythmites overlying the tephra were counted by Clague et al. (2003, p. 248) near Touchet, summing to at least 59 floods. Adding these 31 post-Set-S beds in the Pasco Basin to the 72 preceding Set S at Manila Creek indicates at least 103 floods. Further adding the one or more floods down the Columbia before complete blockage by the Okanogan lobe means at least 104, possibly 108 floods if the four Moses Coulee floods are not recorded in the glacial Lake Columbia stratigraphy at Manila Creek. Most or all of these records are likely incomplete so the actual flood total is likely greater.

The multiple-flood record accords with the record of repeated emptying of glacial Lake Missoula. The most complete stratigraphic records, particularly the Ninemile (Fig. 7b) and Rail Line sections west of Missoula show ~40 lake cycles of slow filling followed by rapid emptying (Alt and Chambers, 1970; Chambers, 1971, 1984) and possibly as many as 80 (Hanson et al., 2012, p. 79). Levish (1997, p. 98–108, irregular pagination) describes 66 "packets" of thinning-up varve sequences in a Mission Valley section (Fig. 1), but attributes these to surging glaciers modulating sediment supply rather than lake-level changes (p. 112–118). From varve counts, Atwater (1986, p. 22, 32; 1987, p.186) proposes a bed-to-bed correlation between the Manila Creek section of interbedded flood and lacustrine deposits and the Ninemile section recording filling and emptying cycles of glacial Lake Missoula.

3.5. Flood magnitudes and routing

How big were the Missoula floods? Thought huge from the very beginning, epithets such as "great vigor," "great flood," and "debacle" (Bretz, 1923a, p. 573, 588; Bretz, 1923b, p. 649) instigated the early controversy regarding their possibility. But soon the estimates became quantitative and in recent decades more sophisticated. Nevertheless, the questions and controversies continue, exemplified by the recent Perron and Venditti (2016) claim of "Megafloods downsized." Also, how did flood sizes vary with timing and routings? These questions link flood size to the evolution of the Channeled Scabland and its flood pathways as well as the history and dynamics of glacial dams impounding glacial lakes Missoula, Columbia, and possibly Spokane. Missoula flood sizes have been studied by a variety of approaches, ranging from stratigraphic to computational. So far, the evidence shows that the sizes of the scores of last-glacial Missoula floods ranged widely in time and by place. Also emerging are more hypotheses, questions and conundrums, lighting paths for future work.

3.5.1. Geomorphic and stratigraphic evidence of flood size

Relative flood size has been inferred by stratigraphic and geomorphic relations, such as from sequences of flood rhythmites inset into coarser flood bars (Benito and O'Connor, 2003, p. 636), up-section thinning and fining of individual floods deposits, and by the relative position of the Set-S tephra within rhythmite sequences at different elevations (Waitt, 1985a, p. 1284). Varve counts between flood beds also indicate relative flood magnitude by clocking lake growth between releases of glacial Lake Missoula. For example, the Manila Creek section in the Sanpoil valley shows general trends in the number of glacial Lake Columbia varves between flood beds (Atwater, 1986, p. 17, 22, 29). Here, the number of varves—probably marking years—ranges from 20 to 40 very low in the section, increases to a maximum of 45–55, and then gradual declines (with short reversals) to only one or two varves between the uppermost flood beds (p. 11, 22). The longer durations suggest bigger lakes and probably larger floods, also signaled by thicker flood beds (Atwater, 1986, p. 18, 22, 27). These stratigraphic and geomorphic approaches show that the Missoula floods perhaps first grew in volume but surely became smaller with time. The last few floods were likely much smaller, involving just a year or two's worth of accumulated water in glacial Lake Missoula, compared with ~50–100 years or so indicated by the longest estimated flood intervals.

Pardee (1942, p. 1598) documented some of this disparity in lake release volumes and rates, noting that the coarse flood bars and giant current dunes emplaced by the "unusual currents" generated by at least one cataclysmic outflow (for example, Fig. 7c) preceded shorelines etched as high as 1080 m. Yet the preservation of shoreline features and associated flood features indicate that these later lakes must have drained more slowly or at least much less vigorously. Deposits associated with these lower lake levels include "varved silt" along the lower parts of stream valleys (Pardee, 1942, p. 1579–1580), locally inset into the higher coarser flood deposits (Smith, 2006; Hanson et al., 2012).

Continued mapping of flood landforms and deposits (e.g. Gordon et al., 2017; Hendrick et al., 2017; Doak et al., 2019), particularly that based on growing lidar availability for the region (Gleason and Markert, 2019), will likely refine geomorphic and stratigraphic relations among flood sizes and routes. Such mapping in conjunction with detailed stratigraphy (e.g. Waitt, 1980; Atwater, 1986; Smith, 1993; Smith, 2017) and geochronology onshore and offshore (e.g. Balbas et al., 2017; Cosma et al., 2008) will provide solid ground supporting inferences drawn from computational flood analyses.

3.5.2. Hydraulic analyses

Starting with Bretz (1925b, p. 257–258), most quantitative estimates of flood discharge derive from hydraulic principles and models, relying on evidence of flood stage in conjunction with estimated channel geometry (Table 2). This approach was adopted in Baker's Ph.D. dissertation in which he employed hydraulic engineering approaches to flow calculation for estimating scabland flood conditions, including sediment transport and current dune formation (Baker, 1973). Other approaches contribute, including sediment-transport criteria, bedrock-erosion thresholds, canyon-formation constraints, and ice-dam erosion modeling. A main emphasis has been to estimate maximum flood discharge for the entire flow, hence efforts focused on: (1) reaches pertinent to maximum lake discharge, including the ice dam itself (Clarke et al., 1984; Beget, 1986), the Rathdrum-Spokane valley just downstream from the ice dam (Baker, 1973; O'Connor and Baker, 1992; Denlinger and O'Connell, 2010), and Eddy Narrows in the Clark Fork Valley within the downstream arm of glacial Lake Missoula (Pardee, 1942, p. 1596); and (2), downstream of Pasco basin where all flood routes converged at Wallula Gap (Bretz, 1925b, p. 257–258; Baker, 1973; Craig and Hanson, 1985, p. 52–53; O'Connor and Baker, 1992; Denlinger and O'Connell, 2010) and downstream within the Columbia River Gorge (Benito and O'Connor, 2003). These areas mostly avoid the complexities of the multiple flood pathways of the Channeled Scabland.

Overall the general finding from the hydraulic analyses is that emplacement of the maximum stage evidence indicates a peak discharge of about 20 million m^3/s near the breached ice dam and about 5–15 million m^3/s through Wallula Gap and downstream in the Columbia River Gorge (Table 2). A further constraint is that Eddy Narrows, the Clark Fork valley constriction in the downstream arm of glacial Lake Missoula (Pardee, 1942, p. 1572, 1594–1597), can at most pass about 30 million m^3/s at critical flow associated with the maximum 1295-m level of glacial Lake Missoula, thus giving an upper limit for the rate of sustained lake outflow. These values place the Missoula floods atop the list of known terrestrial outburst flood discharges (O'Connor et al., 2013). Wide ranges among these values in part reflect different boundary conditions, assumptions, and calculation approaches adopted by different efforts. Uncertainties also owe to the exceptionally energetic and dynamic flows and likely high sediment concentrations, but these are less understood (Carling et al., 2003). Much lower than these hydraulically based calculations are the theoretical and empirical estimates for the peak discharge associated with a subglacial release of the lake by thermal and mechanical erosion of the buoyed ice dam, which for boundary conditions like those of the hydraulic estimates, range from 2.1 to 3.5 million m^3/s, 10–20% of the hydraulic estimates (Clarke et al., 1984; Beget, 1986; O'Connor and Baker, 1992, p. 277).

Discharges calculated for specific flood pathways are tentative and wide-ranging but include about 5–12 million m^3/s for the Cheney-Palouse tract, 6 million m^3/s for the Telford-Crab-Creek track, 0.1–13 million m^3/s for Grand Coulee, 0.6–11 million m^3/s for Moses Coulee, and about 13 million for the Columbia valley *sans* Okanogan lobe near Wenatchee (Table 2). Some of the wide range is explicable by estimates that only account for portions of the flow within a pathway, such as those calculated by the canyon-formation modeling approach of a Lapôtre et al. (2016). In addition, the large ranges for Grand Coulee and Moses Coulee owe to fundamentally distinct calculation approaches explained below. Clear, though, is that these values of the individual flow routes add together to greatly exceed possible outflow from near the ice dam, showing the implausibility of any single flood emplacing the maximum flood evidence on all flood pathways.

The great range in discharge estimates for individual flood routes is partly a result of the wide range of discharges (and volumes) of individual releases of glacial Lake Missoula. For example, a low 0.13 million m^3/s estimate for Grand Coulee is based on the flow velocity required to transport sand of a flood bed within glacial Lake Columbia sediment mantling the floor of Grand Coulee (Atwater, 1987, p. 192). The lacustrine deposits enclosing this sand, however, are locally inset against a bouldery eddy bar built by more energetic floods down Grand Coulee. This eddy bar blocks the mouth of Northrup Canyon cataract complex, itself formed by floods of even higher stages that give the 13 million m^3/s discharge estimated from one-dimensional flow modeling. Even though the 0.13 million m^3/s likely underestimates the discharge associated with the flood bed enclosed within the glacial Lake Columbia deposits (Atwater, 1987, p. 192), and the 13 million m^3/s value from maximum flood evidence and present topography likely overstates the actual maximum discharge because of subsequent coulee enlargement, the 100-fold variation indicates the potential range of discharges and volumes of individual floods from glacial Lake Missoula. Such variation is broadly consistent with the varve-count ranges between flood beds in both glacial Lake Columbia and glacial Lake Missoula, implying order-of-magnitude variations in durations and thus volumes of Lake Missoula between releases (Atwater, 1986, p. 22).

The possibility of wide-ranging flood discharge magnitudes was specifically addressed in the Columbia River Gorge (Benito and O'Connor, 2003). Here, step-backwater flow modeling in conjunction with stratigraphic studies, geomorphic relations, and field evidence of maximum flood stages, shows that at least one flow attained 10 million m^3/s, six or more exceeded 6.5 million m^3/s, 15 were greater than 3 million m^3/s, and at least ten had discharges greater than 1 million m^3/s but less than 3 million m^3/s (p. 634).

3.5.3. Two-dimensional hydrodynamic flow modeling

Two-dimensional, hydrodynamic flow models are now a tractable approach for quantifying Missoula flood flows. Inspired by questions involving multiple flow routes and flood dynamics and enabled by digital topography and improved computer processing power and software, such models are simulating flow at scales ranging from individual flow tracts (e.g. Larsen and Lamb, 2016) to much of the terrestrial flood domain, including Lake Missoula (Baker et al., 2016, p. 15–17; Denlinger et al., in press). Even off-shore modeling for the Willapa Bay submarine canyon under the eastern Pacific Ocean has supported understanding of possible routes of Missoula flood turbidity currents (Beeson et al., 2017, p. 1724–1726). This recent work follows earlier two-dimensional applications by Craig (1987), Komatsu et al. (2000), Miyamoto et al. (2006, 2007), Alho et al. (2010) and Denlinger and O'Connell (2010). The advantage of two-dimensional flood modeling over previous one-dimensional calculations is that it enables evaluation of multiple, interacting flow routes and provides spatially explicit estimates of flow properties such as direction, velocity, shear stress and stream power. The dynamic capability of such models also gives information on the temporal variation of flow through the channel system as well as changes at specific locations.

Recent system-wide Missoula flood modeling gives insights on the overall evolution of the Channeled Scabland. The Denlinger et al. (in press) calculations employing the GeoClaw flow model (Berger et al., 2011) investigate several plausible Missoula-flood scenarios (Fig. 15). All model runs encompass the entire flood system, glacial Lake Missoula to the Pacific Ocean. All are based on overall end-member maximum-flood conditions of a maximum 1295–m glacial Lake Missoula draining from instant removal of the impounding Purcell Trench ice dam. The various runs allow glacial Lake Missoula to freely drain but reflect different conditions of blockage of the Columbia valley and Moses Coulee by the Okanogan lobe, as well as blocked and unblocked states for upper Grand Coulee.

All scenarios show floodwater escaping out of the Spokane and Columbia River valleys into the Cheney-Palouse tract (Denlinger et al., in press). The Telford-Crab-Creek scabland tract is only substantially flooded if glacial Lake Columbia is at a high level, here specified at ~680 m, requiring upper Grand Coulee either not yet incised to its current 470-m bottom elevation, or filled and blocked by ice of the Okanogan lobe. Likewise, Moses Coulee is flooded only when the Okanogan lobe blocks the Columbia valley, but not so far advanced to block floodwater into the coulee (Fig. 6c). Additionally, as postulated by Hanson (1970, p. 64, 79), if Grand Coulee is open, it diverts enough water such that Columbia valley stages are lowered by about 40 m and Moses Coulee is hard to flood, implying that large Moses Coulee floods preceded final cataract retreat and lowering of the entrance to upper Grand Coulee. Similarly, modeled stages for the flood or floods down the Columbia valley *sans* Okanogan lobe (Fig. 6b) more closely approach stages indicated by the highest ice-rafted erratics near Wenatchee (Figs. 6c and 8) if upper Grand Coulee is not yet fully cut. Problematically, no model trials yet produce flood stages high enough to match field evidence for maximum stages near Wenatchee, Pasco Basin, and the Portland Basin—the "volume problem" described below.

As for the simpler and earlier hydraulic assessments, these hydrodynamic models are hindered by uncertainties in boundary conditions (such as channel geometry at the time of the flow, positions of ice margins, and initial input flows); as well as hydraulic parameters such as boundary and internal friction losses, and sediment concentration—all poorly understood for flows so much larger than typically assessed with hydraulic models. Future modeling efforts will likely explore such uncertainties, but ultimately, flow modeling will continue to address various field-testable scenarios to resolve outstanding geologic questions such as progressive erosion and deposition, effects of ice-lobe and glacial-lake positions, effects of isostatic deformation, and possible water sources and sinks.

3.5.4. The volume problem

A persistent and perplexing result from the hydrodynamic modeling so far is that in all but the northern reaches it fails to model flood stages as high as field evidence for maximum flood stages (Baker, 2009, p. 406). This finding is most evident for the Pasco Basin (Fig. 15; Komatsu et al., 2000; Miyamoto et al., 2007; Denlinger and O'Connell, 2010; Denlinger et al., in press).The modeling scenario that comes closest is that for an advanced Okanogan ice lobe blocking the Columbia valley and the entrance to Moses Coulee in combination with upper Grand Coulee blocked with ice or rock. For these conditions, the maximum calculated stage in Pasco Basin is about 338 m, still 20–30 m short of the maximum-stage field evidence of 360–366 m. Other plausible ice margin and channel-evolution scenarios produce maximum model stages in Pasco Basin of 327–334 m. A Pasco Basin stage of 338 m implies a volume of 1065 km^3, 80% of the 1330 km^3 volume required by a 360-m ponding level (Table 1), an instantaneous volume shortfall of ~265 km^3.

Possible explanations for the volume shortfall? It is unlikely that uncertainties and approximations in modeling parameters and conditions, such as boundary and internal friction and grid resolution, could account for all the discrepancy (Miyamoto et al., 2007), but these possibilities should be investigated further. Perhaps Wallula Gap or narrows in the Columbia River Gorge downstream incised or widened during the floods such that first hydraulic ponding was higher and early floods left higher evidence. Yet the geologic evidence of pre-flood valley bottoms close to modern for Wallula Gap (Bretz, 1969, p. 535) and the Columbia River Gorge (O'Connor and Baker, 1992, p. 270) limits this explanation.

Flow bulking by entrained loess and valley-bottom sediment possibly increased flow volume. As noted above, the volume of offshore deposits associated with the thickest flood deposit implies a flood sediment concentration of ~15% by volume. Flow bulking would be most pronounced for the first flood(s) through the Cheney-Palouse and Telford-Crab-Creek scabland tracts, where loess erosion would be greatest. Here, the tops of remnant loess islands commonly stand 50–100 m above scabland basalt channels, enabling a crude estimate of ~200 km^3 for the total volume of eroded loess from the Channeled Scabland upstream of Pasco Basin. This total additional flood volume is likely insufficient to account for the >250 km^3 instantaneous volume shortfall in Pasco Basin indicated by the flow modeling at peak stage, particularly since the ~50% porosity of the in-place loess reduces its effective flow-volume boost. Regardless, such a great bulking scenario would probably be a singular event, creating a first scabland flood substantially more engorged with sediment than subsequent floods coursing across an already eroded landscape. More investigation of this possibility is nevertheless justified, including better offshore sediment volume estimates and perhaps provenance analyses to determine source areas.

Another possibility is multiple or different flood-water sources. This has been posed before, starting with Bretz's search for subglacial volcanism in British Columbia (Bretz, 1927b, p. 107). More recently, Shaw et al. (1999, p. 607–608) and Lesemann and Brennand (2009, p. 2436) proposed that "ice-sheet underbursts" in the Okanogan valley augmented Missoula flood flow, possibly accounting for the missing volume. Or perhaps a different and larger source independently sent a flood down the Columbia valley. The early flood down the Columbia *sans* Okanogan lobe is a candidate for an exceptionally large non-Missoula flood, but no such evidence for either a source or path has yet been described. Another plausible augmenting water source is glacial Lake Columbia or glacial Lake Spokane. An incoming Missoula flood compromising either impounding ice lobe could instigate a tandem release adding to glacial Lake Missoula flooding by the volume of the downstream ice-dammed lake—a possibility proposed by Atwater (1987, p. 199). Rip-up clasts of bedded sand and silt within the bouldery gravel under Okanogan lobe till in the Columbia valley led Hibbert (1985, p. 107) to hypothesize that an early flood entrained lacustrine sediment impounded by an earlier ice advance into Columbia valley. Glacial Lake Spokane seems too small to make enough difference, containing just 50 km^3 at a 600-m stage. Glacial Lake Columbia was possibly large enough; at a high, 715–730-m level, it would hold ~400–500 km^3 of water, perhaps enough to make up the Pasco Basin shortfall if released cataclysmically and added to a large Missoula flood. This scenario is plausible, if not likely, during early phases of glacial Lake Columbia before the Okanogan lobe completely filled the Columbia valley.

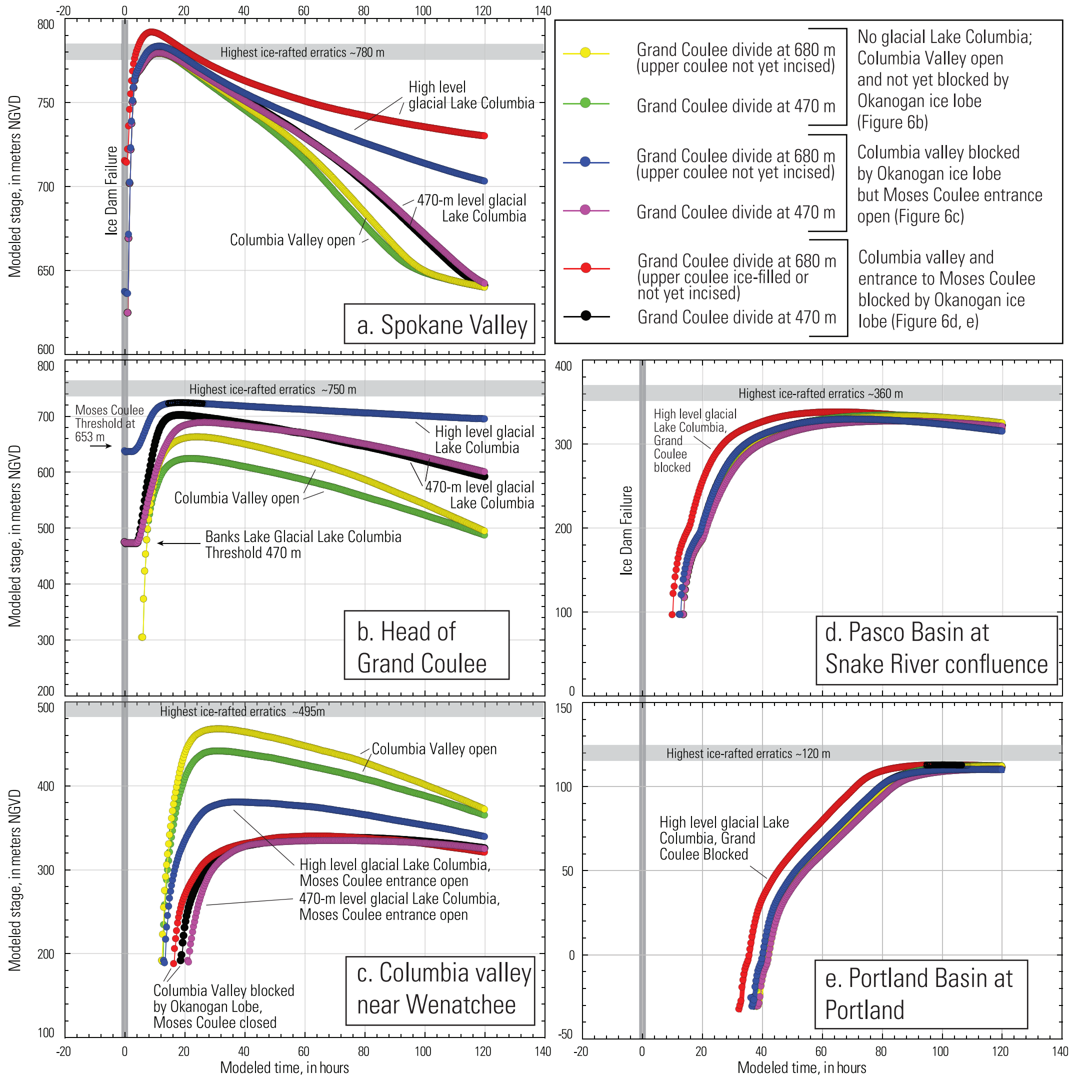

Fig. 15. Stage predictions from hydrodynamic modeling results of Denlinger et al. (in press) employing GeoClaw model (Berger et al., 2011). All model runs fully couple drainage of glacial Lake Missoula from its maximum 1295-m level and downstream flooding due to instantaneous and complete ice-dam removal. Model topography derived from 30 m digital elevation model adjusted for ice-margin positions derived from Waitt and Thorson (1983), reservoir inundation, and Holocene fill of the Columbia River valley below Columbia River Mile River 190 [after Baker et al., 2010 downstream of River Mile 120 and supplemented by drillers logs]. Model calculations implement twenty-minute model time steps. Ice-rafted erratics mark approximate maximum stages at depicted locations (Table 3). Three basic scenarios are modeled for each location: (1) Columbia Valley open to Missoula floods prior to blockage by the Okanogan ice lobe, as shown in Fig. 6b; (2) Columbia Valley blocked by Okanogan lobe but entrance to Moses Coulee open (Fig. 6c); and (3) Okanogan lobe blocking Columbia Valley and entrance to Moses Coulee (Fig. 6d). For each of these three situations, model trials assume that the Grand Coulee threshold is either at the modern 470-m elevation or at a higher, unincised or ice-blocked 680 m, a plausible estimate of the pre-incision elevation of the coulee entrance. (a) Spokane Valley near Spokane (47.6890, -117.1090), about 75 km downstream from ice dam. Maximum inundation is for Missoula flooding coincident with a high-level 680-m glacial Lake Columbia. (b) Columbia River at the head of Grand Coulee (47.9406, -118.9483). Highest stages attained when Columbia River valley downstream is blocked by the Okanogan lobe. No results shown for maximum Okanogan lobe ice extent (Fig. 6d) blocking Columbia Valley, Moses Coulee, and Grand Coulee because site is within presumed ice margin, but this scenario would produce highest local flood stages as indicated by Spokane-area results shown in part a. The two runs for the Columbia-River-open scenario produce the lowest local flood stages but also show the influence of a fully incised upper Grand Coulee diverting substantial flow, enough to reduce local maximum stage by nearly 40 m. (c) Columbia River near Wenatchee (47.4526, -120.3167). Highest flood stages result from unblocked Columbia Valley scenarios, allowing down-Columbia passage of Missoula floods. A Grand Coulee outlet elevation of 470 m diverts enough flow to reduce stages in the Wenatchee area by 27 m. Yet even when Grand Coulee diverts no water, maximum calculated stages fall 20–30 m short of the highest local ice-rafted erratics. The low, 470-m Grand Coulee threshold also markedly reduces Moses Coulee flooding, thereby eliminating backflooding to Wenatchee area, compared to stages 40 m higher when Moses Coulee entrance is open, but the Grand Coulee divide is high at 680 m. (d) Columbia River in Pasco Basin (46.1937, -119.0414). Maximum and most rapid inundation of Pasco Basin predicted for Missoula floods into high-level glacial Lake Columbia. Highest modeled floods levels of 338 m fall short of local field evidence of maximum stages by 20–25 m. (e) Columbia River near Portland (45.6498, -122.756). Similar to Pasco Basin, highest and most rapid flooding associated with high-level glacial Lake Columbia. Maximum predicted flood stages of 112 m about 10 m lower than highest nearby erratics.

3.5.5. An early "Mother of all Floods"?

Recognition of dozens of Missoula floods inspired Waitt's (1980) hypothesis of repeated jökulhlaups from a self-dumping glacial Lake Missoula. Bolstering the jökulhlaup case is an overall decline in flood size (inferred from deposit thickness) and intervals between floods (from varve counts) with time (Waitt, 1984, 1985a; Atwater, 1984, 1986, p. 29), consistent with repeated releases tunneling beneath a thinning Purcell Trench ice dam. However, O'Connor and Baker (1992; p. 277) and Benito and O'Connor (2003), p. 636–637) concluded from stratigraphy, elevations of maximum stage evidence, chronology, and flow modeling that the first Missoula flood (or floods) was perhaps "distinctly larger." O'Connor and Baker (1992, p. 277) suggested this larger flood—affably satirized by Waitt as the "mother of all floods" (personal communications to O'Connor, 1990–2020)—likely came down the Columbia valley before blockage by the Okanogan Ice lobe. Moreover, this inference prompted the hypothesis by O'Connor and Baker (1992, p. 277) that early rupture of the Purcell Trench Lobe ice dam was more cataclysmic and complete than a later self-dumping sub-glacial jökulhlaup mode of ice-dam failure.

Was the early flood preceding the Okanogan lobe's blockage of the Columbia, or any flood, "distinctly larger" and perhaps the result of a more cataclysmic mode of dam failure? Could such a flood address the volume problem? These questions are unresolved. As noted above, Missoula floods had greatly varying discharges. Flood magnitude apparently decreased markedly toward the end of the last glacial episode of flooding. The varve counts among the ~89 floods entering glacial Lake Columbia show an overall trend of a first increasing but then mostly diminished time interval between floods, consistent with floods getting larger then smaller as the damming Purcell Trench lobe thickens and thins, impounding a correspondingly varying glacial Lake Missoula (Atwater, 1984, 1986, p. 11, 29). The latter part of the glacial Lake Columbia record matches that of rhythmite sections along the flood route, which generally show up-section thinning of individual flood layers (Glenn, 1965; Waitt, 1980, 1985a; O'Connor et al., 2001) and is consistent with geomorphic relations in the Columbia River Gorge showing younger and smaller floods post-dating earlier larger ones (Benito and O'Connor, 2003).

Nevertheless it is evident that at least one early and very large last glacial flood came down the Columbia River valley undiverted by the Okanogan lobe, depositing the immense Pangborn and Brays Landing bars (Fig. 13) and emplacing erratics and eddy bars 250–320 m above modern river level for which preliminary hydraulic modeling indicates a peak discharge of 10–13 million m^3/s (Table 2). Also possible is that this was a multi-sourced flood combining volumes of glacial Lake Missoula and lakes Spokane and/or Columbia. TCN dating tentatively correlates (by timing) three high-level ice-rafted erratics from this flood (or floods) near Wenatchee at 18.2 ± 1.5 ka to a high erratic close to maximum flood stage at Wallula Gap dated at 18.2 ± 1.6 ka (Balbas et al., 2017). This flood (or floods) preceded those recorded in glacial Lake Columbia. It may be the flood that left the basal, 57-m thick flood deposit in the eastern Pacific (Zuffa et al., 2000; Normark and Reid, 2003).

As noted previously from geomorphic relations, Pardee (1942, p. 1598) inferred that initial drainage from glacial Lake Missoula was catastrophic, followed by filling and less vigorous emptying recorded in the low lacustrine deposits showing multiple lake-refilling episodes; a conclusion supported by stratigraphy and mapping by Smith (2006), geochronology (Smith et al., 2018) and hydraulic modeling by Alho et al. (2010). Yet Smith (2017) also showed that glacial Lake Missoula filled seven to 12 times to levels containing at least 65% of its maximum volume; and Benito and O'Connor (2003), p. 636) concluded from stratigraphy and modeling that at least six floods had discharges at least 65% of the maximum flow through the Columbia River Gorge. If a singular initial flood was largest, it was not larger by more than ~50% as measured by peak discharge than later floods.

Hydraulic modeling of flow exiting glacial Lake Missoula (Denlinger et al., in press) and modeling of glacial Lake Missoula jökulhlaups by Clarke et al. (1984) address the issue relative to failure mode of the ice-dam. Clarke et al.'s (1984) modeling, considering a variety of ice-dam geometries and parameters, predicted a plausible range of maximum discharge values of 2.6–15.3 million m^3/s for subglacial jökulhlaup-style releases at the ice dam. Beget (1986), from an empirical analysis of historical jökulhlaups, suggested the lower value more likely. These values are mostly lower than the minimum estimate of 14–20 million m^3/s based on high-water evidence for flow through the Spokane-Rathdrum valley reach just downstream from the ice dam (Table 2; O'Connor and Baker, 1992). Clarke et al. (1984, p. 294) described the possibility of catastrophic ice-dam failure caused by thermal erosion of the outflow tunnel to an extent that the "tunnel roof collapses during the flood and the dam is swept away," a possibility affirmed by calculations for maximum lake conditions by Waitt (1985a, p. 1282).

Ice dams release floods both subglacially and by catastrophic failure (Walder and Costa, 1996; O'Connor et al., 2013, p. 481–483). It is possible that both types of ice-dam failure affected the Channeled Scabland; one or more large early release(s) by complete and rapid dam failure initiated either by tunneling or overtopping in conjunction with the maximum stand of glacial Lake Missoula, followed by a hundred or more smaller subglacial jökulhlaup releases in which the ice dam was not entirely removed. The distinction might be recognizable by a question not yet addressed—can the sedimentology of the flood deposits aid in distinguishing the distinctive types of hydrographs produced by different dam-failure modes (e.g. Clarke et al., 1984, p. 294; O'Connor and Baker, 1992, p. 276)? Until more such work is completed, the "mother of all floods" scenario remains hypothetical.

3.5.6. Megafloods downsized?

A conclusion in the opposite direction in terms of flow size is proposed by Larsen and Lamb (2016). They apply a threshold bedrock-erosion model to estimate peak discharges for Missoula floods within Moses Coulee (Fig. 12), leading to "much lower flood discharges than previously thought." Hence, the "Megafloods downsized" leader in *Nature's* accompanying News and Views by Perron and Venditti (2016). Although their overall approach for modeling progressively eroding canyons is a valid new direction for flood research, their "downsizing" of previous Missoula flood discharge estimates relies on incomplete comparisons and assumptions.

Citing earlier estimates of Missoula flood discharges (e.g. Baker, 1973; O'Connor and Baker, 1992; Denlinger and O'Connell, 2010), Larsen and Lamb (2016, p. 229) state "most hydraulic modelling efforts [in bedrock canyons] implicitly assume that the canyons were filled with water to their brims." This statement, which follows from Lapôtre et al. (2016, p. 1232), misrepresents most Missoula flood calculations. Past Missoula flood (and Bonneville flood) modeling in key locations, such as Spokane-Rathdrum Valleys, Wallula Gap, and the Columbia River Gorge, does not rely on assuming a brim-full valley or canyon stage but is based on field evidence of maximum-flow stages such as erosional trim lines, divide crossings and ice-rafted debris (Table 2; Baker, 1973; O'Connor and Baker, 1992; O'Connor, 1993; Benito and O'Connor, 2003). Only for Wallula Gap do these stages approximately coincide with a canyon "brim". Their point, though, addresses the critical assumption for all retrodictive palaeohydraulic modeling—that the modeled channel geometry approximates that extant during emplacement of the highest high-water evidence (e.g. Carling et al., 2003). This assumption indeed leads to overestimates of peak discharge if the valley deepened or widened since maximum flood levels. Alternatively, underestimates result if the valley has filled, as possibly the case for Rathdrum Valley (O'Connor and Baker, 1992, p. 274–276). Despite this limitation, previous flow modeling in the Channeled Scabland has used the modern topography to represent the flow geometry at the time of maximum flow stages. An essential fact not acknowledged by Larsen and Lamb, however, is that for the key estimates of peak discharges, particularly for Wallula Gap and the Columbia River Gorge, strong geologic evidence shows that the modern valley geometry is indeed appropriate for estimating flow (e.g. Bretz, 1969, p. 535; O'Connor and Baker, 1992, p. 270, 274–275). These earlier estimates purposefully were made in valleys that attained nearly all their present depths and widths before the Missoula floods.

In further comparing their threshold bedrock-erosion approach to the "brim-full" model, Larsen and Lamb (2016, p. 229) suggest that brim-full estimates for Channeled Scabland canyons imply a series of floods of increasing size "to maintain brim-full flow while the canyon floor erodes." They describe this scenario as unlikely and inconsistent with the proposed lake-failure mechanisms. This assertion is true, but such a scenario has never to our knowledge been proposed, for it is inconsistent with the field evidence throughout the region. Nor has it been modeled for the Channeled Scabland. Indeed, the opposite has been demonstrated. Hydraulic and stratigraphic evidence throughout the area affected by Missoula floods shows that most earlier floods were larger than later floods (e.g. Waitt, 1985a, 1985b; Benito and O'Connor, 2003).

Mechanistic approaches are certainly a valid direction for increasing knowledge of the Channeled Scabland and other flood-formed landscapes. Threshold bedrock-erosion and canyon-formation models like that applied by Larsen and Lamb (2016) and Lapôtre et al. (2016) are promising approaches for investigating scabland channel erosion in erosional tracts such as Moses Coulee. Nevertheless, Larsen and Lamb's extension of the Moses Coulee analysis and conclusions to earlier estimates of megaflood peak discharges incompletely considers geologic context and key aspects of earlier work.

4. Columbia basin megafloods—other considerations

Although the marks of the Missoula and Bonneville floods are most visible, other floods, topics, and questions continue to be relevant for understanding ice age floods of the Columbia River basin.

4.1. Other last-glacial floods

Several other late Pleistocene floods besides the Missoula and Bonneville floods affected the Columbia River basin within both the Columbia and Snake River systems. Many of the known floods could benefit from more study. Many floods almost certainly have yet to be identified. Intriguing is the possibility of large, subglacial "ice-sheet underbursts" from under the main Cordilleran ice lobes (Lesemann and Brennand (2009, p. 2436).

Huge bars in the Columbia valley between the Methow River confluence and Pasco Basin attest to at least one large down-Columbia flood after the maximum advance of the Okanogan ice lobe. These bars crest as high as 80 m above the Columbia River, are armored with car-size boulders, and are locally mantled by giant current dunes (Figs. 13 and 16). The bars appear to post-date the last Missoula flood. Bretz et al. (1956, p. 993–994) described one of these bars, Beverly Bar, a boulder-studded mound of gravel blocking the Crab Creek's coulee entrance to the Columbia valley (Fig. 5). In this position Beverly Bar would be unlikely to survive substantial flow through Grand Coulee, flow that would then rejoin the Columbia via lower Crab Creek. This relation prompted Bretz et al.'s (1956, p. 994) hypothesis that Beverly Bar, as well as the similar-sized West Bar upstream, owed to "break-up of the Okanogan lobe dam" and rapid emptying of glacial Lake Columbia, an idea posed by earlier Waters (1933, p. 817). Bretz later (1969, p. 525), however, attributed the Beverly Bar flood to Missoula flooding during a temporary pull-back of the Okanogan lobe. Geologic mapping by Waitt (1987, 2016) strengthened Bretz's initial hypothesis, aided by Atwater's (1987, p. 185–187) evidence that glacial Lake Missoula's final flood was at least 200–400 yr before the demise of glacial Lake Columbia.

Conditions were ripe for such a flood after the last Missoula flood but while the Okanogan lobe was still blocking the Columbia River valley at the Okanogan River confluence. The resulting glacial Lake Columbia and the inflowing Columbia River exited through Grand Coulee over the modern coulee floor threshold of 470 m. Because of its outlet, the Okanogan ice-lobe dam and the impounded glacial lake were stable, with a lake surface ~240 m above the modern Columbia River at the Okanogan River confluence. Ice-dam stability diminished as it thinned, probably reaching a critical thickness enabling subglacial tunneling and perhaps failure of the ice dam. Glacial Lake Columbia at this level relative to modern topography would contain about 54 km^3 of water (Table 1), although this value likely overestimates the impounded water volume because of substantial Holocene incision of valley fill within the lake area. Nevertheless, a glacial Lake Columbia outburst is a logical source for coarse deposits traced down the Columbia valley because no such deposits are evident above the Okanogan River confluence (Waitt, 2016, p. 434–436). No evidence has yet been found indicating multiple floods, so a singular rupture of the ice dam may have been the complete and final demise of glacial Lake Columbia.

Available chronology also indicates that glacial Lake Columbia is a plausible source. The demise of the lake post-dated the ~16 ka Set S tephra and preceded the 13.7–13.4 ka Glacier Peak tephra. The downstream flood deposits also precede the 13.7–13.4 ka tephra (Waitt, 2016, 2017). TCN ages on of 11 boulders from five sites inferred deposited or remobilized by the post-Missoula flood range from 12.5 ± 1.2 to 23.1 ± 1.9 (Balbas et al., 2017, Table DR2). Discounting the single high outlier, the remaining ten ages average 14.4 ± 0.3 ka.

Much more work can be done on this flood. Its deposits have neither been systematically mapped nor evaluated. More checking upstream could evaluate the possibility of other water sources, including a possibly reformed glacial Lake Missoula (Atwater, 1987, p. 199). No hydraulic calculations have yet estimated the peak flow of this flood, but abundant field evidence makes this a tractable undertaking. A tentative estimate made by applying Walder and Costa's (1996, p. 708) regression for "non-tunnel drainage" floods from ice-dammed lakes (in contrast to sub-glacial tunneling releases) predicts a peak discharge of 0.13 million m^3/s for a lake volume of $5.4 \cdot 10^{10}$ m^3 associated with the final 470–m Grand Coulee outlet elevation (Table 1, Table 2).

Another later flood followed the same Columbia valley path, leaving bouldery flood bars less than 35 m above the Columbia River (Fig. 16b). These bars are inset against the higher bars likely left by the emptying of glacial Lake Columbia and appear to post-date the 13.7–13.4 Glacier Peak eruption since its tephra has not been found on top of the bars. Searches for a source are still incomplete but a plausible candidate is glacial Lake Kootenay occupying the trough now partly filled with Kootenay Lake in southern British Columbia (Waitt, 2016, p. 436–437). Glacial Lake Kootenay grew in front of the Purcell Trench ice lobe as it receded north, its way to the Columbia via the Kootenay River blocked by the lingering ice lobe. At its maximum, the lake covered 1152 km^2 with 142 km^3 of water as much as 400 m deep. Once the retreating Purcell trench lobe uncovered the outlet valley leading west, the lake drained, possibly cataclysmically. Peters and Brennand (2020) map evidence of the lake and outburst flood in the Kootenay River valley which they attribute to this scenario, although they name the lake "glacial Lake Purcell". Other ice-dammed lakes in the northern Columbia River basin also formed in deglaciated tributary valleys, blocked by lagging ice in the main valleys. Some, like 20 km^3 Elk Lake, dammed by the remnant Purcell Trench lobe, also appear to have emptied rapidly, possibly with a discharge as great as 60,000 m^3/s (Clague, 1975, p. 233–234). Evidence of these floods has yet to be traced downstream.

Additional Pleistocene-age floods from both ice-dammed and pluvial lakes entered the Snake River system. Lamb et al. (2008, 2014) attribute amphitheater-headed canyons entering the Snake River within the Snake River Plain to outburst floods possibly about 45 ka. Candidate sources for these floods are alpine-glacier-dammed lakes in the central mountains of Idaho, such as in the basins in the Lost River headwaters (Rathburn, 1993; Cerling et al., 1994) and Big Wood River (Othberg et al., 2012, p. 4). A large flood from pluvial Lake Alvord in eastern Oregon entered the Owyhee River, tributary to the Snake River, sometime after the ~16 ka Set S tephra fall. Like the Bonneville flood but much smaller, Lake Alvord overtopped and eroded an unconsolidated divide, releasing about 11.3 km^3 of water at a peak flow of about 10,000 m^3/s (Carter et al., 2006). A similar but smaller flood resulted from the breaching of pluvial Lake Millican in central Oregon, sending ~7 km^3 of water into the Deschutes and lower Columbia rivers, possibly in the middle Pleistocene (Vanaman et al., 2006). Smaller Holocene outburst floods from landslide- and moraine-dammed lakes have also affected rivers and streams of the Columbia River basin, summarized in O'Connor et al. (2013).

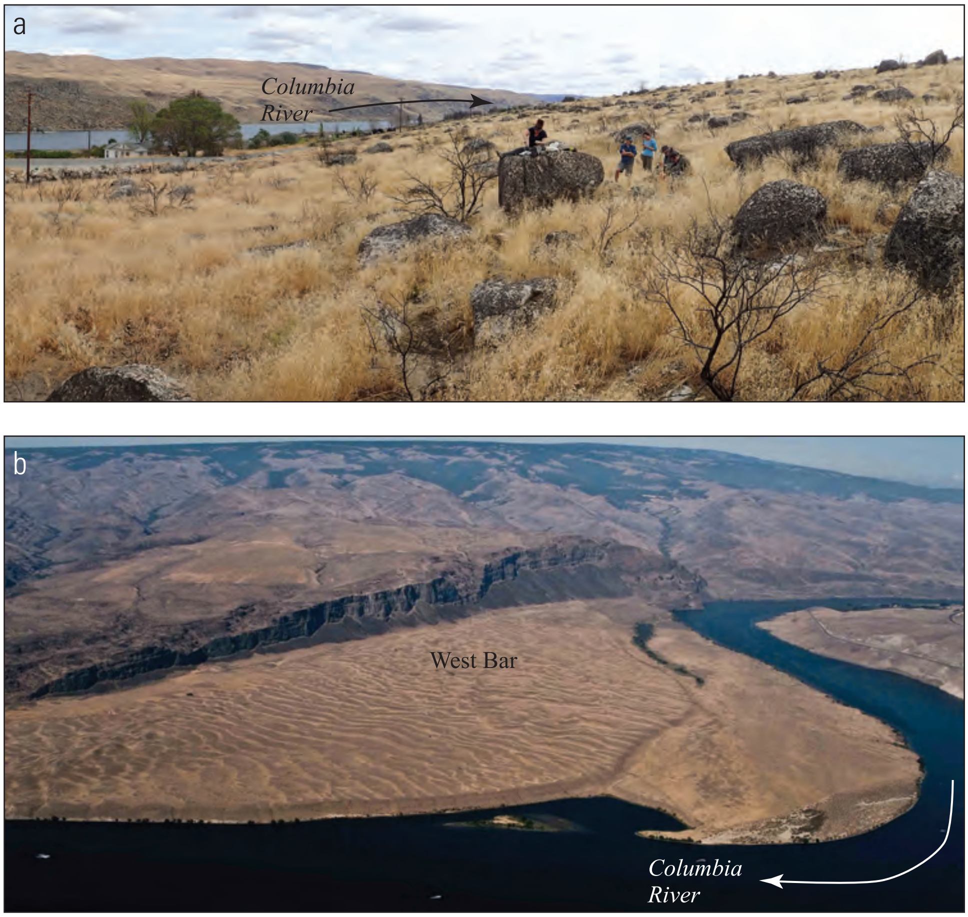

Fig. 16. Deposits left by last-glacial, probably post-Missoula, outburst flood in the Columbia valley which we infer derived from cataclysmic emptying of ice-dammed Lake Columbia. (a) Rounded bar of bouldery gravel downstream of Methow River confluence with Columbia near River Mile 523, 15 km downstream from ice-dam blockage area at the Okanogan River confluence. Terrestrial cosmogenic nuclide analyses of three boulders at this site give results of 13.0 ± 1.1 ka, 13.9 ± 1.1 ka, and 15.3 ± 1.3 ka (Balbas et al., 2017, Table DR2). Scatter among individual boulder ages possibly owes to variable transport and exposure histories of measured boulders perhaps moved by earlier floods or by the overriding Okanogan ice lobe. Photograph by Jim E. O'Connor. (b) Upstream directed aerial view by Bruce Bjornstad of rippled West Bar from near Columbia River Mile 440, inferred deposited by an outburst of glacial Lake Columbia. Bar surface stands 60 m above Columbia River pool level. Inset into West Bar is another flood bar, possibly left by an outburst from glacial Lake Kootenay (Waitt, 2016).

4.2. Older flood episodes

Megaflood evidence of the last ice age is most prevalent and studied. However, deposits of earlier large floods in the Columbia valley near Wenatchee, in Quincy basin, the Cheney-Palouse scabland tract, in Pasco basin, and possibly the Snake River Plain, show that megafloods have a long Quaternary history in the Columbia Basin, likely associated with earlier ice ages (Bretz et al., 1956; Richmond et al., 1965; Baker, 1973; Patton and Baker, 1978b; Baker et al., 1991; Bjornstad et al., 2001; Lamb et al., 2008, 2014; Medley, 2012).

The most compelling evidence of much older flooding is foreset-bedded gravel capped with thick loess sequences containing petrocalcic soil horizons (Fig. 17). Such outcrops are found in the Cheney-Palouse scabland tract (Patton and Baker, 1978b; Bjornstad et al., 2001; Baker et al., 2016, p. 51–52), along the Columbia Valley near Wenatchee, where foreset-bedded gravel capped with a thick petrocalcic horizon lies upslope of the massive last-glacial Pangborn bar, and in the western Quincy basin, where two old floods left carbonate-cemented east-dipping foreset gravels derived from floods spilling out of the Columbia Valley south of Wenatchee (Bretz et al., 1956, p. 985; Bretz, 1969, p. 524; Baker, 1973, p. 8–9).

Pasco basin contains the most studied and diverse evidence for floods throughout the Quaternary. This ranges from foreset-bedded boulder-gravel to rhythmite-like sequences of bedded silt and clay, typically distinguished from last glacial flood deposits by prominent calcic horizons, which last-glacial deposits lack (Bjornstad et al., 1991, p. 228–235; 2001, p. 707–711; Medley, 2012, p. 116–119). Some of the large last-glacial flood bars in Pasco Basin, such as Cold Creek bar, appear cored by older flood bars (Bjornstad et al., 2001, p. 707–709; Pluhar et al., 2006). Bjornstad et al. (1991, 2001) identify at least two episodes of pre-last-glacial flooding; a middle Pleistocene episode sometime between 130 ka and 780 ka, and an early Pleistocene episode older than 780 ka, preceding the Brunhes-Matuyama geomagnetic reversal. In the Walla Walla valley, Bader et al. (2016) identify two pre-last-glacial but post-220-ka diamicts bounding unconformities in a thick loess sequence. The diamicts contain extrabasinal clasts attributed to middle Pleistocene glacial-outburst floods.

Older flood episodes have also been inferred from the soil and loess stratigraphy flanking scabland channels (McDonald and Busacca, 1988, 1992; Busacca and McDonald, 1994; Gaylord et al., 2001; Sweeney et al., 2007; McDonald et al., 2012). Most evidence is indirect and derives from the extensive dune formation and loess deposition nourished by last-glacial flood deposits. Associated with last-glacial flooding is the prominent "L1 loess," which commonly contains the Mount St. Helens Set S tephra near its base, indicating active accumulation during Missoula flooding (McDonald et al., 2012, p. 20). They similarly attribute the underlying L2 loess to possibly larger glacial Lake Missoula outburst floods crossing the Channeled Scabland and into Pasco Basin during oxygen isotope stage 4

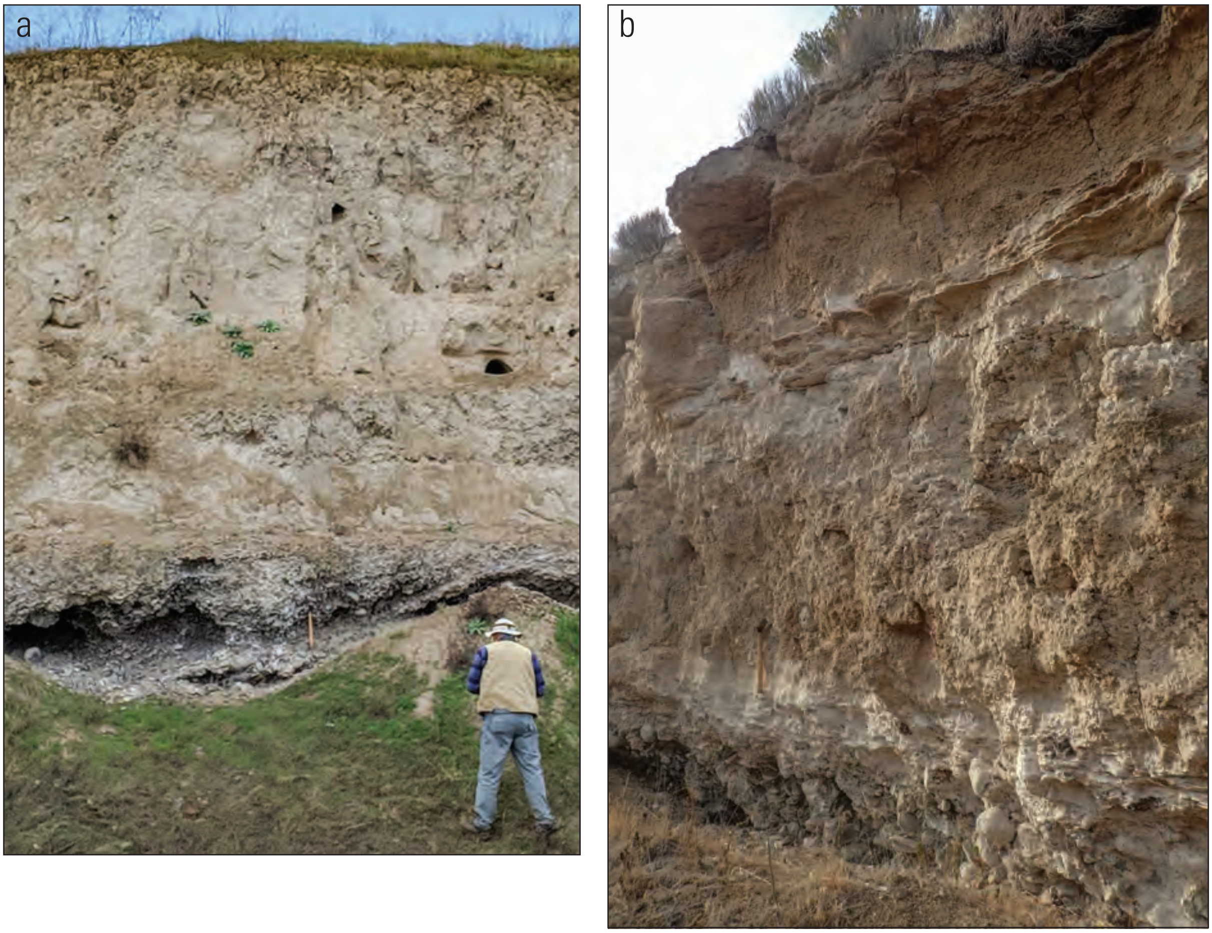

Fig. 17. Thick sequences of loess and paleosols containing strongly formed calcic horizons, capping poorly sorted but stratified and locally foreset bedded basalt cobble gravel. Such deposits indicate flood episodes prior to the last glacial period. Photographs by Jim E. O'Connor. (a) Old Maid Coulee site (46.6045, -118.8599) where Bjornstad et al. (2001, p. 698) reported flood gravel capped by reversed-polarity soil and >350 ka calcic horizon. (b) Marengo bar site within Cheney-Palouse scabland tract (47.0126, -118.2134), also reported by Bjornstad (2001, p. 698) as a flood gravel overlain by reversely magnetized loess and an 800-ka calcic horizon. Hammer 40 cm long.

between about 77 and 44 ka (p. 28), although equivalent age slackwater deposits have yet to be found (p. 24).

So far, evidence for older flooding is confined to the Cheney-Palouse scabland tract, the Columbia valley upstream of Pasco Basin (including that of floods spilling east into Quincy basin), and Pasco basin. Unresolved is if pre-last-glacial floods affected Moses Coulee or Grand Coulee. No older flood deposits have yet been found downstream of Pasco basin, possibly indicating such floods were smaller than the largest of the Missoula floods.

Also unknown is what might have been the sources of earlier floods affecting the Channeled Scabland. No evidence has yet been reported for pre-last-glacial versions of glacial Lake Missoula. Geologic mapping conducted in western Montana reports only late Pleistocene glaciolacustrine sediment, locally overlying flood gravels and bedforms (Berg, 2006, 2005; Lonn et al., 2010, 2007). Additional chronology of the flood gravel and deposits near the terminal positions of the Pend Oreille lobe (Breckenridge et al., 1989; Smyers and Breckenridge, 2003) could resolve whether some of those deposits are older. By contrast, it is likely that there were earlier versions of glacial Lake Columbia, judging from the pre-last-glacial incision and southward shift of the Columbia River through the northernmost extent of the Columbia River Basalt Group (Pardee, 1918, p. 46; Flint, 1935, p. 175–176; Hibbert, 1985, p. 94), most plausibly by earlier glacial blockage of an earlier river route skirting the northern basalt margin.

4.3. Witnesses?

"Were there witnesses?" This common question is addressed by Waitt (2016, p. 442). The question gained notice and was seemingly answered by a stone knife extracted by Luther Cressman in 1953 from a Missoula flood gravel deposit in the Columbia River Gorge (Cressman, 1960, p. 63–66; 1977, p. 50–51, 71). But recent reevaluation suggests that the knife is not actually a knife but rather a "fortuitously flood-bruised basalt pebble" Waitt (2016, p. 442). The knife discounted, no compelling physical evidence is yet reported for humans, their tools, or their habitations within or under ice-age megaflood deposits in the Pacific Northwest.

But the timing of last-glacial floods hints of possible megaflood viewing by early North American inhabitants. Davis et al. (2019) document human occupation of the Coopers Ferry site (Fig. 5) in the Snake River basin by 16.56–15.28 ka (two-sigma uncertainty derived from Bayesian age model of 25 radiocarbon ages). This age post-dates the 18.3–18 ka Bonneville flood down the Snake River, but it very plausibly precedes the last of the Missoula floods, of which ~30 were after the ~16 ka Set S tephra fall. The site itself, at ~420 m elevation, was probably just out of reach from Missoula flood backflooding up the Snake River, where the highest Missoula floods left pebbly silts and ice-rafted erratics up to ~395 m near Lewiston. It is even more likely that that humans were in the region and possibly witnessed the late-glacial floods coming down the Columbia valley after the last of the Missoula floods.

5. Columbia basin megafloods and "Wings of the Morning"

> The region is unique: let the observer take wings of the morning to the uttermost parts of the Earth: he will nowhere find its likeness
> [Bretz, 1928b, p. 446]

The Columbia River basin is where megafloods controversially launched, led by J Harlen Bretz's outrageous hypothesis. Yet other megafloods scarred the region, such as the Bonneville flood spilling out of the Great Basin and down the Snake River, more quietly discovered earlier by G.K. Gilbert. Nearly 100 years from Bretz's first paper outlining and defining the Channeled Scabland, the spectacular flood landscapes of the Columbia River basin still inspire. The region is accessible, well-studied, of huge popular interest, and attracts new tools, techniques, and hypotheses—some outrageous. Hence, the unique landscape continues to breed controversy, motivating even more research into cataclysmic floods and flood processes.

Bretz's relentless advocacy for the singularity of the Channeled Scabland ultimately convinced the scientific community of a great-flood genesis. Yet if the "wings of the morning" had taken him higher in 1928, he would have seen that the region is not unique. Soaring only slightly higher, he would have seen a likeness in the scabland left by the Bonneville flood ~600 km to the southeast, carved by the débâcle described the previous century by G.K. Gilbert. At an orbital level, looking down at the spinning Earth, he would see cataclysmic flood features across terrestrial and submarine landscapes worldwide (e.g. Baker, 2009, p. 405, 2013, 2020). Looking up with eagle eyes, he may have spotted the flood channels scarring other planets (e.g. Baker, 1982; Burr et al., 2009). Yet much of the understanding of these far-flung flood landscapes can be ultimately traced to what was first seen on the ground by Bretz, Gilbert, Malde, and others in the megaflood terrain of the Columbia River basin.

Acknowledgements

Discussions and reviews by Bruce Bjornstad, Brian Atwater, Juergen Herget, and Isaac Larsen have improved this report. Moreover, our work has been aided by discussion, assistance, and support by countless colleagues over the years. We thank Paul Carling, guest editor for this special megaflood volume, for the invitation to contribute and for editorial oversight.

Appendix

Table 1
Columbia Basin megaflood-related lakes and transient impoundment stages; depths, areas, and volumes.

Lake (and surface elevation, m)[a]	Formative mechanism	Calculation notes	Maximum Depth (m)	Area (km^2)	Volume (km^3)	Controls and timing[b]
Pluvial Lake Bonneville						
1552-m Bonneville level	Net hydrologic surplus within closed tectonic basin.	Elevation of Bonneville shoreline at outlet from Currey (1982); area and volume from Adams and Bills (2016, p. 159). Maximum depth estimated by adding 108 m to Provo level maximum depth given by Miller (2016, p. 131).	328	52,110	10,420	Outlet at Marsh Creek divide, Idaho; reached Bonneville level after 18.4 ka, dropped to Provo level by ~18 ka (Oviatt, 2015; Miller, 2016, p. 135–137).
1444-m Provo level	Post-Bonneville-flood overflow of tectonic basin.	Elevation of Provo shoreline at outlet; volume, and measurements from Adams and Bills (2016, p. 159); maximum depth given by Miller (2016, p. 131). Miller et al. (2013) suggest initial Provo shoreline after flood stabilized at 120–125 m below Bonneville level, thus as low as 1427 m.	220	38,150	5,290	Outlet at Red Rock Pass, Idaho; stable from 18.3 to 18.0 Bonneville flood until ~14.8 ka (Oviatt, 2015; Miller, 2016, p.137).
1280-m modern Great Salt Lake	Endorheic lake	Historic average elevation (Arnow and Stephens, 1990, p. 1); volume from fig. 19, p. 18.	10	4,400	18	Closed basin conditions; modern.
Glacial Lake Missoula						
1295-m maximum glacial Lake Missoula	Ice blockage of Clark Fork Purcell Trench lobe in area of Lake Pend Oreille.	Maximum 1295-m shorelines evident on Mount Jumbo, Montana (Table 3); area and volume from GIS assessment of 30-m resolution digital elevation model adjusted for glacial extents as depicted on Fig. 1; maximum depth estimated on basis of modern elevation of Clark Fork River at ice terminus; volume includes part of Kootenay River basin that would not fully drain in a Missoula flood; volume does not account for post-lake erosion of lacustrine and glaciofluvial sediment.	650	10,700	2,540	Ice-dam, no alternative outlet below 1440-m Lookout Pass; lake extant for 3–4 ky during 20–14 ka; gone by 13.7–13.4 ka.
Glacial Lake Spokane						
600-m glacial Lake Spokane	Ice blockage of Spokane River valley by Columbia ice lobe at Columbia-Spokane confluence.	Multiple possible levels between 550 and 671 m identified by Kiver and Stradling (1995, p. 137–139); ~600 m indicated by Latah Creek section (Gaylord et al., 2016, p. 18–26); includes 0.3 km^3 volume of modern Lake Spokane determined by Welch et al. (2015, p. 158); maximum depth estimated from historic river grade at Columbia-Spokane River confluence; volume does not account for post-lake erosion of lacustrine and glaciofluvial sediment.	270	850	48	Ice-dam, no alternative outlet below 703 m; timing uncertain but apparently formed after the ~18.5 ka initiation of glacial Lake Columbia and ended before the ~15.5 ka demise of glacial Lake Columbia (see text).
Glacial Lake Columbia						
750 m flood-swollen glacial Lake Columbia	Maximum transient level of glacial Lake Columbia from Missoula flood inflow, marked by ice-rafted erratics in the Sanpoil River valley (Atwater, 1986, p. 5–7)	Area and volume from GIS assessment of 30-m resolution digital elevation model adjusted for maximum glacial extents approximately shown in Fig. 7d; Lake Spokane volume added from Welch et al. (2015); volume does not account for post-lake erosion of lacustrine and glaciofluvial sediment. Depths estimated for 750-m, 730-m, and 715-m levels on basis of historic Columbia River level of about 270 m near Grand Coulee (River Mile 596)	480	6,140	590	Ice-dammed lake hydrodynamically raised by Missoula floods; multiple spill points south at this stage; possibly only attained this stage when Okanogan lobe was near its maximum extent at ~16 ka.
730-m glacial Lake Columbia	Ice blockage of Columbia River by Okanogan lobe; local and poorly formed shorelines at 725–730 m marking transient inundation by "Lake Columbia I" (Waitt and Thorson, 1983, p. 57; Kiver and Stradling, 1995, p. 59, 141)		460	5,100	475	Short-lived(?) ice-dammed lake, possibly outleting by way of Cheney-Palouse or Telford Crab Creek scabland tracts; Cr hydrodynamically raised by Missoula floods; multiple spill points south at this stage; possibly only attained this stage when Okanogan lobe was near its maximum extent at ~16 ka.
715 m-glacial Lake Columbia	Ice blockage of Columbia River by Okanogan lobe; 715 m shoreline gravel (Atwater, 1986, p. 6) marking transient high-stand		445	4,330	405	
653-m glacial Lake Columbia	Ice blockage of Columbia River; possible stable but likely short-lived drainage at ~653 m, corresponding to outlet divide into Moses Coulee	Area and volume from GIS assessment of 30-m resolution digital elevation model adjusted for maximum glacial extents approximately shown in Fig. 7c; Lake Spokane volume added from Welch et al. (2015); volume does not account for post-lake erosion of lacustrine and	423	3,380	340	Possible short-lived ice-dammed lake outletting by way of Moses Coulee; such a lake could only exist if Grand Coulee divide higher than 653 m and its most likely timing would be between first blockage of the Columbia valley at ~18.5–18 ka and prior to maximum

(continued on next page)

Table 1 (*continued*)

Lake (and surface elevation, m)[a]	Formative mechanism	Calculation notes	Maximum Depth (m)	Area (km^2)	Volume (km^3)	Controls and timing[b]
		glaciofluvial sediment. Does not include volume of Banks Lake. Depths estimated for 470-m, 520-m, and 653-m levels on basis of historic Columbia River level of 230 m at Okanogan River confluence (River Mile 535)				ice-lobe extent at ~16 ka.
520-m glacial Lake Columbia	Ice blockage of Columbia River by Okanogan lobe; stable drainage at 520 m, corresponding to prominent terrace correlated to "Lake Columbia II" (Kiver and Stradling, 1995, pg. 139–140)		290	930	95	Persistent ice-dammed lake outletting through Grand Coulee; different elevations possibly the result of outlet erosion; possible extant at these levels for as long as 3 ky between 18.5 and 15.5 ka, including during the ~16 ka Mount St. Helens Set S tephra fall.
470 m glacial Lake Columbia	Ice blockage of Columbia River by Okanogan lobe near Okanogan River confluence; stable drainage over 470-m modern Grand Coulee rock threshold at south end of Banks Lake; scenario likely associated with glacial Lake Columbia outburst		240	595	54	
390-m Franklin D. roosevelt Lake Roosevelt	Impounded in 1941 by Grand Coulee Dam	Lake dimensions from Martin and Hanson (1966, p. 103)	120	320	12	Constructed dam completed in 1942.
Hydrodynamic Lake Lewis (upstream of Wallula Gap constriction)						
366-m Lake Lewis	Approximate maximum Pasco Basin flood stage marked by highest Pasco Basin erratics (Bjornstad, 2014, p. 54)	Area and volume from GIS assessment of 30-m resolution digital elevation model adjusted for glacial extent approximately shown on Fig. 1; volume does not account for post-lake erosion of lacustrine and glaciofluvial sediment; depth measured from historical 90-m river level at Wallula Gap.	276	13,680	1,410	Transient inundation levels during Missoula floods created by hydraulic ponding by downstream constrictions at Wallula Gap and in the Columbia River Gorge; all levels likely attained during last-glacial period of flooding during 20–14 ka.
360-m Lake Lewis	Approximate maximum flow stage at Wallula Gap (O'Connor and Baker, 1992, site 17)		270	13,100	1,330	
338-m Lake Lewis	Approximate maximum flow stage predicted by hydrodynamic flow modeling (Fig. 14d)		248	11,020	1,065	

Notes:

[a] All elevations are relative to the National Geodetic Vertical Datum of 1988 and are modern positions. Many Pleistocene landform elevations were affected by isostatic deformation owing to loading by water and ice.

[b] Timing information summarized from material in text.

Table 2
Columbia Basin megaflood discharge estimates.

Location	Reference	Discharge	Comments
Bonneville flood			
Red Rock Pass Outlet Area	Gilbert (1890, p. 177)	"Flood volume of the Missouri"	Estimate for Portneuf valley, 40 km downstream of Red Rock Pass; equivalent to about $0.02{\cdot}10^6$ m^3/s (O'Connor, 2016, p. 118).
	O'Connor (1993, p. 16–19); O'Connor (2016, p. 119)	$0.85–1{\cdot}10^6$ m^3/s; $1.5–1.6{\cdot}10^6$ m^3/s	Low range for step-backwater computations fit to highwater evidence for 90 km reach between Red Rock Pass outlet of Lake Bonneville and Snake River Plain near Pocatello; high range maximum possible critical-flow estimate for Red Rock Pass of $1.1–1.6{\cdot}10^6$ m^3/s for maximum Lake Bonneville elevation of 1552 m and a total lake-level drop of 108–125 m.
	Abril-Hernández et al. (2018)	$0.85{\cdot}10^6$ m^3/s	Two-dimensional hydrodynamic model coupling outlet erosion and drainage of Lake Bonneville.
Downstream Estimates	Stearns (1962)	$0.3{\cdot}10^6$ m^3/s	At head of Hells Canyon (Snake River Mile 285); probably by Mannings equation, as inferred by Malde (1968, p. 12).
	Malde (1968, p. 12, 25)	$0.35–0.46{\cdot}10^6$ m^3/s	Critical depth calculation based on highwater evidence at Swan Falls constriction (Snake River Mile 459).
	Jarrett and Malde (1987)	$0.79–1.0{\cdot}10^6$ m^3/s	Step-backwater computations fit to new highwater evidence at Swan Falls constriction.
	O'Connor (1993, p. 12–40)	$0.6–1.0{\cdot}10^6$ m^3/s	Ten separate analyses between Pocatello and Lewiston, ID; all step-backwater computations fit to highwater evidence; includes $0.82–0.91{\cdot}10^6$ m^3/s estimate for Swan Falls constriction.
	Lapôtre et al. (2016, p. 1249, 1252)	$0.010–0.011{\cdot}10^6$ m/s	Canyon formation model applied to two amphitheater-headed canyons conveying a portion of the total Bonneville flood volume; combined estimate.
Missoula Floods			
Lake Missoula Outlet area			
Eddy Narrows (Clark Fork River Mile 215; internal Lake Missoula constriction regulating outflow; ~75% of lake volume upstream)	Pardee (1942, p. 1596–1597)	$10.9{\cdot}10^6$ m^3/s	Manning and Chezy formulas, using cross-sectional area associated flow with 305 m deep and regional modern river gradient
	This report	$29{\cdot}10^6$ m^3/s	Critical flow calculation for parabolic channel (as formulated by Benito and O'Connor, 2013, p. 466), assuming 1295 lake level 563 m above channel floor, and cross-sectional top width of 2250 m; yields maximum possible outflow rate.

Table 2 (*continued*)

Location	Reference	Discharge	Comments
Ice dam (assuming subglacial, jökulhlaup-style water release)	Clarke et al. (1984)	2.6–3.5·10^6 m^3/s; 13–15.3·10^6 m^3/s	Model simulating subglacial release of Lake Missoula from near maximum 1265 m lake level; low range is for conditions of downstream backwater (hydraulic ponding or glacial Lake Columbia) and delayed conveyance of flow to ice dam owing to complex lake geometry; high range is for conditions of downstream backwater and immediate conveyance of lake volume to release point.
	Beget (1986)	2.1·10^6 m^3/s	Extrapolated empirical relation between ice-dammed lake volume and associated jökulhlaup peak discharges; for Lake Missoula volume of 2180 km^3.
Rathdrum Prairie - Spokane Valley (conveying most, but not all, flow exiting from ice dam area)	Baker (1973, p. 14–22)	21.3·10^6 m^3/s	Slope-area method based on water-surface slope defined by high-water evidence
	O'Connor and Baker (1992, p. 272–276)	14–20·10^6 m^3/s; 27·10^6 m^3/s	Step-backwater computations fit to high-water evidence; low range is minimum flow required to emplace maximum stage evidence for present topography; high value is for valley bottom 60 m lower, consistent with bottom of marginal lake bottoms.
	Denlinger and O'Connell (2010, p. 685)	35·10^6 m^3/s	Instantaneous peak discharge entering "head of Rathdrum valley" on basis of present topography and two-dimensional hydrodynamic flow model of coupled lake emptying and downstream flooding for 1250 m maximum Lake Missoula level and instantaneous dam removal.
Wallula Gap and Columbia River Gorge			
Wallula Gap (conveying all flow downstream from Pasco Basin)	Bretz (1925b, p. 257–258)	1.9·10^6 m^3/s	Chezy equation calculation by D.F. Higgins, assuming stage of 312 m within the constriction and a water-surface slope defined by local highwater evidence.
	Baker (1973, p. 14–22)	9.1·10^6 m^3/s	Contracted opening equation assuming 350 m ponding level in Pasco Basin
	Craig and Hanson (1985, p. 52–53)	12.5·10^6 m^3/s	Step-backwater computation assuming critical flow through constriction and maximum Pasco Basin ponding level of 350 m
	O'Connor and Baker (1992, p. 269–272)	7.5–12·10^6 m^3/s; 14·10^6 m^3/s	Step-backwater computation assuming maximum Pasco Basin ponding level of 370–385 m; low range incorporates effects of downstream ponding behind Columbia River Gorge constrictions; 14·10^6 m^3/s value is for critical flow and a Pasco Basin ponding level of 375 m, representing an upper limit for discharge through the present constriction geometry.
	Denlinger and O'Connell (2010, p. 685–687)	6.2·10^6 m^3/s	Two-dimensional hydrodynamic finite-volume flow modeling of entire flood system, assuming modern topography, 1250-m level of Lake Missoula, and instantaneous ice-dam removal; peak flow through Wallula Gap associated with maximum Pasco Basin ponding level of 336 m.
Columbia River Gorge (conveying all flow)	Benito and O'Connor (2003)	10·10^6 m^3/s	Step-backwater computations fit to Columbia Valley high-water evidence for 190 km between Columbia River Mile 240 and River Mile 120 (eastern Portland Basin), encompassing critical flow constrictions within the Columbia River Gorge.
Individual Flood Tracts Conveying Partial Flood Discharges			
Columbia Valley; Grand Coulee to Sentinel Gap	Craig and Hanson (1985, p. 53–66)	9.5·10^6 m^3/s	Step-backwater computation assuming assuming Pasco Basin ponding level of 350 m and 366-m maximum stage upstream of Sentinel Gap
	Harpel (1996); Harpel et al. (2000); Waitt (2017, p. 192; Waitt et al., 2019, 332–333)	10·10^6 m^3/s; 13·10^6 m^3/s	Step-backwater computations fit to Columbia Valley high-water evidence between Sentinel Gap and Grand Coulee; low value is Harpel's original determination; high value is updated calculation by Austin Rains reported by Waitt (2017); primarily fit to Wenatchee area erratics (Waitt et al., 2019).
Moses Coulee	Hanson (1970, p. 59)	1.1–9.8·10^6 m^3/s	Manning equation calculations; range incorporates estimated maximum and minimum cross-sectional areas for two different cross sections
	Harpel (1996, p. 10); Harpel et al. (2000)	11·10^6 m^3/s	Step-backwater computations fit to highwater evidence for lower 30 km of Moses Coulee, assuming present topography and 366-m ponding level at downstream end of analysis reach.
	Larsen and Lamb (2016)	$\leq$0.6·10^6 m^3/s	Threshold shear stress model predictions of flow required for local, incremental, bedrock incision of Rattlesnake Springs scabland.
Grand Coulee - Quincy Basin	Baker (1973, p. 14–22)	4.5·10^6 m^3/s	Contracted opening equation applied to lower Grand Coulee on basis of high-water evidence
	Atwater (1987, p. 192)	$\geq$0.13·10^6 m^3/s	Local velocity estimate on basis of grain size and bedforms in flood beds deposited in glacial Lake Columbia sediment near Steamboat Rock within upper Grand Coulee; extrapolated for entire Grand Coulee cross section with assumed maximum flood stage of 493 m; applicable to final (probably small) Missoula floods.
	Harpel (1996, p. 5–9); Harpel et al. (2000); Waitt et al. (2000)	12–14·10^6 m^3/s	Step-backwater computations fit to highwater evidence for upper 50 km of Grand Coulee, assuming present topography and water surface descending from 750 m at the coulee head to 530 m at downstream end of upper Grand Coulee.
	Lapôtre et al. (2016, p. 1249, 1252)	0.10–0.65·10^6 m^3/s	Canyon formation model applied to two amphitheater-headed canyons forming part of the Dry Falls cataract complex within Grand Coulee; combined estimates for both of the assessed

(continued on next page)

Table 2 (*continued*)

Location	Reference	Discharge	Comments
			cataracts; accounts for a portion of the total Grand Coulee flow.
	Lapôtre et al. (2016, p. 1249, 1252)	$0.21–1.91 \cdot 10^6$ m^3/s	Canyon formation model applied to the pairs of amphitheater-headed canyons (cataracts) forming Potholes Coulee and Frenchman Coulee for a portion of the flow exiting Quincy Basin; combined estimates for the four individually assessed amphitheater-headed canyons; accounts for portions of two of the four major flow paths out flow exiting Quincy Basin.
Telford Crab Creek Scabland Tract	Baker (1973, p. 14–22)	$6.1 \cdot 10^6$ m^3/s	Slope-area method based on water-surface slope defined by high-water evidence; includes Crab Creek flow in part derived from Cheney Palouse tract.
Cheney Palouse Scabland Tract	Baker (1973, p. 14–22)	$1–12 \cdot 10^6$ m^3/s	Multiple slope-area estimates based on maximum stage evidence for individual channels; $12 \cdot 10^6$ m^3/s value based on combined estimates for the two major channel complexes entering head of Washtucna Coulee.
	Harpel (1996, p. 10–14); Harpel et al. (2000); Waitt et al. (2000)	$5.5–7 \cdot 10^6$ m^3/s	Step-backwater computations fit to highwater evidence for 51-km reach of Snake River between River Mile 20 and 52, encompassing canyon reach downstream from major Cheney-Palouse tract flow input by way of Palouse River canyon and adjacent scabland tracts breaching the divide between Washtucna Coulee and the Snake River Canyon.
	Harpel (1996, p. 14); Harpel et al. (2000); Waitt et al. (2000)	$2 \cdot 10^6$ m^3/s	Step-backwater computations fit to highwater evidence for 45-km reach of Washtucna Coulee between Washtucna and Connell; together with Snake River canyon estimate, implies $\geq 7.5–9 \cdot 10^6$ m^3/s for Cheney-Palouse tract flow.
Glacial Lake Columbia Outburst Flood			
Ice Dam	This report	$0.13 \cdot 10^6$ m^3/s	Based on regression relating peak discharge to lake volume for "non-tunnel drainage" ice-dam failures (Walder and Costa, 1996, p. 708, equation 4) and a glacial Lake Columbia volume of $5.4 \cdot 10^{10}$ m^3 associated with its final, 470 m stabilized level. This volume, based on modern topography, probably overestimates flood volume because of significant post-glacial incision of the glacial Lake Columbia sediment within the Columbia Valley.

Table 3
Evidence of maximum Missoula flood stages, selected from multiple sources and new observations (as noted).

General location	Feature	Missoula flood context	Minimum constraint	Maximum constraint	Latitude	Longitude	Location source	Elevation (ft)	Elevation (m)	Elevation source
Glacial Lake Missoula										
Missoula Valley; Mount Jumbo	Shoreline	Maximum level glacial Lake Missoula	X		46.87154	-113.96558	Larry Smith, Dec. 6, 2017, written communication	4250	1295	USGS quad, 40 ft contour interval
Missoula Valley; Mount Sentinel	Shoreline	Maximum level glacial Lake Missoula	X		46.84543	-113.97624	Charles Cannon interpretation (Dec. 7, 2017, written communication)	4250	1295	5-ft resolution raster derived from 2002 lidar
Bitterroot Valley; east of Carlton	Shoreline	Maximum level glacial Lake Missoula	X		46.68070	-114.00100	Charles Cannon interpretation (Dec. 7, 2017, written communication)	4240	1292	USGS quad, 20 ft contour interval
Rathdrum Prairie - Spokane Valley										
Scotia Channel	Divide crossing	Flow spilling north of Rathdrum Prairie through Scotia Channel	X		48.08520	-117.24180	Waitt et al. (2016)	2680	817	USGS quad, 40 ft contour interval
Rathdrum Prairie	Eddy bar	Flow in Rathdrum Prairie	X		47.90917	-116.61874	Modified from O'Connor and Baker (1992, p. 274, site 20)	2660	811	USGS quad, 20 ft contour interval
Spirit Lake, Rathdrum Prairie	Tractive gravel bar	Flow in Rathdrum Prairie	X		47.96160	-116.83160	O'Connor interpretation	2640	805	USGS quad, 40 ft contour interval
Coeur d'Alene Lake	Erratics	Spill from Rathdrum Prairie	X		47.48053	-116.86952	Richmond et al. (1965, p. 239); Baker (1973, p. 67)	2665	813	As reported; location representative
Coeur d'Alene Lake	Erratics	Spill from Rathdrum Prairie	X		47.48053	-116.86952	Dort (1960); as reported by Smyers and Breckenridge (2003, p. 6)	2600	792	As reported; location representative
Coeur d'Alene Lake spillover	Divide crossing	Spill from Rathdrum Prairie	X		47.45156	-117.00237	O'Connor interpretation	2580	786	USGS quad, 20 ft contour interval

Table 3 (*continued*)

General location	Feature	Missoula flood context	Minimum constraint	Maximum constraint	Latitude	Longitude	Location source	Elevation (ft)	Elevation (m)	Elevation source
Setters Divide, Coeur d'Alene Lake spillover	Divide crossing	Spill from Rathdrum Prairie	X		47.47368	-117.03221	Modified from Richmond et al. (1965, p. 239)	2600	792	USGS quad, 20 ft contour interval
Spokane Valley	Divide crossing	Flow in Spokane Valley	X		47.64530	-117.04960	Baker (1973, appendix 1, high-water mark 64); O'Connor and Baker (1992, site 8)	2560	780	USGS quad, 20 ft contour interval
Spokane-Cheney-Palouse divide	Erratic	Flow in Spokane Valley	X		47.60204	-117.34469	Baker (1973, appendix 1, high-water mark 60; location reported as S2 T24N 43E); Bretz (1923a, p. 581)	2525	770	As reported
Near Reardon, south of Spokane Valley	Divide crossing	Flow in Spokane Valley, near Cheney Palouse divide	X		47.68710	-117.82353	O'Connor interpretation	2500	762	USGS quad, 20 ft contour interval
Near Reardon, south of Spokane Valley	Erratic	Flow in Spokane Valley, near Cheney Palouse divide	X		47.70215	-117.83102	O'Connor, unpublished, field site 09-17-2019-03	2550	777	USGS quad, 20 ft contour interval
Near Reardon, south of Spokane Valley	Divide crossing	Flow in Spokane Valley, spilling south into Cheney-Palouse tract	X		47.68746	-117.83147	O'Connor interpretation	2520	768	USGS quad, 20 ft contour interval
Near Reardon, south of Spokane Valley	Divide crossing	Flow in Spokane Valley, spilling south into Cheney-Palouse tract	X		47.70774	-117.84262	O'Connor interpretation	2560	780	USGS quad, 20 ft contour interval
Near Reardon, south of Spokane Valley	Divide crossing	Flow in Spokane Valley, spilling south into Cheney-Palouse tract	X		47.67189	-117.99988	O'Connor interpretation	2520	768	USGS quad, 20 ft contour interval
Near Davenport, south of Spokane Valley	Divide not crossed	Flow in Spokane Valley, near Telford Crab Creek divide		X	47.72712	-118.06360	O'Connor interpretation	2530	771	USGS quad, 10 ft contour interval
Upper Columbia River Valley; Spokane River confluence to Grand Coulee										
Columbia Valley near Hawk Creek	Erratic	Flow in Columbia Valley, near Telford Crab Creek divide	X		47.83279	-118.30521	O'Connor, unpublished, field site 05-19-2019-04	2500	762	USGS quad, 40 ft contour interval
Columbia Valley near Hawk Creek	Erratic	Flow in Columbia Valley, near Telford Crab Creek divide	X		47.83315	-118.30567	O'Connor, unpublished, field site 05-19-2019-05	2520	768	USGS quad, 40 ft contour interval
Columbia Valley near Hawk Creek	Erratic	Flow in Columbia Valley, near Telford Crab Creek divide	X		47.78741	-118.37063	O'Connor, unpublished, field site 05-19-2019-14	2500	762	USGS quad, 40 ft contour interval
Columbia Valley near Hawk Creek	Erratic	Flow in Columbia Valley, near Telford Crab Creek divide	X		47.80833	-118.48868	Thomas Cooney, Oct. 17, 2018, written communication	2480	756	USGS quad, 40 ft contour interval
South of Columbia Valley near Creston	Divide crossing	Flow spilling south of Columbia Valley into Telford Crab Creek tract	X		47.80932	-118.50511	O'Connor interpretation	2497	761	USGS quad, checked elevation
South of Columbia Valley near Creston	Divide crossing	Flow spilling south of Columbia Valley into Telford Crab Creek tract	X		47.80190	-118.50730	O'Connor interpretation	2469	753	USGS quad, checked elevation
South of Columbia Valley near Creston	Divide not crossed	Flow in Columbia Valley, near Telford Crab Creek divide		X	47.81468	-118.51147	O'Connor interpretation	2540	774	USGS quad, 20 ft contour interval
Sanpoil River valley backflooding	Many erratics	Flow in Columbia Valley, backflooding Sanpoil Valley	X		47.96000	-118.69000	Atwater (1986, p. 7, plate 2A)		750	As reported, location representative
Columbia Valley at Grand Coulee	Divide not crossed	Head of Grand Coulee		X	47.90540	-118.93730	O'Connor interpretation	2600	792	USGS quad, 40 ft contour interval
Columbia Valley at Grand Coulee	Stripped basalt	Head of Grand Coulee	X		47.90330	-118.95160	O'Connor interpretation	2480	756	USGS quad, 40 ft contour interval
Columbia Valley; Chelan to Pasco Basin										
Entiat area	Trim line	Flow in Columbia Valley, Chelan area	X		47.75280	-120.12500	O'Connor interpretation	1560	475	USGS quad, 40 ft contour interval
Entiat area	Eddy bar	Flow in Columbia Valley, Chelan area	X		47.77480	-120.19530	O'Connor interpretation	1440	439	USGS quad, 40 ft contour interval

(continued on next page)

Table 3 (*continued*)

General location	Feature	Missoula flood context	Minimum constraint	Maximum constraint	Latitude	Longitude	Location source	Elevation (ft)	Elevation (m)	Elevation source
Wenatchee area	Erratic	Flow in Columbia Valley, Wenatchee area	X		47.49558	-120.33101	O'Connor, unpublished, field site 09-06-2017-01	1560	475	USGS quad, 20 ft contour interval
Wenatchee area	Erratic	Flow in Columbia Valley, Wenatchee area	X		47.49549	-120.33125	O'Connor, unpublished, field site 09-14-2015-09	1550	472	USGS quad, 20 ft contour interval
Wenatchee area	Erratic	Flow in Columbia Valley, Wenatchee area	X		47.44346	-120.26008	Waitt et al. (2019; p. 324, L12)		483	As reported
Wenatchee area	Erratic	Flow in Columbia Valley, Wenatchee area	X		47.43614	-120.24321	Waitt et al. (2019; p. 324, L11)		476	As reported
Wenatchee area	Erratic	Flow in Columbia Valley, Wenatchee area	X		47.42241	-120.20271	Waitt et al. (2019; p. 324, L10)		495	As reported
Wenatchee area	Erratic	Flow in Columbia Valley, Wenatchee area	X		47.41527	-120.19279	Waitt et al. (2019; p. 324, L8)		488	As reported
Wenatchee area	Erratic	Flow in Columbia Valley, Wenatchee area	X		47.41271	-120.17858	Waitt et al. (2019; p. 324, L7)		490	As reported
Wenatchee area	Erratic	Flow in Columbia Valley, Wenatchee area	X		47.40155	-120.13408	Waitt et al. (2019; p. 324, W2)		481	As reported
Lynch Coulee erratic	Erratic	Backflooding from Columbia Valley or possibly Quincy Basin outflow	X		47.29992	-119.95224	Bruce Bjornstad, Aug. 29, 2017, written communication	1420	433	USGS quad, 20 ft contour interval
Babcock Ridge divide crossing	Divide crossing	Flow in Columbia Valley, Quincy Basin overflow	X		47.21190	-119.96744	O'Connor interpretation	1420	433	USGS quad, 20 ft contour interval
Babcock Bench eddy bar	Eddy bar	Flow in Columbia Valley	X		47.21370	-119.97820	O'Connor interpretation	1358	414	USGS quad checked elevation
Babcock Bench	Erratic	Flow in Columbia Valley	X		47.17011	-119.97455	Ken Lacy, Jan. 7, 2020 email (erratic 2)	1340	408	USGS quad, 10 ft contour interval
Babcock Bench	Erratic	Flow in Columbia Valley	X		47.17092	-119.97488	Ken Lacy, Jan. 7, 2020 email (erratic 1)	1360	415	USGS quad, 10 ft contour interval
Babcock Bench	Erratic	Flow in Columbia Valley	X		47.16904	-119.97662	O'Connor, unpublished, field site 08-03-2017-02	1280	390	USGS quad, 20 ft contour interval
Babcock Bench	Stripped basalt	Flow in Columbia Valley	X		47.16632	-119.97801	O'Connor, unpublished, field site 08-03-2017-01	1370	418	USGS quad, 20 ft contour interval
Ginkgo State Park	Erratic	Flow in Columbia Valley, upstream of Sentinel Gap	X		46.94480	-120.02631	Andrea Balbas, Aug. 5, 2017, written communication	1200	366	USGS quad, 20 ft contour interval
Sentinel Gap	Erratic	Flow in Columbia Valley, upstream of Sentinel Gap	X		46.90410	-120.03480	Dunbar (1998, fig. 7)	1210	369	USGS quad, 20 ft contour interval
Sentinel Gap	Divide not crossed	Flow in Columbia Valley, Sentinel Gap		X	46.81099	-119.94754	O'Connor interpretation	1560	475	USGS quad, 20 ft contour interval
Rattlesnake Mountain	Erratic	Pasco Basin ponding	X		46.39701	-119.53631	Bjornstad (2014; and written communication)	1190	363	USGS quad, 40 ft contour interval
Red Mountain	Erratic	Pasco Basin ponding	X		46.30100	-119.44800	Bjornstad (2014, p. 54)		366	As reported by Bjornstad (2014)
Walla Walla River backflooding	Upper limits of erratics, pebbly silts	Pasco Basin ponding	X		46.07040	-118.82950	Bretz (1929b, p. 516–517)	1100	335	Uncertain; location representative
Columbia Valley; Wallula Gap to Pacific Ocean										
Wallula Gap	Top of loess scarp	Flow through Wallula Gap, Pasco Basin ponding level at outlet		X	46.03890	-118.90371	O'Connor and Baker (1992, site 18)	1240	378	USGS quad, 20 ft contour interval
Wallula Gap	Base of loess scarp, near boulder bar	Flow through Wallula Gap, Pasco Basin ponding level at outlet	X		46.03991	-118.90386	O'Connor interpretation	1180	360	USGS quad, 20 ft contour interval

Table 3 (*continued*)

General location	Feature	Missoula flood context	Minimum constraint	Maximum constraint	Latitude	Longitude	Location source	Elevation (ft)	Elevation (m)	Elevation source
Wallula Gap	Divide not crossed	Flow through Wallula Gap, Pasco Basin ponding level at outlet		X	46.03362	-118.91791	O'Connor interpretation	1220	372	USGS quad, 20 ft contour interval
Wallula Gap	Divide crossing	Flow through Wallula Gap, Pasco Basin ponding level at outlet	X		46.05252	-118.99107	Bretz (1969); O'Connor and Baker (1992, site 19)	1140	347	USGS quad, 20 ft contour interval
Wallula Gap	Unflooded loess; top of loess scarp	Flow through Wallula Gap, Pasco Basin ponding level at outlet		X	46.05195	-118.99401	O'Connor interpretation	1223	373	Surveyed section corner
Columbia Valley east of John Day River	Erosional trim line	Flow in Columbia River valley, Columbia River Gorge	X		45.68806	-120.39167	Benito and O'Connor (2003, site 1)	1120	341	USGS quad, 20 ft contour interval
The Nook at John Day River	Divide not crossed	Flow in Columbia River valley, Columbia River Gorge		X	45.68660	-120.50080	Benito and O'Connor (2003, site 5)	1140	347	USGS quad, 20 ft contour interval
John Day River confluence	Divide not crossed	Flow in Columbia River valley, Columbia River Gorge		X	45.73470	-120.62000	Benito and O'Connor (2003, site 8)	1180	360	USGS quad, 20 ft contour interval
Deschutes River confluence	Erosional trim line	Flow in Columbia River valley, Columbia River Gorge	X		45.63100	-120.88460	Benito and O'Connor (2003, site 13)	1000	305	USGS quad, 40 ft contour interval
Deschutes River confluence	Divide not crossed	Flow in Columbia River valley, Columbia River Gorge		X	45.66450	-120.91840	Benito and O'Connor (2003, site 14)	1120	341	USGS quad, 40 ft contour interval
Fifteenmile Creek	Erratics	Flow in Columbia River valley, Columbia River Gorge	X		45.61600	-120.94913	O'Connor and Cannon, unpublished, field site 04-26-2016-10	1120	341	USGS quad, 40 ft contour interval
Fulton Ridge east of Fairbanks Gap	Divide not crossed	Flow in Columbia River valley, Columbia River Gorge		X	45.63590	-120.99180	O'Connor interpretation	1200	366	USGS quad, 40 ft contour interval
Dalles Basin trimline	Base of erosional trim line	Flow in Columbia River valley, Columbia River Gorge	X		45.64875	-121.24395	Benito and O'Connor (2003, site 20)	1040	317	USGS quad, 40 ft contour interval
Tom McCall Point	Base of erosional trim line	Flow in Columbia River valley, Columbia River Gorge	X		45.67642	-121.30160	Benito and O'Connor (2003, site 21)	920	280	USGS quad, 40 ft contour interval
Balch Lake	Divide not crossed	Flow in Columbia River valley, Columbia River Gorge		X	45.72480	-121.30904	Benito and O'Connor (2003, site 22); Allison (1933, p. 709)	1000	305	USGS quad, 40 ft contour interval
Hood River Valley	Erratic	Flow in Columbia River valley, Columbia River Gorge	X		45.65407	-121.50511	Benito and O'Connor (2003, site 27); Bretz (1919); Allison (1933, p. 709); Newcomb (169, plate 1)	840	256	USGS quad, 40 ft contour interval
Underwood Mountain	Prominent trimline	Flow in Columbia River valley, Columbia River Gorge	X		45.73570	-121.55390	O'Connor interpretation	900	274	USGS quad, 40 ft contour interval
Eastern Portland basin	Divide not crossed	Flow in Columbia River valley, Portland basin		X	45.57350	-122.28280	O'Connor interpretation	520	158	USGS quad, 20 ft contour interval
Eastern Portland basin	Prominent trimline	Flow in Columbia River valley, Columbia River Gorge	X		45.57405	-122.29107	O'Connor interpretation	420	128	USGS quad, 20 ft contour interval
Eastern Portland basin	Divide not crossed	Flow in Columbia River valley, Portland basin		X	45.52860	-122.32520	O'Connor interpretation	540	165	USGS quad, 40 ft contour interval
Eastern Portland basin	Divide crossing	Flow in Columbia River valley, Portland basin	X		45.60970	-122.38360	O'Connor interpretation	360	110	USGS quad, 20 ft contour interval

(continued on next page)

Table 3 (*continued*)

General location	Feature	Missoula flood context	Minimum constraint	Maximum constraint	Latitude	Longitude	Location source	Elevation (ft)	Elevation (m)	Elevation source
Portland Basin	Eddy bar	Flow in Columbia River valley, Portland basin	X		45.51908	-122.41630	Evarts and O'Connor (2008)	370	113	USGS quad, 10 ft contour interval
Southern Willamette Valley	Erratic	Flow spilling backflooding Willamette River basin from Columbia valley	X		44.04825	-123.17255	Minervini et al. (2003, erratic 307)	391	119	USGS quad, near checked elevation
Cowlitz River Valley	Erratic	Flow backflooding Cowlitz River valley from Columbia River estuary near Longview, WA	X		46.30526	-122.92697	O'Connor, unpublished, field site 05-24-2006-01	200	61	USGS quad, 20 ft contour interval
Cheney-Scabland tract										
Medical Lake	Eddy deposit	Flow entering Cheney Palouse tract	X		47.55710	-117.64200	O'Connor interpretation	2460	750	USGS quad, 20 ft contour interval
Near Cheney	Divide not crossed	Flow within Cheney Palouse tract		X	47.51457	-117.61260	O'Connor interpretation	2540	774	USGS quad, 20 ft contour interval
Near Cheney	Divide crossing	Flow within Cheney Palouse tract	X		47.50295	-117.59277	O'Connor interpretation	2440	744	USGS quad, 20 ft contour interval
Near Cheney	Divide crossing	Flow within Cheney Palouse tract	X		47.42040	-117.91490	Baker (1973, appendix 1, high-water mark 54)	2320	707	USGS quad, 20 ft contour interval
Near Cheney	Divide crossing	Flow within Cheney Palouse tract	X		47.42630	-117.69150	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 50)	2400	732	USGS quad, 10 ft contour interval
Near Cheney	Divide not crossed	Flow within Cheney Palouse tract		X	47.42540	-117.69720	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 50)	2420	738	USGS quad, 10 ft contour interval
Near Cheney	Divide crossing	Flow within Cheney Palouse tract	X		47.34300	-117.70680	Baker (1973, appendix 1, high-water mark 51)	2290	698	USGS quad, 10 ft contour interval
Upper Rock Creek	Divide crossing	Flow within Cheney Palouse tract	X		47.36110	-117.53060	Suzanna Doak interpretation	2420	738	USGS quad, 20 ft contour interval
Upper Rock Creek	Divide crossing, stripped basalt	Flow within Cheney Palouse tract	X		47.34620	-117.54650	Suzanna Doak interpretation	2380	725	USGS quad, 20 ft contour interval
Alkali Lake	Divide crossing, stripped basalt	Flow within Cheney Palouse tract	X		47.3722	-117.6936	Suzanna Doak interpretation	2330	710	USGS quad, 10 ft contour interval
Alkali Lake	Divide not crossed	Flow within Cheney Palouse tract		X	47.3682	-117.6998	Suzanna Doak interpretation	2380	725	USGS quad, 10 ft contour interval
Hog Lake, Northeast of Sprague	Divide crossing	Flow within Cheney Palouse tract	X		47.3807	-117.8448	O'Connor/Doak interpretation	2270	692	USGS quad, 10 ft contour interval
Hog Lake, Northeast of Sprague	Divide not crossed	Flow within Cheney Palouse tract		X	47.3747	-117.8426	O'Connor/Doak interpretation	2320	707	USGS quad, 10 ft contour interval
Near Sprague	Divide crossing	Flow within Cheney Palouse tract	X		47.34300	-117.94060	Baker (1973, appendix 1, high-water mark 52)	2130	649	USGS quad, 10 ft contour interval
Northeast of Sprague	Divide crossing	Flow within Cheney Palouse tract	X		47.31760	-117.78570	Baker (1973, appendix 1, high-water mark 53)	2220	677	USGS quad, 10 ft contour interval
West of Sprague Lake	Divide crossing	Flow within Cheney Palouse tract			47.20750	-118.20810	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 39)	2010	613	USGS quad, 10 ft contour interval
West of Sprague Lake	Divide crossing	Flow within Cheney Palouse tract	X		47.21660	-118.19140	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 39)	1990	607	USGS quad, 10 ft contour interval
Rock Creek, Rock Lake	Divide not crossed	Flow within Cheney Palouse tract		X	47.16860	-117.67560	Baker (1973, appendix 1, high-water mark 44)	2140	652	USGS quad, 10 ft contour interval

Table 3 (*continued*)

General location	Feature	Missoula flood context	Minimum constraint	Maximum constraint	Latitude	Longitude	Location source	Elevation (ft)	Elevation (m)	Elevation source
Rock Creek, Rock Lake	Divide crossing	Flow within Cheney Palouse tract	X		47.17300	-117.66460	Baker (1973, appendix 1, high-water mark 44)	2100	640	USGS quad, 10 ft contour interval
Rock Creek, Rock Lake	Divide not crossed	Flow within Cheney Palouse tract		X	47.15670	-117.68530	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 45)	2030	619	USGS quad, 10 ft contour interval
Rock Creek, Rock Lake	Divide crossing	Flow within Cheney Palouse tract	X		47.14960	-117.74840	Baker (1973, appendix 1, high-water mark 47)	1940	591	USGS quad, 10 ft contour interval
East of Ritzville, west of Rock Creek	Divide not crossed	Flow within Cheney Palouse tract		X	47.12980	-118.25480	Baker (1973, appendix 1, high-water mark 43)	1870	570	USGS quad, 10 ft contour interval
East of Ritzville, west of Rock Creek	Divide crossing	Flow within Cheney Palouse tract	X		47.11910	-118.26240	Baker (1973, appendix 1, high-water mark 43)	1840	561	USGS quad, 10 ft contour interval
Cottonwood Valley backflooding	Upper limits of erratics, pebbly silts	Flow within Cheney Palouse tract	X		47.11170	-117.70200	Bretz (1929a, p. 400)	1940	591	As reported by Bretz (1929a); location representative
Rock Creek	Divide not crossed	Flow within Cheney Palouse tract		X	47.10030	-117.88490	Baker (1973, appendix 1, high-water mark 36)	1870	570	USGS quad, 10 ft contour interval
Rock Creek	Divide crossing	Flow within Cheney Palouse tract	X		47.09940	-117.88100	Baker (1973, appendix 1, high-water mark 36)	1830	558	USGS quad, 10 ft contour interval
Between Cow and Rock Creeks	Divide crossing	Flow within Cheney Palouse tract	X		47.09780	-118.05540	Baker (1973, appendix 1, high-water mark 37)	1820	555	USGS quad, 10 ft contour interval
Rock Creek	Divide crossing	Flow within Cheney Palouse tract	X		47.06610	-117.76410	Baker (1973, appendix 1, high-water mark 48)	1890	576	USGS quad, 10 ft contour interval
Rock Creek	Divide crossing	Flow within Cheney Palouse tract	X		47.02870	-117.84530	Baker (1973, appendix 1, high-water mark 49)	1800	549	USGS quad, 10 ft contour interval
Rock Creek	Divide not crossed	Flow within Cheney Palouse tract		X	47.02420	-117.85120	O'Connor interpretation	1900	579	USGS quad, 10 ft contour interval
Near Marengo	Divide not crossed	Flow within Cheney Palouse tract		X	47.01770	-118.16870	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 38)	1750	533	USGS quad, 10 ft contour interval
Near Marengo	Divide crossing; stripped basalt	Flow within Cheney Palouse tract	X		47.02210	-118.17820	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 38)	1720	524	USGS quad, 10 ft contour interval
Palouse River backflooding	Upper limits of erratics, pebbly silts	Flow within Cheney Palouse tract	X		46.98890	-117.75660	Bretz (1929a, p. 401)	1615	492	As reported by Bretz (1929a); location representative
Palouse River at Rebel Creek	Divide Crossing	Flow within Cheney Palouse tract	X		46.93270	-117.79510	O'Connor interpretation of Bretz (1929a, p. 401–402)	1605	489	USGS quad, 5 m contour interval
Rock Creek	Divide crossing	Flow within Cheney Palouse tract	X		46.94830	-117.96230	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 35)		495	USGS quad, 5 m contour interval
Rock Creek	Divide not crossed	Flow within Cheney Palouse tract		X	46.93980	-117.95310	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 35)		500	USGS quad, 5 m contour interval
North of Benge	Divide crossing	Flow within Cheney Palouse tract	X		46.96760	-118.07820	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 34)		515	USGS quad, 5 m contour interval
North of Benge	Divide crossing	Flow within Cheney Palouse tract		X	46.95240	-118.07940	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 34)		515	USGS quad, 5 m contour interval
Union Flat Creek backflooding	Upper limits of erratics, pebbly silts	Flow within Cheney Palouse tract	X		46.86000	-117.86000	Bretz (1929a, p. 402–403)	1680	512	As reported by Bretz (1929a); location representative

(continued on next page)

Table 3 (*continued*)

General location	Feature	Missoula flood context	Minimum constraint	Maximum constraint	Latitude	Longitude	Location source	Elevation (ft)	Elevation (m)	Elevation source
West of Cow Creek	Divide crossing	Flow within Cheney Palouse tract	X		46.83120	-118.17880	Baker (1973, appendix 1, high-water mark 32)		495	USGS quad, 5 m contour interval
West of Cow Creek	Divide not crossed	Flow within Cheney Palouse tract		X	46.81500	-118.17090	Baker (1973, appendix 1, high-water mark 33)		500	USGS quad, 5 m contour interval
North of Washtucna Coulee	Divide crossing	Flow within Cheney Palouse tract	X		46.83960	-118.33260	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 30)	1664	507	Near benchmark on USGS quad
North of Washtucna Coulee	Divide crossing	Flow within Cheney Palouse tract	X		46.81460	-118.30990	Baker (1973, appendix 1, high-water mark 31)	1525	465	Near benchmark on USGS quad
Willow Creek backflooding	Upper limits of erratics, pebbly silts	Flow within Cheney Palouse tract	X		46.79050	-117.89500	Bretz (1929a, p. 405)	1550	472	As reported by Bretz (1929a); location representative
Washtucna Coulee/Snake divide east of Palouse R.	Divide crossing, stripped basalt	Flow spilling across Washtucna Coulee divide into Snake River	X		46.73350	-118.07650	O'Connor interpretation	1510	460	USGS quad, 5 m contour interval
Washtucna Coulee/Snake divide east of Palouse R.	Divide not crossed	Flow spilling across Washtucna Coulee divide into Snake River		X	46.73530	-118.07940	O'Connor interpretation	1670	510	USGS quad, 5 m contour interval
Washtucna Coulee/Snake divide	Divide crossing	Flow spilling across Washtucna Coulee divide into Snake River	X		46.72450	-118.31050	Baker (1973, appendix 1, high-water mark 25)		425	USGS quad, 5 m contour interval
Washtucna Coulee/Snake divide	Divide crossing	Flow spilling across Washtucna Coulee divide into Snake River	X		46.70280	-118.35650	Near or equivalent to Baker (1973, appendix 1, highwater mark 23)		410	USGS quad, 5 m contour interval
Washtucna Coulee/Snake divide	Divide crossing	Flow spilling across Washtucna Coulee divide into Snake River	X		46.70210	-118.34570	Baker (1973, appendix 1, high-water mark 23)		435	USGS quad, 5 m contour interval
Washtucna Coulee/Snake divide	Divide not crossed	Flow spilling across Washtucna Coulee divide into Snake River		X	46.70150	-118.34290	Baker (1973, appendix 1, high-water mark 24)		450	USGS quad, 5 m contour interval
Washtucna Coulee/Snake divide	Divide not crossed	Flow spilling across Washtucna Coulee divide into Snake River		X	46.67760	-118.34740	Baker (1973, appendix 1, high-water mark 26)	1360	420	USGS quad, 5 m contour interval
Palouse Snake confluence	Divide not crossed	Flow within Snake River Canyon		X	46.62430	-118.24480	O'Connor interpretation	1360	415	USGS quad, 5 m contour interval
Palouse Snake confluence	Divide crossing	Flow within Snake River Canyon	X		46.61720	-118.24220	Baker (1973, appendix 1, high-water mark 27; elevation reinterpreted)		390	USGS quad, 5 m contour interval
Palouse Snake confluence	Stripped basalt	Flow within Snake River Canyon	X		46.61260	-118.23510	O'Connor interpretation	1345	410	USGS quad, checked elevation
Kamiah (Clearwater)	Erratic	Snake River backflooding from Cheney Palouse tract inflow	X		46.23371	-116.02961	Kurt Othberg, Sept. 12, 2017, written communication	1204	367	USGS quad, checked elevation
Snake River backflooding	Upper limits of erratics, pebbly silts	Snake River backflooding from Cheney Palouse tract inflow	X		46.5571	-118.17420	Bretz (1929a, p. 405–427); Bretz (1929b, p. 505–509)	1300	396	As reported by Bretz (1929a, 1929b); location representative
Tucannon River backflooding	Erratic	Tucannon River backflooding from Cheney Palouse tract inflow	X		46.5571	-118.17420	Bretz (1929b, p. 509-516)	1320	402	As reported by Bretz (1929b); location representative
Lind Coulee	Divide crossing	Flow within Lind Coulee, from Cheney Palouse tract to Quincy Basin	X		46.91490	-118.74460	Baker (1973, appendix 1, high-water mark 41)	1330	405	USGS quad, 10 ft contour interval

Table 3 (*continued*)

General location	Feature	Missoula flood context	Minimum constraint	Maximum constraint	Latitude	Longitude	Location source	Elevation (ft)	Elevation (m)	Elevation source
Telford-Crab-Creek tract										
Telford Crab Creek tract	Stripped basalt	Flow in Telford Crab Creek tract, Lake Creek coulee	X		47.62160	-118.39680	O'Connor, unpublished, field site 05-20-2019-09	2340	713	USGS quad, 10 ft contour interval
Telford Crab Creek tract	Top of streamlined hille	Flow in Telford Crab Creek tract, Lake Creek coulee		X	47.61821	-118.39856	O'Connor, unpublished, field site 05-20-2019-09	2391	729	Checked elevation on USGS quadrangle
Telford Crab Creek tract	Stripped basalt	Flow in Telford Crab Creek tract, Lake Creek coulee	X		47.58450	-118.45990	O'Connor, unpublished, field site 05-20-2019-11	2310	704	USGS quad, 10 ft contour interval
Telford Crab Creek tract	Top of streamlined hille	Flow in Telford Crab Creek tract, Lake Creek coulee		X	47.58380	-118.46090	O'Connor, unpublished, field site 05-20-2019-11	2347	715	Checked elevation on USGS quadrangle
Wilson Creek Scabland	Stripped basalt	Flow in Telford Crab Creek tract, Wilson Creek coulee	X		47.63463	-118.91333	O'Connor, unpublished, field site 08-15-2018-08	2010	613	USGS quad, 10 ft contour interval
Wilson Creek Scabland	Divide crossing	Flow in Telford Crab Creek tract, Wilson Creek coulee	X		47.45990	-119.04460	Baker (1973, appendix 1, high-water mark 57)	1627	496	Checked elevation on USGS quadrangle
Wilson Creek Scabland	Divide crossing, stripped basalt	Flow in Telford Crab Creek tract, Crab Creek coulee	X		47.38710	-119.05840	Baker (1973, appendix 1, high-water mark 56)	1630	497	USGS quad, 10 ft contour interval
Crab Creek	Trimline, stripped basalt	Flow in Telford Crab Creek tract, Crab Creek coulee	X		47.41050	-119.20620	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 59)	1620	494	USGS quad, 20 ft contour interval
West of Wilson Creek	Trimline, stripped basalt	Flow in Telford Crab Creek tract, entering Quincy Basin south of Crab Creek	X		47.39010	-119.21280	O'Connor interpretation, near Baker (1973, appendix 1, high-water mark 58)	1520	463	USGS quad, 20 ft contour interval
Grand Coulee Quincy Basin Othello Basin Route										
Grand Coulee	Divide not crossed	Head of Grand Coulee		X	47.90540	-118.93730	O'Connor interpretation	2600	792	USGS quad, 40 ft contour interval
Grand Coulee	Stripped basalt	Head of Grand Coulee	X		47.90330	-118.95160	O'Connor interpretation	2480	756	USGS quad, 40 ft contour interval
Grand Coulee	Divide crossing	Upper Grand Coulee	X		47.85010	-119.00920	O'Connor interpretation	2440	744	USGS quad, 10 ft contour interval
Grand Coulee	Stripped basalt	Upper Grand Coulee	X		47.81440	-119.11580	O'Connor interpretation	2357	718	Checked elevation on USGS quadrangle
Grand Coulee	Divide crossing	Grand Coulee, near Dry Falls	X		47.58210	-119.25750	Baker (1973, appendix 1, high-water mark 2)	1810	552	USGS quad, 10 ft contour interval
Grand Coulee	Erratic	Lower Grand Coulee	X		47.50950	-119.47410	Baker (1973, appendix 1, high-water mark 3)	1790	546	USGS quad, 10 ft contour interval
Grand Coulee	Divide not crossed	Lower Grand Coulee		X	47.47600	-119.30210	Baker (1973, appendix 1, high-water mark 5)	1830	558	USGS quad, 10 ft contour interval
Grand Coulee	Divide crossing	Lower Grand Coulee	X		47.47340	-119.28040	Baker (1973, appendix 1, high-water mark 6)	1720	524	USGS quad, 10 ft contour interval
Grand Coulee	Divide crossing	Lower Grand Coulee	X		47.46890	-119.19940	Baker (1973, appendix 1, high-water mark 8)	1700	518	USGS quad, 20 ft contour interval
Grand Coulee	Divide not crossed	Lower Grand Coulee		X	47.46690	-119.20560	Baker (1973, appendix 1, high-water mark 7)	1740	530	USGS quad, 20 ft contour interval
Grand Coulee	Divide crossing	Lower Grand Coulee	X		47.42470	-119.45260	Baker (1973, appendix 1, high-water mark 9)	1460	445	USGS quad, 20 ft contour interval
Grand Coulee	Divide not crossed	Lower Grand Coulee		X	47.41620	-119.44260	Baker (1973, appendix 1, high-water mark 10)	1520	463	USGS quad, 20 ft contour interval
Quincy Basin via Grand Coulee	Scabland	Lower Grand Coulee	X		47.25290	-119.54610	Baker (1973, appendix 1, high-water mark 12)	1340	408	USGS quad, 10 ft contour interval

(*continued on next page*)

Table 3 (*continued*)

General location	Feature	Missoula flood context	Minimum constraint	Maximum constraint	Latitude	Longitude	Location source	Elevation (ft)	Elevation (m)	Elevation source
Quincy Basin	Divide crossing	Lower Grand Coulee	X		47.04930	-119.97040	Baker (1973, appendix 1, high-water mark 13)	1340	408	USGS quad, 10 ft contour interval
Quincy Basin, Drumheller Channels	Erratic	Quincy Basin, Drumheller Channels	X		46.95350	-119.30430	Baker (1973, appendix 1, high-water mark 17)	1340	408	USGS quad, 10 ft contour interval
Quincy Basin, Lind Coulee	Erratic	Quincy Basin, Lind Coulee	X		46.95060	-119.03730	Baker (1973, appendix 1, high-water mark 16)	1350	411	USGS quad, 10 ft contour interval
Quincy Basin; Frenchman Coulee	Divide not crossed	Quincy Basin, Frenchman Coulee		X	46.94470	-119.56810	Baker (1973, appendix 1, high-water mark 18)	1379	420	Checked elevation on USGS quadrangle
Quincy Basin; Frenchman Coulee	Erratic	Quincy Basin, Frenchman Coulee	X		46.94420	-119.54080	Baker (1973, appendix 1, high-water mark 15)	1360	415	USGS quad, 10 ft contour interval
Othello Basin; Crab Creek	Erratic	Othello Basin, lower Crab Creek	X		46.93460	-119.56870	Baker (1973, appendix 1, high-water mark 19	1180	360	USGS quad, 10 ft contour interval
Othello Basin; Saddle Mountains	Divide not crossed	Othello Basin, Othello Channels		X	46.74430	-119.21920	Baker (1973, appendix 1, high-water mark 20)	1210	369	USGS quad, 10 ft contour interval
Othello Basin; Saddle Mountains	Divide crossing	Othello Basin, Othello Channels	X		46.73150	-119.20490	Baker (1973, appendix 1, high-water mark 21)	1155	352	Checked elevation on USGS quadrangle
Moses Coulee										
Moses Coulee, Burton Draw	Crossed divide	Flow into head of Moses Coulee	X		47.63730	-119.59420	Hanson (1970, p. 63; elevation modified by O'Connor)	2310	704	USGS quad, 10 ft contour interval
Moses Coulee, Burton Draw	Divide not crossed	Flow into head of Moses Coulee		X	47.62840	-119.60730	O'Connor interpretation	2380	725	USGS quad, 10 ft contour interval
Moses Coulee, near Coyote Canyon	Stripped basalt	Flow within Moses Coulee	X		47.52030	-119.68360	O'Connor interpretation	1910	582	USGS quad, 10 ft contour interval
Moses Coulee, near Billingsley Ranch	Stripped basalt	Flow within Moses Coulee	X		47.43890	-119.87580	O'Connor interpretation	1850	564	USGS quad, 10 ft contour interval
Moses Coulee, near mouth	Flood bar top	Moses Coulee at Columbia Valley confluence	X		47.30590	-120.02880	O'Connor interpretation	920	280	USGS quad, 10 ft contour interval

All sites were reevaluated using modern imagery and topographic maps, resulting in many elevation and location adjustments from original sources. Elevations of positive evidence of floods, such as eroded rock, crossed divides, and ice-rafted erratics were attributed to the highest elevation contour below the feature, except where nearby benchmarks or checked elevations provided more precise information. Similarly, elevations of features inferred not flooded, such as divides not crossed, were attributed to the lowest elevation contour above the feature. Where original evidence, particularly erratics, could not be evaluated remotely, elevation and location information are noted "as reported."

References

Abril-Hernández, J.M., Periáñez, R., O'Connor, J.E., Garcia-Castellanos, D., 2018. Computational Fluid Dynamics simulations of the Late Pleistocene Lake Bonneville Flood. J. Hydrol. 561, 1–15. https://doi.org/10.1016/J.JHYDROL.2018.03.065.

Adams, K.D., Bills, B.G., 2016. Isostatic rebound and palinspastic restoration of the Bonneville and Provo shorelines in the Bonneville Basin, UT, NV, and ID. Dev. Earth Surf. Process, 145–164 https://doi.org/10.1016/B978-0-444-63590-7.00008-1.

Alho, P., Baker, V.R., Smith, L.N., 2010. Paleohydraulic reconstruction of the largest Glacial Lake Missoula draining(s). Quat. Sci. Rev. 29, 3067–3078. https://doi.org/10.1016/j.quascirev.2010.07.015.

Allen, J.E., Burns, M., Burns, S., 2009. Cataclysms on the Columbia: The Great Missoula Floods. 2nd ed. Ooligan Press.

Allison, I.S., 1933. New version of the spokane flood. Bull. Geol. Soc. Am. 44, 675–722. https://doi.org/10.1130/GSAB-44-675.

Allison, I.S., 1935. Glacial erratics in Willamette Valley. Bull. Geol. Soc. Am. 46, 615–632. https://doi.org/10.1130/GSAB-46-615.

Allison, I.S., 1941. Flint's fill hypothesis of origin of scabland. J. Geol. 49, 54–73. https://doi.org/10.1086/624944.

Allison, I.S., 1978. Late Pleistocene sediments and floods in the Willamette Valley. The Ore Bin 40 (177–191), 192–202.

Alt, D., Chambers, R.L., 1970. Repetition of the Spokane flood [abs.]. American Quaternary Association Meeting. Montana State University, Bozeman 1.

Amidon, W.H., Clark, A.C., 2015. Interaction of outburst floods with basaltic aquifers on the Snake River Plain: implications for Martian canyons. Bull. Geol. Soc. Am. 127, 688–701. https://doi.org/10.1130/B31141.1.

Amidon, W.H., Farley, K.A., 2011. Cosmogenic 3He dating of apatite, zircon and pyroxene from Bonneville flood erosional surfaces. Quat. Geochronol. 6, 10–21. https://doi.org/10.1016/j.quageo.2010.03.005.

Archaeological Investigations Northwest, 2015. Archaeological Damage Assessment and Monitoring Summary for the Mill Creek North Mitigation Site for the SR 502 Corridor Widening Project, Clark County, Washington Report 3360 (for Washington State Department of Transportation, Vancouver, WA).

Arnow, T., Stephens, D.W., 1990. Hydrologic characteristics of the Great Salt Lake, Utah, 1847-1986. U.S. Geol. Sur. Water Supply Pap. 2332. https://doi.org/10.3133/wsp2332.

Atwater, B.F., 1984. Periodic floods from glacial Lake Missoula into the Sanpoil arm of glacial Lake Columbia, northeastern Washington. Geology 12, 464–467. https://doi.org/10.1130/0091-7613(1984)12<464:PFFGLM>2.0.CO;2.

Atwater, B.F., 1986. Pleistocene glacial-lake deposits of the Sanpoil River Valley, northeastern Washington. U.S. Geol. Surv. Bull. 1661. https://doi.org/10.3133/b1661.

Atwater, B.F., 1987. Status of glacial Lake Columbia during the last floods from glacial Lake Missoula. Quat. Res. 27, 182–201. https://doi.org/10.1016/0033-5894(87)90076-7.

Atwater, B.F., Smith, G.A., Waitt, R.B., 2000. The Channeled Scabland: back to Bretz? Comment. Geology 28, 574–575. https://doi.org/10.1130/0091-7613(2000)28<576:TCSBTB>2.0.CO;2.

Bader, N.E., Spencer, P.K., Bailey, A.S., Gastineau, K.M., Tinkler, E.R., Pluhar, C.J., Bjornstad, B.N., 2016. A loess record of pre-Late Wisconsin glacial outburst flooding, Pleistocene paleoenvironment, and Irvingtonian fauna from the Rulo site, southeastern Washington, USA. Palaeogeogr. Palaeoclimatol. Palaeoecol. 462, 57–69. https://doi.org/10.1016/j.palaeo.2016.08.042.

Baker, V.R., 1973. Paleohydrology and Sedimentology of Lake Missoula Flooding in Eastern Washington. Geological Society of America Special Paper 144. Geological Society of America https://doi.org/10.1130/SPE144-p1.

Baker, V.R., 1978a. The Spokane Flood controversy and the Martian outflow channels. Science 202, 1249–1256. https://doi.org/10.1126/science.202.4374.1249.

Baker, V.R., 1978b. Quaternary geology of the channeled scabland and adjacent areas. In: Baker, V.R., Nummeddal, D. (Eds.), The Channeled Scabland. National Aeronautics and Space Administration Planetary Geology Program, Washington, D.C., pp. 17–35

Baker, V.R. (Ed.), 1981. Catastrophic Flooding: The Origin of the Channeled Scabland. Dowden, Hutchinson & Ross, Stroudsburg, PA.

Baker, V.R., 1982. The Channels of Mars. University of Texas Press, Austin, Texas.

Baker, V.R., 1995. Joseph Thomas Pardee and the Spokane flood controversy. GSA Today 5, 169–173.

Baker, V.R., 1998. Catastrophism and uniformitarianism: logical roots and current relevance in geology. In: Blundell, D.J., Scott, A.C. (Eds.), Lyell: The Past is the Key to the Present, pp. 171–182. https://doi.org/10.1144/GSL.SP.1998.143.01.15 Geological Society of London, Special Publication 143.

Baker, V.R., 2008. The Spokane Flood debates: historical background and philosophical perspective. In: Grapes, R.H., Oldroyd, D., Grigelis, A. (Eds.), History of Geomorphology and Quaternary Geology, pp. 33–50. https://doi.org/10.1144/SP301.3 Geological Society of London, London.

Baker, V.R., 2009. The channeled scabland: a retrospective. Annu. Rev. Earth Planet. Sci. 37, 393–411. https://doi.org/10.1146/annurev.earth.061008.134726.

Baker, V.R., 2013. Global late quaternary fluvial paleohydrology: with special emphasis on paleofloods and megafloods. In: Wohl, E.E. (Ed.), Treatise on Geomorphology. Elsevier, pp. 511–527. https://doi.org/10.1016/B978-0-12-374739-6.00252-9.

Baker, V.R., 2020. Global megaflood paleohydrology. In: Herget, J., Fontana, A. (Eds.), Palaeohydrology. Geography of the Physical Environment. Springer, Heidelberg, pp. 3–28. https://doi.org/10.1007/978-3-030-23315-0_1.

Baker, V.R., Bunker, R.C., 1985. Cataclysmic late Pleistocene flooding from glacial Lake Missoula: a review. Quat. Sci. Rev. 4, 1–41. https://doi.org/10.1016/0277-3791(85)90027-7.

Baker, V.R., Milton, D.J., 1974. Erosion by catastrophic floods on Mars and Earth. Icarus 23, 27–41. https://doi.org/10.1016/0019-1035(74)90101-8.

Baker, V.R., Bjornstad, B.N., Busacca, A.J., Fecht, K.R., Kiver, E.P., Moody, U.L., Rigby, J.G., Stradling, D.F., Tallman, A.M., 1991. Quaternary geology of the Columbia Plateau. In: Morrison, R.G. (Ed.), Quaternary Nonglacial Geology. Geological Society of America, Boulder, Colorado, pp. 215–250. https://doi.org/10.1130/DNAG-GNA-K2.215.

Baker, D., Peterson, C., Hemphill-Haley, E., Twichell, D., 2010. Latest Pleistocene and Holocene (2–16ka) sedimentation in the Columbia River Estuary, Oregon. USA. Mar. Geol. 273, 83–95. https://doi.org/10.1016/j.margeo.2010.02.005.

Baker, V.R., Bjornstad, B.N., Gaylord, D.R., Smith, G.A., Meyer, S.E., Alho, P., Breckenridge, R.M., Sweeney, M.R., Zreda, M., 2016. Pleistocene megaflood landscapes of the channeled scabland. Exploring the Geology of the Inland Northwest. Geol. Soc. Am. Field Trip Guide 41, pp. 1–73. https://doi.org/10.1130/2016.0041(01).

Balbas, A.M., Barth, A.M., Clark, P.U., Clark, J., Caffee, M., O'Connor, J., Baker, V.R., Konrad, K., Bjornstad, B., O'Connor, J., Baker, V.R., Konrad, K., Bjornstad, B., 2017. ^{10}Be dating of late Pleistocene megafloods and Cordilleran Ice Sheet retreat in the northwestern United States. Geology 45, 583–586. https://doi.org/10.1130/G38956.1.

Beeson, J.W., Goldfinger, C., Fortin, W.F., 2017. Large-scale modification of submarine geomorphic features on the Cascadia accretionary wedge caused by catastrophic flooding events. Geosphere 13, 1713–1728. https://doi:10.1130/GES01388.1.

Beget, J.E., 1986. Comment on "outburst floods from glacial Lake Missoula" by G. K. C. Clarke, W. H. Mathews, and R. T. Pack. Quat. Res. 25, 136–138. https://doi.org/10.1016/0033-5894(86)90050-5.

Benito, G., 1997. Energy expenditure and geomorphic work of the cataclysmic Missoula flooding in the Columbia River Gorge, USA. Earth Surf. Process. Landf. 22, 457–472. https://doi.org/10.1002/(SICI)1096-9837(199705)22:5<457::AID-ESP762>3.0.CO;2-Y.

Benito, G., O'Connor, J.E., 2003. Number and size of last-glacial Missoula floods in the Columbia River valley between the Pasco Basin, Washington, and Portland. Oregon Geol. Soc. Amer. Bull. 115, 624–638. https://doi.org/10.1130/0016-7606(2003)115%3C0624:NASOLM%3E2.0.CO;2.

Berg, R.B., 2005. Geologic map of the Upper Clark Fork Valley between Garrison and Bearmouth, southwestern Montana. Open-File Report 523. Montana Bureau of Mines and Geology, Butte.

Berg, R.B., 2006. Geologic map of the Upper Clark Fork Valley between Bearmouth and Missoula, southwestern Montana. Open-File Report 535. Montana Bureau of Mines and Geology, Butte.

Berger, M.J., George, D.L., LeVeque, R.J., Mandli, K.T., 2011. The GeoClaw software for depth-averaged flows with adaptive refinement. Adv. Water Resour. 34, 1195–1206. https://doi.org/10.1016/J.ADVWATRES.2011.02.016.

Bjornstad, B.N., 1980. Sedimentology and Depositional Environment of the Touchet Beds, Walla Walla River Basin, Washington. Eastern Washington University.

Bjornstad, B.N., 2014. Ice-rafted erratics and bergmounds from Pleistocene outburst floods, Rattlesnake Mountain, Washington, USA. Quat. Sci. J. 63, 44–59. https://doi.org/10.3285/eg.63.1.03.

Bjornstad, B.N., Fecht, K.R., Tallman, A.M., 1991. Quaternary stratigraphy of the Pasco Basin, south-central Washington. Baker, V.R., Bjornstad, B.N., Busacca, A.J., Fecht, K.R., Kiver, E.P., Moody, U.L., Rigby, J.G., Stradling, D.F., and Tallman, A.M., authors. In: Morrison, R.G. (Ed.), Quaternary Geology of the Columbia Plateau, pp. 238–250. https://doi.org/10.1130/DNAG-GNA-K2 Quaternary Nonglacial Geology; Conterminous U.S.: Boulder; Colorado; Geological Society of America; Geology of North America; v. K-2.

Bjornstad, B.N., Fecht, K.R., Pluhar, C.J., 2001. Long history of pre-Wisconsin, ice age cataclysmic floods: evidence from southeastern Washington state. J. Geol. 109, 695–713. https://doi.org/10.1086/323190.

Bolke, E.L.V., Washington, J.J., 1980. Digital model simulation of the hydrologic flow system, with emphasis on ground water in Spokane Valley, Washington and Idaho. U.S. Geol. Surv. Open-File Report 80-1300. https://doi.org/10.3133/ofr801300.

Booth, D.B., 1986. Mass balance and sliding velocity of the Puget lobe of the Cordilleran ice sheet during the last glaciation. Quat. Res. 25, 269–280. https://doi.org/10.1016/0033-5894(86)90001-3.

Breckenridge, R.M., Othberg, K.L., 1998a. Surficial geologic map of the post falls quadrangle and part of the Liberty Lake Quadrangle, Kootenai County, Idaho. Idaho Geol. Surv. Surficial Geol. Maps SGM-5.

Breckenridge, R.M., Othberg, K.L., 1998b. Surficial geologic map of the Rathdrum quadrangle and part of the Newman Lake quadrangle, Kootenai County, Idaho. Idaho Geol. Surv. Surficial Geol. Maps SGM-6.

Breckenridge, R.M., Phillips, W.M., 2010. New cosmogenic 10Be surface exposure ages for the Purcell Trench lobe of the Cordilleran ice sheet in Idaho [abs.]. Geol. Soc. Am. Abstr. Programs 42 (5), 309.

Breckenridge, R.M., Atwater, B.F., Baker, V.R., Busacca, A.J., Chambers, R.L., Curry, R.R., Hanson, L.G., Kever, E.P., McDonald, E.V., Stradling, D.F., Waitt, R.B., 1989. Glacial Lake Missoula and the channeled scabland. International Geological Congress, 28th, Guidebook. American Geophysical Union, Washington D.C. https://doi.org/10.1029/FT310.

Bretz, J H., 1919. The late Pleistocene submergence in the Columbia Valley of Oregon and Washington. J. Geol. 27, 489–506. https://www.jstor.org/stable/30058413.

Bretz, J H., 1923a. Glacial drainage on the Columbia Plateau (with discussion by M. M. Leighton and O. E. Meinzer). Bull. Geol. Soc. Am. 34, 573–608. https://doi.org/10.1130/GSAB-34-573.

Bretz, J H., 1923b. The channeled scablands of the Columbia Plateau. J. Geol. 31, 617–649. https://doi.org/10.1086/623053.

Bretz, J H., 1924. The Dalles type of river channel. J. Geol. 32, 139–149. https://doi.org/10.1086/623074.

Bretz, J H., 1925a. The Spokane flood beyond the channeled scablands (I). J. Geol. 33, 97–115. https://doi.org/10.1086/623179.

Bretz, J H., 1925b. The Spokane Flood beyond the Channeled Scablands (II). J. Geol. 33, 236–259. https://doi.org/10.1086/623197.

Bretz, J H., 1927a. Channeled scabland and the Spokane Flood. J. Wash. Acad. Sci. 17, 200–211.

Bretz, J H., 1927b. What caused the Spokane flood [abs.]? Bull. Geol. Soc. Am. 28, 107.

Bretz, J H., 1928a. Bars of channeled Scabland. Bull. Geol. Soc. Am. 39, 643–701. https://doi.org/10.1130/GSAB-39-643.

Bretz, J H., 1928b. Channeled scabland of eastern Washington. Geogr. Rev. 18, 446–477. https://doi.org/10.2307/208027.

Bretz, J H., 1929a. Valley deposits immediately east of the Channeled Scabland of Washington I. J. Geol. 37, 393–427. https://doi.org/10.1086/623636.

Bretz, J H., 1929b. Valley deposits immediately east of the channeled Scabland of Washington. II. J. Geol. 37, 505–541. https://doi.org/10.1086/623636.

Bretz, J H., 1930. Valley deposits immediately west of the Channeled Scabland. J. Geol. 38, 385–422. https://doi.org/10.1086/623737.

Bretz, J H., 1932. The Grand Coulee. Special Publication No. 15. American Geographical Society,, New York.

Bretz, J H., 1969. The Lake Missoula floods and the Channeled Scabland. J. Geol. 77, 505–543. https://doi.org/10.1086/627452.

Bretz, J H., Smith, H.T.U., Neff, G.E., 1956. Channeled scabland of Washington: new data and interpretations. Bull. Geol. Soc. Am. 67, 957–1049. https://doi.org/10.1130/0016-7606(1956)67[957:CSOWND]2.0.CO;2.

Bright, R.C., 1963. Pleistocene Lakes Thatcher and Bonneville, Southeastern Idaho. University of Minnesota.

Bronk Ramsey, C., 2009. Bayesian analysis of radiocarbon dates. Radiocarbon 51, 337–360. https://doi.org/10.1017/S0033822200033865.

Brunner, C.A., Normark, W.R., Zuffa, G.G., Serra, F., 1999. Deep-sea sedimentary record of the late Wisconsin cataclysmic floods from the Columbia River. Geology 27, 463–466. https://doi.org/10.1130/0091-7613(1999)027%3C0463:DSSROT%3E2.3.CO;2.

Buehler, J.S., Shearer, P.M., 2010. Pn tomography of the western United States using USArray. J. Geophys. Res. 115, B09315. https://doi.org/10.1029/2009JB006874.

Bunker, R.C., 1982. Evidence of multiple late-Wisconsin floods from glacial Lake Missoula in Badger, Coulee, Washington. Quat. Res. 18, 17–31. https://doi.org/10.1016/0033-5894(82)90019-9.

Burr, D.M., Carling, P.A., Baker, V.R. (Eds.), 2009. Megaflooding on Earth and Mars. Cambridge University Press, New York ISBN-10:0521868521.

Busacca, A.J., McDonald, E.V., 1994. Regional sedimentation of late Quaternary loess on the Columbia Plateau: Sediment source areas and loess distribution patterns. Wash. Div. Geol. Earth Resour. Bull. 80, 181–190.

Carling, P.A., Kidson, R., Cao, Z., Herget, J., 2003. Palaeohydraulics of extreme flood events; reality and myth. In: Gregory, K., Benito, G. (Eds.), Palaeohydrology: Understanding Global Change. Wiley, Chichester, pp. 325–336.

Carrara, P.E., Kiver, E.P., Stradling, D.F., 1996. The southern limit of cordilleran ice in the Colville and Pend Oreille valleys of northeastern Washington during the Late Wisconsin glaciation. Can. J. Earth Sci. 33, 769–778. https://doi.org/10.1139/e96-059.

Carter, D.T., Ely, L.L., O'Connor, J.E., Fenton, C.R., 2006. Late Pleistocene outburst flooding from pluvial Lake Alvord into the Owyhee River, Oregon. Geomorphology 75, 346–367. https://doi.org/10.1016/J.GEOMORPH.2005.07.023.

Cerling, T.E., Poreda, R.J., Rathburn, S.L., 1994. Cosmogenic ^{3}He and ^{21}Ne age of the big lost river flood, Snake River Plain, Idaho. Geology 22, 227–230. https://doi.org/10.1130/0091-7613(1994)022<0227:CHANAO>2.3.CO;2.

Chambers, R.L., 1971. Sedimentation in Glacial Lake Missoula. University of Montana, Missoula, Missoula.

Chambers, R.L., 1984. Sedimentary evidence for multiple glacial Lakes Missoula. In: McBane, J.D., Garrison, P.B. (Eds.), Northwest Montana and Adjacent Canada. Montana Geological Society, Billings, pp. 189–199.

Clague, J.J., 1975. Sedimentology and paleohydrology of late Wisconsinan outwash, Rocky Mountain Trench, southeastern British Columbia. In: Jopling, A.V., McDonald, B.C.

(Eds.), Glaciofluvial and Glaciolacustrine Sedimentation. Society of Economic Paleontologists and Mineralogists, pp. 223–237 Special Publication 23.
Clague, J.J., 1980. Late Quaternary geology and geochronology of British Columbia. Geol. Surv. Can. Pap. 80-13.
Clague, J.J., 1989. Cordilleran ice sheet. In: Fulton, R.J. (Ed.), Quaternary Geology of Canada and Greenland. Vol. 1 Geological Survey of Canada, Geology of Canada, Ottawa, Ontario, 40–42 (also Geological Society of America, The Geology of North America, Vol. K-1) Boulder, CO.
Clague, J.J., Armstrong, J.E., Mathews, W.H., 1980. Advance of the late Wisconsin Cordilleran ice sheet in southern British Columbia since 22000 yr. BP. Quat. Res. 13, 322–326. https://doi.org/10.1016/0033-5894(80)90060-5.
Clague, J.J., Barendregt, R., Enkin, R.J., Foit, F.F., 2003. Paleomagnetic and tephra evidence for tens of Missoula floods in southern Washington. Geology 31, 247–250. https://doi.org/10.1130/0091-7613(2003)031<0247:PATEFT>2.0.CO;2.
Clarke, G.K.C., Mathews, W.H., Pack, R.T., 1984. Outburst floods from glacial Lake Missoula. Quat. Res. 22, 289–299. https://doi.org/10.1016/0033-5894(84)90023-1.
Clynne, M.A., Calvert, A.T., Wolfe, E.W., Evarts, R.C., Fleck, R.J., Lanphere, M.A., 2008. The Pleistocene eruptive history of Mount St. Helens, Washington, from 300,000 to 12,800 years before present. U.S. Geol. Surv. Prof. Pap. 1750, 593–628. https://doi.org/10.3133/pp175028.
Cochran, D.S., Warlow, J.C., 1980. Engineering and Environmental Geology, Mt. Tolman projet, Ferry County, Washington Nespelem.
Cordero, D.I., 1997. Early to Middle Pleistocene Catastrophic Flood Deposits. Portland State University, The Dalles, Oregon.
Cosma, T., Hendy, I.L., 2008. Pleistocene glaciomarine sedimentation on the continental slope off Vancouver Island, British Columbia. Mar. Geol. 255, 45–54. https://doi.org/10.1016/j.margeo.2008.07.001.
Cosma, T.N., Hendy, I.L., Chang, A.S., 2008. Chronological constraints on Cordilleran Ice Sheet glaciomarine sedimentation from core MD02-2496 off Vancouver Island (western Canada). Quat. Sci. Rev. 27, 941–955.
Craig, R.G., 1987. Dynamics of a Missoula flood. In: Mayer, L., Nash, D. (Eds.), Catastrophic Flooding. Allen & Unwin, London, pp. 305–332.
Craig, R.G., Hanson, J.P., 1985. Erosion potential from Missoula floods in the Pasco Basin, Washington. Battelle Pacific Northwest Lab. Doc. PNL-5684.
Cressman, L.S., 1960. Cultural sequences at the Dalles, Oregon—a contribution to Pacific Northwest prehistory. Trans. Am. Philos. Soc. 50 (10), 1–108. https://doi.org/10.2307/1005853.
Cressman, L.S., 1977. Prehistory of the Far West—homes of Vanished Peoples. University of Utah Press ISBN 10: 0874801133.
Currey, D.R., 1982. Lake Bonneville: selected features of relevance to neotectonic analyis. U.S. Geol. Surv. Open-File Rep. https://doi.org/10.3133/ofr821070 82-1070.
Curry, R.R., Lister, J.C., Stoffel, K., 1977. Glacial history of Flathead Valley and Lake Missoula floods. In: Smith, D.G., Curry, R.R., Lister, J.C., Stoffel, K. (Eds.), Glacial Geology of Flathead Valley and Catastrophic Drainage of Glacial Lake Missoula, pp. 14–38 Geological Society of America, Rocky Mountain Section 30th Annual Meeting, field Guide, Missoula.
Davis, L.G., et al., 2019. Late Upper Paleolithic occupation at Cooper's Ferry, Idaho, USA, ~16,000 years ago. Science 365, 891–897. https://doi.org/10.1126/science.aax9830.
Dawson, W.L., 1898. Glacial phenomena in Okanogan County, Washington. Am. Geol. 22, 203–217.
Denlinger, R.P., O'Connell, D.R.H., 2010. Simulations of cataclysmic outburst floods from Pleistocene glacial lake Missoula. Bull. Geol. Soc. Am. 122, 678–689. https://doi.org/10.1130/B26454.1.
Denlinger, R.P. George, D.L., Cannon, C., O'Connor, J., Waitt, R.B., in press. Diverse cataclysmic floods from Pleistocene glacial Lake Missoula. Geological Society of America Special Paper 574, 10.1130/2020.2548(17).
Dethier, D.P., Pessi Jr., F., Keuler, R.F., Balzarini, M.A., Pevear, D.R., 1995. Late Wisconsinan glaciomarine deposition and isostatic rebound, northern Puget Lowland, Washington. Bull. Geol. Soc. Am. 107, 1288–1303. https://doi.org/10.1130/0016-7606(1995)107<1288:LWGDAI>2.3.CO;2.
Doak, S., O'Connor, J.E., Haugerud, R., 2019. Mapping the Missoula floods, eastern Washington. Geol. Soc. Am. Abstr. Programs 51-4.
Dunbar, M., 1998. Missoula flood deposits in Kittitas County. Central Washington University.
Evarts, R.C., O'Connor, J.E., 2008. Geologic map of the Camas Quadrangle, Clark County, Washington, and Multnomah County, Oregon. U.S. Geol. Surv. Sci. Inv. Map 3017. https://doi.org/10.3133/sim3017.
Evarts, R., O'Connor, J.E., Wells, R.E., Madin, I.P., 2009. The Portland basin: a (big) river runs through it. GSA Today, 4–10 https://doi.org/10.1130/GSATG58A.1.
Fields, R.L., Woods, P.F., Berenbrock, C., 1996. Bathymetric map of Lake Pend Oreille and Pend Oreille River, Idaho. U.S. Geol. Surv. Water-Resour. Invest. Rep. https://doi.org/10.3133/wri964189 96-4189.
Flint, R.F., 1935. Glacial features of the southern Okanogan region. Bull. Geol. Soc. Am. 46, 169–194. https://doi.org/10.1130/GSAB-46-169.
Flint, R.R., 1936. Stratified drift and deglaciation of eastern Washington. Bull. Geol. Soc. Am. 47, 1849–1884. https://doi.org/10.1130/GSAB-47-1849.
Flint, R.F., 1937. Pleistocene drift border in eastern Washington. Bull. Geol. Soc. Am. 48, 203–232. https://doi.org/10.1130/GSAB-48-203.
Flint, R.F., 1938. Origin of the Cheney-Palouse scabland tract, Washington. Bull. Geol. Soc. Am. 49, 461–524. https://doi.org/10.1130/GSAB-49-461.
Flint, R.F., Irwin, W.H., 1939. Glacial geology of Grand Coulee Dam, Washington. Bull. Geol. Soc. Am. 50, 661–680. https://doi.org/10.1130/GSAB-50-661.
Foley, L.L., 1976. Slack Water Sediments in the Alpowa Creek Drainage. Washington State University, Washington.
Fritz, W.J., Smith, G.A., 1993. Revisiting the Ninemile section: Problems with relating glacial Lake Missoula stratigraphy to the Scabland-floods stratigraphy [abs.]. EOS (Transactions, American Geophysical Union). American Geophysical Union 74, 302.
Fryxell, R., 1962. A radiocarbon limiting date for scabland flood. Northwest Sci. 6, 113–119.
Fulton, R.J., 1971. Radiocarbon geochronology of Southern British Columbia. Geol. Surv. Canada 71, 28. https://doi.org/10.4095/102467.
Garcia-Castellanos, D., O'Connor, J.E., 2018. Outburst floods provide erodability estimates consistent with long-term landscape evolution. Sci. Rep. 8, 10573. https://doi.org/10.1038/s41598-018-28981-y.
Gaylord, D.R., Foit Jr., F.F., Schatz, J.K., Coleman, A.J., 2001. Smith Canyon dune field, Washington, U.S.A.: relation to glacial outburst floods, the Mazama eruption, and Holocene paleoclimate. J. Arid Environ. 47, 403–424. https://doi.org/10.1006/jare.2000.0731.
Gaylord, D.R., Meyer, S.E., Breckenridge, R.N., 2016. Glacial lake Columbia sedimentation: sedimentology and stratigraphy of late quaternary megaflood-influenced sedimentary deposits in the Latah Creek Valley near Spokane, Washington. Pleistocene megaflood landscapes of the Channeled Scabland. In: Baker, V.R., Bjornstad, B.N., Gaylord, D.R., Smith, G.A., Meyer, S.E., Alho, P., Zreda, M. (Eds.), Exploring the Geology of the Inland Northwest. Geological Society of America, pp. 18–26 https://doi.org/10.1130/2016.0041(01).
Gerstel, W.J., Palmer, S.P., 1994. Geologic and geophysical mapping of the Spokane Aquifer-Relevance to growth management. Wash. Geol. 22, 18–24.
Gilbert, G.K., 1890. Lake Bonneville. U.S. Geol. Surv. Monogr. 1. https://doi.org/10.3133/m1.
Gilmour, D.M., Butler, V.L., O'Connor, J.E., Davis, E.B., Culleton, B.J., Kennett, D.J., Hodgins, G., 2015. Chronology and ecology of late Pleistocene megafauna in the northern Willamette Valley, Oregon. Quat. Res. 83, 127–136. https://doi.org/10.1016/j.yqres.2014.09.003.
Gleason, A., Markert, J., 2019. Washington State Lidar Plan: A Plan for Statewide Lidar Coverage in Washington. Olympia. https://www.dnr.wa.gov/publications/ger_wa_lidar_plan_2019.pdf accessed Nov. 7, 2019.
Glenn, J.L., 1965. Late Quaternary Sedimentation and Geologic History of the North Willamette Valley, Oregon. Oregon State University.
Gombiner, J.H., Hemming, S.R., Hendy, I.L., Bryce, J.G., Blichert-Toft, J., 2016. Isotopic and elemental evidence for Scabland Flood sediments offshore Vancouver Island. Quat. Sci. Rev. 139, 129–137. https://doi.org/10.1016/j.quascirev.2016.02.026.
Gordon, G., Haugerud, R., O'Connor, J.E., 2017. Mapping the Missoula floods: the Columbia River Gorge. Geol. Soc. Am. Abstr. Programs 49. https://doi.org/10.1130/abs/2017AM-301734.
Goudie, A.S., 2008. The quaternary. In: Burt, T.P., Chorley, R.J., Brunsden, D., Cox, H.J., Goudie, A.S. (Eds.), The History of the Study of Landforms or the Development of Geomorphology, volume 4, Quaternary and Recent Processes and Forms (1890–1965) and the Mid-Century Revolutions. The Geological Society, London, pp. 397–509 Chapter 10. ISBN 10: 1862392498.
Gould, S.J., 1978. The great scablands debate. Nat. Hist. 87, 12–18.
Gould, S.J., 1980. The great scablands debate. The Panda's Thumb: More Reflections in Natural History. Norton, New York, pp. 194–203 ISBN: 9780393013801.
Griggs, G.B., Kulm, L.D., Waters, A.C., Fowler, G.A., 1970. Deep-sea gravel from Cascadia Channel. J. Geol. 78, 611–619. https://doi.org/10.1086/627560.
Hanson, L.G., 1970. The Origin and Deformation of Moses Coulee and Other Scabland Features on the Waterville Plateau. University of Washington.
Hanson, M.A., 2013. Sedimentological and Paleomagnetic Study of Glacial Lake Missoula Lacustrine and flood Sediment. Simon Fraser University http://summit.sfu.ca/item/12949.
Hanson, M.A., Clague, J.J., 2016. Record of glacial Lake Missoula floods in glacial Lake Columbia, Washington. Quat. Sci. Rev. 133, 62–76. https://doi.org/10.1016/j.quascirev.2015.12.009.
Hanson, M.A., Lian, O.B., Clague, J.J., 2012. The sequence and timing of large late Pleistocene floods from glacial Lake Missoula. Quat. Sci. Rev. 31, 67–81. https://doi.org/10.1016/j.quascirev.2011.11.009.
Hanson, M.A., Enkin, R.J., Barendregt, R.W., Clague, J.J., 2015. Provenance and deposition of glacial Lake Missoula lacustrine and flood sediments determined from rock magnetic properties. Quat. Res. 83, 166–177. https://doi:10.1016/j.yqres.2014.09.005.
Harpel, C.J., 1996. Paleodischarges andinferred hydraulics of the Missoula Ffloods: Columbia River valley and Channeled Scabland, eastern Washington Western Washington University Senior Thesis.
Harpel, C.J., Waitt, R.B., O'Connor, J.E., 2000. Paleodischarges of the Late Pleistocene Missoula Floods, Eastern Washington, U.S.A. [abs.]: Orkustofnun Rept. OS-2000/036, Reykjavík, Extremes of the Extremes Conference, 21.
Hendrick, M., Haugerud, R., O'Connor, J.E., 2017. Mapping the Missoula floods: North-central Washington. Geol. Soc. Am. Abstr. Programs 49 (6). https://doi.org/10.1130/abs/2017AM-301715.
Hendy, I., 2009. A fresh perspective on the Cordilleran Ice Sheet. Geology 37, 95–96. https://doi.org/10.1130/focus012009.1.
Hibbert, D., 1985. Quaternary geology and the history of the landscape along the Columbia River between Chief Joseph and Grand Coulee Dams. In: Campbell, S.K. (Ed.), Summary of Results, Chief Joseph Dam Cultural Resources Project, Washington. Office of Public Archaeology Institue for Environmental Studies, University of Washington , pp. 85–145. https://apps.dtic.mil/docs/citations/ADA166700.
Hodge, E.T., 1931. Exceptional moraine-like deposits in Oregon. Bull. Geol. Soc. Am. 42, 985–1010. https://doi.org/10.1130/GSAB-42-985.
Hodge, E.T., 1934. Origin of the Washington scabland. Northwest Sci. 8, 4–11.
Hodge, E.T., 1938. Geology of the lower Columbia River. Bull. Geol. Soc. Am. 49, 831–930. https://doi.org/10.1130/GSAB-49-831.
Hoffman, M.H., Hendrix, M.S., 2010. Depositional processes and the inferred history of ice-margin retreat associated with the deglaciation of the Cordilleran Ice Sheet: the sedimentary record from Flathead Lake, northwest Montana, USA. Sediment. Geol. 223, 61–74. https://doi.org/10.1016/j.sedgeo.2009.10.004.
Hunt, C.B., 1982. Pleistocene Lake Bonneville, ancestral Great Salt Lake, as described in the notebooks of G. K. Gilbert, 1875–1880. Brigham Young Univ. Geol. Stud. 29, 225.

Jarrett, R.D., Malde, H.E., 1987. Paleodischarge of the late Pleistocene Bonneville Flood, Snake River, Idaho, computed from new evidence. Bull. Geol. Soc. Am. 99, 127–134. https://doi.org/10.1130/0016-7606(1987)99<127:POTLPB>2.0.CO;2.

Jones, F.O., Embody, D.R., Peterson, W.L., Hazlewood, R.M., 1961. Landslides along the Columbia River Valley, northeastern Washington. US Geol. Surv. Prof. Pap. 367. https://doi.org/10.3133/pp367.

Kahle, S.C., Bartolino, J.R., 2007. Hydrogeologic framework and ground-water budget of the Spokane Valley-Rathdrum Prairie Aquifer, Spokane County, Washington, and Bonner and Kootenai Counties, Idaho. US Geol. Surv. Sci. Investig. Rep. https://doi.org/10.3133/sir20075041 5007-5041.

Keszthelyi, L.P., Baker, V.R., Jaeger, W.L., Gaylord, D.R., Bjornstad, B.N., Greenbaum, N., Self, S., Thordarson, T., Porat, N., Zreda, M.G., 2009. Floods of water and lava in the Columbia River Basin: analogs for Mars. Volcanoes to Vineyards: Geologic Field Trips through the Dynamic Landscape of the Pacific Northwest. Geological Society of America, pp. 845–874 https://doi.org/10.1130/2009.fld015(34).

Kiver, E.P., Stradling, D.F., 1982. Upper Columbia River terrace system of eastern Washington-Preliminary interpretations [abs]. American Quaternary Association Program and Abstracts. 117. American Quaternary Association ISSN: 0741-059X.

Kiver, E.P., Stradling, D.F., 1985. Comments on "Periodic Jökulhlaups from Pleistocene Glacial Lake Missoula—New Evidence from Varved Sediments in Northern Idaho and Washington," by Richard B. Waitt, Jr. Quat. Res. 24, 354–356. https://doi.org/10.1016/0033-5894(85)90057-2.

Kiver, E.P, Stradling, D.F., 1989. The Spokane Valley and northern Columbia Plateau. In: Breckenridge, R.M., Atwater, B.F., Baker, V.R., Busacca, A.J., Chambers, R.L., Curry, R.R., Waitt, R.B. (Eds.), Glacial Lake Missoula and the Channeled Scabland, in: International Geological Congress, 28th, Guidebook. American Geophysical Union, Washington D.C., pp. 23–35. https://doi.org/10.1029/FT310.

Kiver, E.P., Stradling, D.F., 1995. Geology of the Franklin D. Roosevelt Shoreline: Glacial Geology, Terraces, Landslides, and Lineaments. Grand Coulee Power Office, Bureau of Reclamation, United States Department of the Interior https://archive.org/details/geologyoffrankli00kive/.

Kiver, E.P., Stradling, D.F., Moody, U.L., 1989. Glacial and multiple flood history of the northern borderlands—Trip B. In: Joseph, N.L., et al. (Eds.), Geologic Guidebook for Washington and Adjacent Areas. Washington Division of Geology and Earth Resources Information Circular 86, pp. 321–335.

Kiver, E.P., Moody, U.L., Rigby, J.G., Stradling, D.F., 1991. Late Quaternary stratigraphy of the Channeled Scabland and adjacent areas. Baker, V.R., Bjornstad, B.N., Busacca, A.J., Fecht, K.R., Kiver, E.P., Moody, U.L., Rigby, J.G., Stradling, D.F., and Tallman, A.M., authors. In: Morrison, R.G. (Ed.), Quaternary Geology of the Columbia Plateau, pp. 238–250. https://doi.org/10.1130/DNAG-GNA-K2 Quaternary Nonglacial Geology; Conterminous U.S.: Boulder; Colorado; Geological Society of America; Geology of North America; v. K-2.

Komatsu, G., Miyamoto, H., Ito, K., Tosaka, H., Tokunaga, T., 2000. The channeled scabland: back to bretz?: comment and reply. Geology 28, 573. https://doi.org/10.1130/0091-7613(2000)28<573:TCSBTB>2.0.CO;2.

Konizeski, R.L., Brietkrietz, A., McMurtrey, R.G., 1968. Geology and ground water resources of the Kalispell Valley, northwestern Montana. Mont. Bur. Mines Geol. Bull. 68 Butte.

Kovanen, D.J., Slaymaker, O., 2004. Glacial imprints of the Okanogan Lobe, southern margin of the Cordilleran Ice Sheet. J. Quat. Sci. 19, 547–565. https://doi.org/10.1002/jqs.855.

Kuehn, S.C., Froese, D.G., Carrara, P.E., Foit, F.F., Pearce, N.J.G., Rotheisler, P., 2009. Major- and trace-element characterization, expanded distribution, and a new chronology for the latest Pleistocene Glacier Peak tephras in western North America. Quat. Res. 71, 201–216. https://doi.org/10.1016/j.yqres.2008.11.003.

Lamb, M.P., Dietrich, W.E., Aciego, S.M., Depaolo, D.J., Manga, M., 2008. Formation of Box Canyon, Idaho, by megaflood: implications for seepage erosion on Earth and Mars. Science 320, 1067–1070. https://doi.org/10.1126/science.1156630.

Lamb, M.P., Mackey, B.H., Farley, K.A., 2014. Amphitheater-headed canyons formed by megaflooding at Malad Gorge, Idaho. Proc. Natl. Acad. Sci. U. S. A. 111, 57–62. https://doi.org/10.1073/pnas.1312251111.

Lapôtre, M.G.A., Lamb, M.P., Williams, R.M.E., 2016. Canyon formation contraints on the discharge of catastrophic outburst floods of Earth and Mars. J. Geophys. Res. Planets. 121, 1232–1263. https://doi.org/10.1002/2016JE005061.

Larsen, I.J., Lamb, M.P., 2016. Progressive incision of the Channeled Scablands by outburst floods. Nature 538, 229–232. https://doi.org/10.1038/nature19817.

Lesemann, J.-E., Brennand, T.A., 2009. Regional reconstruction of subglacial hydrology and glaciodynamic behaviour along the southern margin of the Cordilleran Ice Sheet in British Columbia, Canada and northern Washington State, USA. Quat. Sci. Rev. 28, 2420–2444. https://doi.org/10.1016/J.QUASCIREV.2009.04.019.

Levish, D.R., 1997. Late Pleistocene Sedimentation in glacial Lake Missoula and Revised Glacial History of the Flathead Lobe of the Cordilleran Ice Sheet, Mission Valley, Montana. University of Colorado, Boulder.

Lewis, R.S., Burmester, R.F., Breckenridge, R.M., McFaddan, M.D., Kauffman, J.D., 2002. Geologic map of the Coeur d'Alene 30 x 60 minute quadrangle. Idaho. Idaho Geol. Surv. Geol. Maps GM-33.

Lillquist, K., Lundblad, S., Barton, B.R., 2005. The Moxee (Washington) mammoth morphostratigraphic, taphonomic, and taxonomic considerations. West. N. Am. Nat. 65, 417–428.

Locke, W., 1995. Modeling of icecap glaciation of the northern Rocky Mountains of Montana. Geomorphology 14, 123–130. https://doi.org/10.1016/0169-555X(95)00053-5.

Lonn, J.D., Smith, L.N., McCulloch, R.B., 2007. Geologic Map of the Plains 30' x 60' Quadrangle, Western Montana, Montana Bureau of Mines and Geology Open-File Report Butte.

Lonn, J.D., McDonald, C., Sears, J.W., Smith, L.N., 2010. Geologic Map of the Missoula East 30' x 60' Quadrangle, Western Montana, Montana Bureau of Mines and Geology Open-File Report Butte.

Lopes, C., Mix, A.C., 2009. Pleistocene megafloods in the Northeast Pacific. Geology 37, 79–82. https://doi.org/10.1130/G25025A.1.

MacEachern, J.A., Roberts, M.C., 2013. Ichnological evidence of jökulhlaup deposit recolonization from the Touchet Beds, Mabton, WA,USA. Quat. Res. 79, 37–48. https://doi.org/10.1016/j.yqres.2012.09.001.

Malde, H.E., 1960. Evidence in the Snake River Plain, Idaho, of a catastrophic flood from Pleistocene Lake Bonneville. US Geol. Surv. Prof. Pap. 400-B. https://doi.org/10.3133/pp400B.

Malde, H.E., 1968. The catastrophic late Pleistocene Bonneville flood in the Snake River Plain, Idaho. US Geol. Surv. Prof. Pap. 596. https://doi.org/10.3133/ppP596.

Malde, H.E., Powers, H.A., 1962. Upper cenozoic stratigraphy of western Snake River plain. Idaho. Geol. Soc. Am. Bull. 73, 1197–1220. https://doi.org/10.1130/0016-7606(1962)73[1197:ucsows]2.0.co;2.

Martin, R.O.R., Hanson, R.L., 1966. Reservoirs in the United States. U.S. Geol. Surv. Water Suppl. Pap. 1838. https://doi.org/10.3133/wsp1838.

McDonald, E.V., Busacca, A.J., 1988. Record of pre-late Wisconsin giant floods in the Channeled Scabland interpreted from loess deposits. Geology 16, 728–731. https://doi.org/10.1130/0091-7613(1988)016<0728:ROPLWG>2.3.CO;2.

McDonald, E.V., Busacca, A.J., 1992. Late Quaternary stratigraphy of loess in the Channeled Scabland and Palouse regions of Washington state. Quat. Res. 38, 141–156. https://doi.org/10.1016/0033-5894(92)90052-K.

McDonald, E.V., Sweeney, M.R., Busacca, A.J., 2012. Glacial outburst floods and loess sedimentation documented during Oxygen Isotope Stage 4 on the Columbia Plateau, Washington State. Quat. Sci. Rev. 45, 18–30. https://doi.org/10.1016/j.quascirev.2012.03.016.

Medley, E., 2012. Ancient Cataclysmic Floods in the Pacific Northwest: Ancestors to the Missoula Floods. Portland State University https://doi.org/10.15760/etd.581.

Meinzer, O.E., 1918. The glacial history of the Columbia River in the Big Bend Region. Wash. Acad. Sci. J. 8, 411–412.

Meyer, S.E., 1999. Depositional History of Pre-late and late Wisconsin Outburst Flood Deposits in Northern Washington and Idaho: Analysis of Flood Paths and Provenance. Washington State University, Pullman.

Miller, D.M., 2016. The Provo shoreline of Lake Bonneville. Dev. Earth Surf. Process. 20, 127–144. https://doi.org/10.1016/B978-0-444-63590-7.00007-X.

Miller, D.M., Oviatt, C.G., McGeehin, J.P., 2013. Stratigraphy and chronology of Provo shoreline deposits and lake-level implications, late Pleistocene Lake Bonneville, eastern Great Basin, USA. Boreas 42, 342–361. https://doi.org/10.1111/j.1502-3885.2012.00297.x.

Minervini, J.M., O'Connor, J.E., Wells, R.E., 2003. Maps showing inundation depths, ice-rafted erratics, and sedimentary facies of late pleistocene missoula floods in the Williamette Valley, Oregon. U.S. Geol. Surv. Open-File Rep. 03-408. https://pubs.usgs.gov/of/2003/of03-408/.

Miyamoto, H., Itoh, K., Komatsu, G., Baker, V.R., Dohm, J.M., Tosaka, H., Sasaki, S., 2006. Numerical simulations of large-scale cataclysmic floodwater: a simple depth-averaged model and an illustrative application. Geomorphology 76, 179–192. https://doi.org/10.1016/J.GEOMORPH.2005.11.002.

Miyamoto, H., Komatsu, G., Baker, V.R., Dohm, J.M., Ito, K., Tosaka, H., 2007. Cataclysmic scabland flooding: insights from a simple depth-averaged numerical model. Environ. Model. Softw. 22, 1400–1408. https://doi.org/10.1016/j.envsoft.2006.07.006.

Moody, U.L., 1978. Microstratigraphy, Paleoecology, and Tephrochronology of the Lind Coulee Site, Central Washington Washington State University Ph.D thesis.

Moody, U.L., 1987. Late Quaternary Stratigraphy of the Channeled Scabland and Adjacent Areas University of Idaho Ph.D. thesis.

Mullineaux, D.R., Wilcox, R.E., Ebaugh, W.F., Fryxell, R., 1978. Age of the last major scabland flood of the Columbia Plateau in eastern Washington. Quat. Res. 10, 171–180. https://doi.org/10.1016/0033-5894(78)90099-6.

Newcomb, R.C., 1969. Effect of tectonic structure on the occurrence of ground water in the basalt of the Columbia River Group of The Dalles area, Oregon and Washington. US Geol. Surv. Prof. Pap. 383–C. https://doi.org/10.3133/pp383C.

Normark, W.R., Reid, J.A., 2003. Extensive deposits on the pacific plate from late pleistocene North American glacial lake outbursts. J. Geol. 111, 617–637. https://doi.org/10.1086/378334.

O'Connor, J.E., 1993. Hydrology, Hydraulics, and Geomorphology of the Bonneville Flood. Geological Society of America Special Paper 274. Boulder. https://doi.org/10.1130/SPE274-p1.

O'Connor, J.E., 2016. The Bonneville flood—a veritable Débâcle. Dev. Earth Surf. Process. 20, 105–126. https://doi.org/10.1016/B978-0-444-63590-7.00006-8.

O'Connor, J.E., Baker, V.R., 1992. Magnitudes and implications of peak discharges from glacial Lake Missoula. Bull. Geol. Soc. Am. 104, 267–279. https://doi.org/10.1130/0016-7606(1992)104<0267:MAIOPD>2.3.CO;2.

O'Connor, J.E., Burns, S.F., 2009. Cataclysms and controversy—aspects of the geomorphology of the Columbia River Gorge. Volcanoes to Vineyards: Geologic Field Trips through the Dynamic Landscape of the Pacific Northwest. Geological Society of America, pp. 237–251 https://doi.org/10.1130/2009.fld015(12).

O'Connor, J.E., Waitt, R.B., 1995. Beyond the Channeled Scabland; a field trip to Missoula flood features in the Columbia, Yakima, and Walla Walla valleys of Washington and Oregon; Part 3; Field trip, days two and three. Oregon Geol., Oregon Geol. 57, 99–115 ISSN: 0164-3304.

O'Connor, J.E., Sarna-Wojcicki, A., Wozniak, K.C., Polette, D.J., Fleck, R.J., 2001. Origin, extent, and thickness of quaternary geologic units in the Willamette Valley, Oregon. US Geol. Surv. Prof. Pap. 1620. https://doi.org/10.3133/pp1620.

O'Connor, J.E., Clague, J.J., Walder, J.S., Manville, V., Beebee, R.A., 2013. 9.25 Outburst floods. Treatise Geomorphol., 475–510 https://doi.org/10.1016/B978-0-12-374739-6.00251-7.

Othberg, K.L., Kauffman, J.D., Gillerman, V.S., Garwood, D.L., 2012. Geologic Map of the Twin Falls 30 x 60 Minute Quadrangle, Idaho Idaho Geological Survey. Geologic Map 49.

Oviatt, C.G., 2015. Chronology of Lake Bonneville, 30,000 to 10,000 yr B.P. Quat. Sci. Rev. 110, 166–171. https://doi.org/10.1016/J.QUASCIREV.2014.12.016.

Oviatt, C.G., Jewell, P.W., 2016. The Bonneville shoreline: reconsidering Gilbert's interpretation. Dev. Earth Surf. Process. 20, 88–104. https://doi.org/10.1016/B978-0-444-63590-7.00005-6.

Oviatt, C.G., Shroder, J.F., 2016. Lake Bonneville: A Scientific Update. 1st ed. Elsevier 9780444635907.

Pardee, J.T., 1910. The glacial Lake Missoula. J. Geol. 18, 376–386. https://doi.org/10.1086/621747.

Pardee, J.T., 1918. Geology and mineral deposits of the Colville Indian Reservation, Washington. U.S. Geol. Surv. Bull. 186. https://doi.org/10.3133/b677.

Pardee, J.T., 1942. Unusual currents in glacial lake Missoula, Montana. Bull. Geol. Soc. Am. 53, 1569–1600. https://doi.org/10.1130/GSAB-53-1569.

Parker, S., 1844. Journal of an Exploring Tour Beyond the Rocky Mountains: Under the Direction of the A. B. C. F. M. Containing a Description of the Geography, Geology, Climate, Production of the Country, and the Numbers, Manners, and Customs of the Natives: With a Map of Oregon Territory. 4th ed. Andrus, Woodruff, & Gauntlett, Ithaca, N.Y https://lccn.loc.gov/17015014.

Patton, P.C., Baker, V.R., 1978a. Origin of the Cheney-Palouse scabland tract. In: Baker, V.R., Nummedal, D. (Eds.), The Channeled Scabland, pp. 117–130 National Aeronautics and Space Administration Planetary Geology Program, Washington, D.C.

Patton, P.C., Baker, V.R., 1978b. New evidence for pre-Wisconsin flooding in the channeled scabland of eastern Washington. Geology 6, 567–571. https://doi.org/10.1130/0091-7613(1978)6<567:NEFPFI>2.0.CO;2.

Pederson, J.L., Janecke, S.U., Reheis, M.C., Kaufman, D.S., Oaks, R.Q., 2016. The Bear River's history and diversion: constraints, unsolved problems, and implications for the Lake Bonneville record. Dev. Earth Surf. Process. 20, 28–59. https://doi.org/10.1016/B978-0-444-63590-7.00002-0.

Perron, J.T., Venditti, J.G., 2016. Megafloods downsized. Nature 538, 174–175. https://doi.org/10.1038/538174a.

Peters, J.L., Brennand, T.A., 2020. Palaeogeographical reconstruction and hydrology of glacial Lake Purcell during MIS 2 and its potential impact on the Channeled Scabland, USA. Boreas https://doi.org/10.1111/bor.12434.

Peterson, C.D., Minor, R., Peterson, G.L., Gates, E.B., 2011. Pre-and post-Missoula flood geomorphology of the Pre-Holocene ancestral Columbia River Valley in the Portland forearc basin, Oregon and Washington, USA. Geomorphology 129, 276–293. https://doi.org/10.1016/J.GEOMORPH.2011.02.022.

Pluhar, C.J., Bjornstad, B.N., Reidel, S.P., Coe, R.S., 2006. Magnetostratigraphic evidence from the Cold Creek bar for onset of ice-age cataclysmic floods in eastern Washington during the early Pleistocene. Quat. Res. 65, 123–135. https://doi.org/10.1016/j.yqres.2005.06.011.

Porter, S.C., 1978. Glacier Peak tephra in the North Cascade Range, Washington: stratigraphy, distribution, and relationship to late-glacial events. Quat. Res. 10, 30–41. https://doi.org/10.1016/0033-5894(78)90011-X.

Praetorius, S.K., Condron, A., Mix, A.C., Walczak, M.H., McKay, J.L., Du, J., 2020. The role of northeast Pacific meltwater events in deglacial climate change. Sci. Adv. 6. https://doi.org/10.1126/sciadv.aay2915.

Prytulak, J., Vervorrt, J.D., Plank, T., Yu, C., 2006. Astoria Fan sediments, DSDP site 174, Cascadia Basin: Hf–Nd–Pb constraints on provenance and outburst flooding. Chem. Geol. 233, 276–292. https://doi.org/10.1016/j.chemgeo.2006.03.009.

Rathburn, S.L., 1993. Pleistocene cataclysmic flooding along the Big Lost River, east central Idaho. Geomorphology 8, 305–319. https://doi.org/10.1016/0169-555X(93)90026-X.

Reimer, P., et al., 2013. IntCal13 and Marine13 radiocarbon age calibration curves 0–50,000 years cal BP. Radiocarbon 55, 1869–1887. https://doi.org/10.2458/azu_js_rc.55.16947.

Richmond, G.M., Fryxell, R., Neff, G.E., Weis, P.L., 1965. The Cordilleran ice sheet of the northern Rocky Mountains, and related Quaternary history of the Columbia Plateau. In: Wright, H.E., Frey, D.G. (Eds.), The Quaternary of the U.S. Princeton University Press. ISBN: 9780691624426, pp. 231–242.

Riedel, J.L., 2017. Deglaciaton of the North Cascade Range, Washington and British Columbia, from the last glacial maximum to the Holocene. Cuad. Investig. Geogr. 43, 467–496. https://doi.org/10.1016/0169-555X(95)00053-5.

Rigby, J.G., 1982. The Sedimentology, Mineralogy, and Depositional Environment of a Sequence of Quaternary Catastrophic Flood-Derived Lacustrine Turbidites Near Spokane, Washington. University of Idaho, Moscow.

Russell, I.C., 1893. Geologic reconnoissance in central Washington. U.S. Geol. Surv. Bull. 108. https://doi.org/10.3133/b108.

Schmidt, K.L., Kauffman, J.D., Stewart, D.E., Garwood, D.L., Othberg, K.L., Lewis, R.S., 2009. Geologic map of the Grave Point quadrangle, Idaho County, Idaho, and Wallowa County, Oregon. Idaho Geol. Surv. Digit. Web Maps DWM-111, scale 1:24,000. https://www.idahogeology.org/pub/Digital_Data/Digital_Web_Maps/Lucile_DWM-126-m.pdf.

Scott, W.E., Pierce, K.L., Bradbury, J.P., Forester, R.M., 1982. Revised Quaternary stratigraphy and chronology in the American Falls area, southeastern Idaho. In: Bonnichsen, B., Breckenridge, R.M. (Eds.), Cenozoic Geology of Idaho. Idaho Bureau of Mines and Geology Bulletin 26, pp. 581–595.

Shaw, J., Munro-Stasiuk, M., Sawyer, B., Beaney, C., Lesemann, J.-E., Musacchio, A., Rains, B., Young, R.R., 1999. The channeled scabland: Back to Bretz? Geology 27, 605–608. https://doi.org/10.1130/0091-7613(1999)027.

Shaw, J., Munro-Stasiuk, M., Sawyer, B., Beaney, C., Lesemann, J.-E., Musacchio, A., Rains, B., Young, R.R., 2000. The channeled scabland: Back to Bretz?: comment and reply. Geology 28, 577. https://doi.org/10.1130/0091-7613(2000)28<577:TCSBTB>2.0.CO;2.

Smith, G.A., 1993. Missoula flood dynamics and magnitudes inferred from sedimentology of slack-water deposits on the Columbia Plateau, Washington. Bull. Geol. Soc. Am. 105, 77–100. https://doi.org/10.1130/0016-7606(1993)105<0077:MFDAMI>2.3.CO;2.

Smith, L.N., 2004. Late Pleistocene stratigraphy and implications for deglaciation and subglacial processes of the Flathead Lobe of the Cordilleran Ice Sheet, Flathead Valley, Montana, USA. Sediment. Geol. 165, 295–332. https://doi.org/10.1016/j.sedgeo.2003.11.013.

Smith, L.N., 2006. Stratigraphic evidence for multiple drainings of glacial Lake Missoula along the Clark Fork River, Montana, USA. Quat. Res. 66, 311–322. https://doi.org/10.1016/j.yqres.2006.05.009.

Smith, L.N., 2017. Repeated sedimentation and exposure of glacial Lake Missoula sediments: a lake-level history at Garden Gulch, Montana, USA. Quat. Sci. Rev. 155, 114–126. https://doi.org/10.1016/J.QUASCIREV.2016.11.018.

Smith, L.N., Hanson, M.A., 2014. Sedimentary record of glacial Lake Missoula along the Clark Fork River from deep to shallow positions in the former lakes: St. Regis to near Drummond, Montana. In: Shaw, C.A., Tikoff, B. (Eds.), Exploring the Northern Rocky Mountains. Geological Society of America, pp. 51–63. https://doi.org/10.1130/9780813700373.

Smith, L.N., Sohbati, R., Buylaert, J.-P., Lian, O.B., Murray, A., Jain, M., 2018. Timing of lake-level changes for a deep last-glacial Lake Missoula: optical dating of the Garden Gulch area, Montana, USA. Quat. Sci. Rev. 183, 23–35. https://doi.org/10.1016/j.quascirev.2018.01.009.

Smyers, N.B., Breckenridge, R.M., 2003. Glacial Lake Missoula, Clark Fork ice dam, and the floods outburst area: Northern Idaho and western Montana. In: Swanson, T.W. (Ed.), GSA Field Guide 4: Western Cordillera and Adjacent Areas. Geological Society of America, Boulder, pp. 1–15. https://doi.org/10.1130/0-8137-0004-3.1.

Spencer, P.K., 1989. A small mammal fauna from the Touchet Beds of Walla Walla County, Washington—support for the multiple flood hypothesis. Northwest Sci. 63, 167–174.

Spencer, P.K., Jaffee, M.A., 2002. Pre-Late Wisconsin glacial outburst floods in southeastern Washington—the indirect record. Wash. Geol. 30, 9–16.

Spencer, P.K., Knapp, A.N., 2010. New stratigraphic markers in the late Pleistocene Palouse loess: Novel fossil gastropods, absolute age constraints and nonaeolian facies. Sedimentology 57, 41–52. https://doi.org/10.1111/j.1365-3091.2009.01085.x.

Stearns, H.T., 1962. Evidence of Lake Bonneville flood along Snake River below King Hill, Idaho. Geol. Soc. Am. Bull. 73, 385–387. https://doi.org/10.1130/0016-7606(1962)73[385:EOLBFA]2.0.CO;2.

Steele, W.K., 1991. Paleomagnetic evidence for repeated glacial Lake Missoula floods from sediments of the Sanpoil River Valley, Northeastern Washington. Quat. Res. 35, 197–207. https://doi.org/10.1016/0033-5894(91)90067-F.

Stuiver, M., Reimer, P.J., Reimer, R.W., 2020. CALIB 7.1. [WWW program] at. http://calib.org accessed 2020-2-7.

Swanson, T.W., Caffee, M.L., 2001. Determination of ^{36}Cl production rates derived from the well-dated deglaciation surfaces of Whidbey and Fidalgo Islands, Washington. Quat. Res. 56, 366–382. https://doi.org/10.1006/qres.2001.2278.

Sweeney, M.R., Gaylord, D.R., Busacca, A.J., 2007. The evolution of eureka flat: a dust producing engine of the Palouse loess, USA. Quat. Int. 162 (163), 76–96. https://doi.org/10.1016/j.quaint.2006.10.034.

Symons, T.W., 1882. Report of an examination of the upper Columbia river and the territory in its vicinity in September and October, 1881, to determine its navigability, and adaptability to steamboat transportation.: Made by direction of the commanding general of the Department of the Columbia, U. S. 47th Cong., 1st sess. Senate. Ex. Doc. 186. Government Printing Office, Washington.

Thorson, R.M., 1989. Glacio-isostatic response of the Puget Sound area, Washington. Geol. Soc. Am. Bull. 101, 1163–1174. https://doi.org/10.1130/0016-7606(1989)101<1163:GIROTP>2.3.CO;2.

Trimble, D.E., 1963. Geology of Portland, Oregon, and adjacent areas. U.S. Geol. Surv. Bull. 1119. https://doi.org/10.3133/b1119.

Trimble, D.E., Carr, W.J., 1961. The Michaud delta and Bonneville River near Pocatello, Idaho. U.S. Geol. Surv. Prof. Paper 424-B, B164–B166. https://doi.org/10.3133/pp424B.

Vanaman, K.M., O'Connor, J.E., Riggs, N., 2006. Pleistocene history of Lake Millican, central Oregon [abs.]. Geol. Soc. Am. Abstr. Programs 38 (7), 71.

Waitt, R.B., 1977a. Missoula flood *sans* Okanogan lobe. Geol. Soc. Am. Abstr. Programs 9 (6), 770.

Waitt, R.B., 1977b. Guidebook to quaternary geology of the Columbia, Peshastin, and upper Yakima valleys, west-central Washington: U.S. Geol. Surv. Open-File Rept. 77-753.

Waitt, R.B., 1980. About forty last-glacial Lake Missoula jökulhlaups through southern Washington. J. Geol. 88, 653–679. https://doi.org/10.1086/628553.

Waitt, R.B., 1982. Surficial geology. In: Tabor, R.W., Waitt, R., Frizzell Jr., V.A., Swanson, D.A., Byerly, G.R., Bentley, R.D. (Eds.), Geologic Map of the Wenatchee 1:100,000 Quadrangle, Central Washington. U.S. Geol. Surv. Miscellaneous Investigations Map I-1311 . https://pubs.usgs.gov/imap/i1311/.

Waitt, R.B., 1984. Periodic jökulhlaups from Pleistocene glacial Lake Missoula—New evidence from varved sediment in northern Idaho and Washington. Quat. Res. 22, 46–58. https://doi.org/10.1016/0033-5894(84)90005-X.

Waitt, R.B., 1985a. Case for periodic, colossal jökulhlaups from Pleistocene glacial Lake Missoula. Bull. Geol. Soc. Am. 96, 1271–1286. https://doi.org/10.1130/0016-7606(1985)96<1271:CFPCJF>2.0.CO;2.

Waitt, R.B., 1985b. Reply to Comment on "Periodic Periodic jökulhlaups from Pleistocene glacial Lake Missoula—New evidence from varved sediment in northern Idaho and Washington". Quat. Res. 24, 357–360. https://doi.org/10.1016/0033-5894(85)90058-4.

Waitt, R.B., 1987. Erosional landscape and surficial deposits. In: Tabor, R.W., Frizzell, V.A.J., Waitt, R.B., Swanson, D.A., Byerly, G.R., Booth, D.B., Hetherington, M.J., Zartman, R.E. (Eds.), Geologic Map of the Chelan 3-Minute by 60-Minute Quadrangle, Washington. U.S. Geol. Surv. Misc. Inv. Series Map I-1661. https://doi.org/10.3133/i1661.

Waitt, R.B., 1994. Scores of gigantic, successively smaller Lake Missoula floods through Channeled scabland and Columbia valley. In: Swanson, D.A., Haugerud, R.A. (Eds.), Geological field trips in the Pacific Northwest. Geological Society of America Annual Meeting Field Trip Guidebook 1. University of Washington, Seattle, pp. K1–K88.

Waitt, R.B., 2016. Megafloods and Clovis cache at Wenatchee, Washington. Quat. Res. 85, 430–444. https://doi.org/10.1016/j.yqres.2016.02.007.

Waitt, R.B., 2017. Pleistocene glaciers, lakes, and floods in north-central Washington State. In: Haugerud, R.A., Kelsey, H.M. (Eds.), From the Puget Lowland to East of the Cascade

Range: Geologic Excursions in the Pacific Northwest. Geological Society of America Field Guide 49, pp. 175–205. https://doi.org/10.1130/2017.0049(08).
Waitt, R.B., Thorson, R.M., 1983. The Cordilleran ice sheet in Washington, Idaho, and Montana. In: Wright, H.E., Porter, S.C. (Eds.), Late-Quaternary Environments of the United States. University of Minnesota Press, Minneapolis. ISBN: 978-0816669547, pp. 53–70.
Waitt, R.B., O'Connor, J.E., Harpel, C.J., 2000. Varying routings of repeated colossal jökulhlaups through the channeled scabland of Washington, USA [abs.]. Orkustofnun Rept. OS-2000/036, Reykjav´ık, Extremes of the Extremes Conference 27.
Waitt, R.B., Breckenridge, R.M., Kiver, E.P., Stradling, D.F., 2016. Late Wisconsin Cordilleran icesheet and colossal floods in northeast Washington and north Idaho. In: Cheney, E.S. (Ed.), The Geology of Washington and Beyond from Laurentia Laurentia to Cascadia. University of Washington Press, Seattle. ISBN: 978-0295995274, pp. 233–256.
Waitt, R.B., Long, W.A., Stanton, K.M., 2019. Erratics and other evidence of late Wisconsin Missoula outburst floods in lower Wenatchee and adjacent Columbia valleys, Washington. Northwest Sci. 92, 318–337. https://doi.org/10.3955/046.092.0503.
Walder, J.S., Costa, J.E., 1996. Outburst floods from glacier-dammed lakes: the effect of mode of lake drainage on flood magnitude. Earth Surf. Process. Landf. 21, 701–723. https://doi.org/10.1002/(SICI)1096-9837(199608)21:8<701::AID-ESP615>3.0.CO;2-2.
Washington Department of Natural Resources, Division of Geology and Earth Resources, 2016. The Cheney-Palouse tract of Washington's Channeld Scalands. https://www.dnr.wa.gov/publications/ger_presentations_coe_cheney_palouse_scablands.pdf?vqt8q accessed Nov. 7, 2019.
Waters, A.C., 1933. Terraces and coulees along the Columbia River near Lake Chelan, Washington. Bull. Geol. Soc. Am. 44, 783–820. https://doi.org/10.1130/GSAB-44-783.
Weis, P.L., 1968. Geologic map of the Greenacres quadrangle, Washington and Idaho. U.S. Geol. Surv. Quadrang. Map 734, 4. https://doi.org/10.3133/gq734.
Weis, P.L., Richmond, G.M., 1965. Maximum extent of late Pleistocene cordilleran glaciation in northeastern Washington and northern Idaho. US Geol. Surv. Prof. Pap. 525–C, C128–C132. https://doi.org/10.3133/pp525C.
Welch, E.G., Brattebo, S.K., Gibbons, H.L., Plotnikoff, R.W., 2015. A dramatic recovery of Lake Spokane water quality following wastewater phosphorus reduction. Lake Reserv. Manag. 31, 157–165. https://doi.org/10.1080/10402381.2015.1040566.
Wells, R.E., Haugerud, R.A., Niem, A.R., Niem, W.A., Ma, L., Evarts, R.C., O'Connor, J.E., Madin, I.P., Sherrod, D.R., Beeson, M.H., Tolan, T.L., Wheeler, K.L., Hanson, W.B., Sawlan, M.G., 2020. Geologic map of the greater Portland metropolitan area and surrounding region, Oregon and Washington. U.S. Geol. Surv. Sci. Inv. Map 3443 (in press).
Woods, P.F., Berenbrock, C.E., 1994. Bathymetric map of Coeur D'Alene Lake, Idaho. U.S. Geol. Surv. Water-Resour. Invest. Rep. 94-4119. https://doi.org/10.3133/wri944119.
Zuffa, G.G., Normark, W.R., Serra, F., Brunner, C.A., 2000. Turbidite megabeds in an oceanic rift valley recording jokulhlaups of late Pleistocene glacial lakes of the western United States. J. Geol. 108, 253–274. https://doi.org/10.1086/314404.

Update

Earth-Science Reviews
Volume 210, Issue , November 2020, Page

DOI: https://doi.org/10.1016/j.earscirev.2020.103401

Earth-Science Reviews 210 (2020) 103401

Contents lists available at ScienceDirect

Earth-Science Reviews

journal homepage: www.elsevier.com/locate/earscirev

Corrigendum

Corrigendum to "The Missoula and Bonneville floods—A review of ice-age megafloods in the Columbia River basin" [Earth-Science Reviews volume 208 (2020) 103181]

Jim E. O'Connor[a,*], Victor R. Baker[b], Richard B. Waitt[c], Larry N. Smith[d], Charles M. Cannon[a], David L. George[c], Roger P. Denlinger[c]

[a] U.S. Geological Survey, Portland, Oregon, USA
[b] University of Arizona, Tucson, AZ, USA
[c] U.S. Geological Survey, Vancouver, Washington, USA
[d] Montana Technological University, Butte, MT, USA

The authors regret two errors:

1. In section 3.3, *Missoula flood timing synopsis*, we refer to the age of the Glacier Peak tephra as 13.7–13.5 ka. Because of a typographical error, this should be 13.7–13.4 ka, as stated elsewhere in the report.
2. Due to unfortunate modifications to the text and tables during the revision process, we failed to properly acknowledge and credit Thomas Cooney's discovery of several of the ice-rafted erratics reported in Table 3, "*Evidence of maximum Missoula flood stages.*" In particular, Mr. Cooney (mistakenly referred to as Thomas Cooley in the original version posted on-line) discovered and guided us to the four erratics listed as near "Columbia Valley near Hawk Creek" in the section of the table specifying maximum flood evidence for the "Upper Columbia River Valley; Spokane River confluence to Grand Coulee." In addition, we should have acknowledged that Mr. Cooney graciously guided us to several field sites helpful to our understanding of flood inundation and processes in the little studied Telford Crab Creek scabland tract.

The authors would like to apologise for any inconvenience caused.

DOI of original article: https://doi.org/10.1016/j.earscirev.2020.103181
* Corresponding author.
E-mail address: oconnor@usgs.gov (J.E. O'Connor).

https://doi.org/10.1016/j.earscirev.2020.103401

Available online 13 October 2020
0012-8252/ Published by Elsevier B.V.