Peak flow responses to landscape disturbances caused by the cataclysmic 1980 eruption of Mount St. Helens, Washington

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ABSTRACT

Years of discharge measurements that precede and follow the cataclysmic 1980 eruption of Mount St. Helens, Washington, provide an exceptional opportunity to examine the responses of peak flows to abrupt, widespread, devastating landscape disturbance. Multiple basins surrounding Mount St. Helens (300–1300 km^2 drainage areas) were variously disturbed by: (1) a debris avalanche that buried 60 km^2 of valley; (2) a lateral volcanic blast and associated pyroclastic flow that destroyed 550 km^2 of mature forest and blanketed the landscape with silt-capped lithic tephra; (3) debris flows that reamed riparian corridors and deposited tens to hundreds of centimeters of gravelly sand on valley floors; and (4) a Plinian tephra fall that blanketed areas proximal to the volcano with up to tens of centimeters of pumiceous silt, sand, and gravel. The spatially complex disturbances produced a variety of potentially compensating effects that interacted with and influenced hydrological responses. Changes to water transfer on hillslopes and to flow storage and routing along channels both enhanced and retarded runoff. Rapid post-eruption modifications of hillslope surface textures, adjustments of channel networks, and vegetation recovery, in conjunction with the complex nature of the eruptive impacts and strong seasonal variability in regional climate hindered a consistent or persistent shift in peak discharges. Overall, we detected a short-lived (5–10 yr) increase in the magnitudes of autumn and winter peak flows. In general, peak flows were larger, and moderate to large flows (>$Q_{2\,yr}$) were more substantively affected than predicted by early modeling efforts. Proportional increases in the magnitudes of both small and large flows in basins subject to severe channel disturbances, but not in basins subject solely to hillslope disturbances, suggest that eruption-induced modifications to flow efficiency along alluvial channels that have very mobile beds differentially affected flows of various magnitudes and likely played a prominent, and additional, role affecting the nature of the hydrological response.

Keywords: Mount St. Helens, volcano, eruption, streamflow, hydrology, landscape disturbance, hydrologic response, tephra, landslide, debris flow, watershed.

INTRODUCTION

Explosive volcanism directly impacts landscape hydrology by damaging or destroying vegetation and by depositing sediment on hillslopes and in river channels. Like other forms of landscape disturbances (e.g., forest practices, wildfire, urbanization, mining, agriculture), ecological and geomorphic impacts of explosive volcanism can alter key components of the hydrological cycle and affect the character, magnitude, timing, and duration of runoff. Streamflow responses to volcanic disturbances, however, are not well documented.

The nature and longevity of hydrological impacts caused by landscape disturbances, and methods for assessing those impacts, are subjects of debate. Scale is a chief influence on the hydrological response of any basin (Kirkby, 1988). In small basins, hillslope processes, such as infiltration, percolation, and throughflow (interflow), strongly influence hydrological response because water spends most of its time on and within a hillslope. In contrast, channel hydraulics and processes greatly affect responses of large basins as channel residence times become predominant. In small (<1 km^2) headwater basins subject to forest practices, post-disturbance discharge peaks can be double pre-disturbance peaks, and disturbance effects can persist for decades (e.g., Jones and Grant, 1996; Thomas and Megahan, 1998; Jones, 2000). In urbanized basins, flood magnitudes can increase hundreds of percent (Konrad et al., 2005). There is considerable debate, however, whether landscape disturbances influence the magnitude and timing of peak flows in large (>100 km^2) river basins in any consistent manner, or whether hydrological perturbations are buffered at that scale (e.g., Lyons and Beschta, 1983; Jones and Grant, 1996; Thomas and Megahan, 1998; Beschta et al., 2000). Principal difficulties with analyses of hydrological responses to landscape changes in large basins typically center on two themes: (1) land cover and land use commonly change more or less continuously over periods of decades, so that distinct predisturbance and postdisturbance periods cannot be well defined, and (2) there are seldom true control basins of appropriate size against which to compare disturbance effects.

In contrast to the progressive landscape changes caused by some disturbances, large, explosive volcanic eruptions can cause abrupt and widespread landscape modifications. For example, eruptions of Ilopango caldera (El Salvador) (e.g., Sheets, 2004), Katmai (Novarupta) (Alaska, USA) (Martin, 1913; Griggs, 1922), Lamington (Papua New Guinea) (Taylor, 1958), Bezymianny (Kamchatka, Russia) (Gorshkov, 1959), Mount St. Helens (Washington, USA) (Lipman and Mullineaux, 1981), and Mount Pinatubo (Philippines) (Newhall and Punongbayan, 1996) all caused instantaneous, widespread (10^2–10^3 km^2) devastation that radically altered landscape geomorphology and ecology. These eruptions varied in magnitude and intensity, but each destroyed broad swaths of vegetation and deposited from 4 to 40 km^3 of volcaniclastic sediment.

The impacts from the cataclysmic 1980 eruption of Mount St. Helens, Washington, provide an exceptional opportunity to evaluate the nature

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and longevity of streamflow change caused by drastic landscape changes in large (>300 km^2) basins. First, the eruption reconfigured hundreds of square kilometers of heavily forested landscape, deposited great quantities of sediment on hillslopes and in channels (Table 1; Lipman and Mullineaux, 1981), and radically altered hydrological processes in several watersheds (e.g., Janda et al., 1984; Collins and Dunne, 1986; Leavesley et al., 1989). Second, stations gauging discharges of water and sediment were established along several rivers soon after the eruption, and continuous posteruption data were gathered at principal stations for at least 14 yr. Third, relatively long-term pre-eruption records of discharge exist for several of the rivers where stations were reestablished. Fourth, the discharge from a comparably sized basin near the volcano, which was unaffected by the eruption, has been gauged continuously since 1930, and that record can be used as a control in paired-basin comparisons. Fifth, the variety of eruptive impacts allows comparisons of responses among basins affected by different volcanic processes.

Modeling studies conducted shortly after the eruption (Lettenmaier and Burges, 1981; Orwig and Mathison, 1982; Datta et al., 1983) concluded that runoff peaks and volumes would increase. Predicted peak discharges of posteruption unit hydrographs were 50% greater and had rise times that were ~25% faster than pre-eruption unit hydrographs (Orwig and Mathison, 1982), and predicted magnitudes of floods of given frequencies increased by 20%–60% (Lettenmaier and Burges, 1981), with changes greatest for small- to moderate-magnitude events. To date, these predictions have not been tested rigorously.

Here, we evaluate the nature, seasonality, and longevity of the hydrologic responses of large-basin (300–1300 km^2 drainage area) peak flows to the cataclysmic 1980 eruption of Mount St. Helens. We compare discharge responses to initial modeling predictions, link the responses to geomorphic, hydrologic, and ecologic changes in the affected basins, and provide new insights regarding controls of hydrologic responses related to landscape disturbances.

LANDSCAPE DISTURBANCES CAUSED BY THE 1980 ERUPTION

The eruption of Mount St. Helens on 18 May 1980 consisted of an ensemble of volcanic processes that reconfigured the landscapes of several watersheds (Lipman and Mullineaux, 1981; Table 1). Within minutes to hours of its onset, the eruption reconfigured hundreds of square kilometers through a voluminous debris avalanche, a directed volcanic blast, debris flows, pyroclastic flows, and extensive proximal tephra fall (Figs. 1–4). The nature and severity of impact in a particular basin depended upon the disturbance process and proximity to the volcano. Multiple processes impacted both hillslopes and channels in basins broadly north, east, and within 10 km of the volcano, whereas single processes chiefly impacted either hillslopes or channels in basins to the west, south, and those beyond 10 km from the volcano (Fig. 1; Tables 2 and 3).

The eruption began with a colossal failure of the volcano's north flank (Voight, 1981). The resulting debris avalanche deposited 2.5 km^3 of poorly sorted rock, soil, ice, and organic debris in the upper North Fork Toutle River valley (Figs. 1, 3, and 4A; Glicken, 1998), buried 60 km^2 of the valley to a mean depth of 45 m, truncated tributary channels, impounded lakes, and disrupted the drainage network (Lehre et al., 1983; Janda et al., 1984).

A synchronous laterally directed blast and its consequent high-mobility pyroclastic flow and associated fall followed the debris avalanche off the volcano, devastated 550 km^2 of rugged terrain in a 180° arc north of the volcano (Fig. 1), and blanketed the landscape with up to 1 m of normally graded silt-capped lithic tephra (Figs. 2, 4B, and 4C; Hoblitt et al., 1981; Waitt, 1981). Close to the volcano, the blast flow (and

TABLE 1. CHARACTERISTICS OF DEPOSITS FROM THE 18 MAY 1980 MOUNT ST. HELENS ERUPTION

Event	Volume of uncompacted deposit (km^3)	Area affected (km^2)	Deposit thickness (m)
Debris avalanche	2.5	60	10–195
Blast	0.20	550	0.01–1
Debris flows	0.05	50	0.1–3
Pyroclastic flows	0.3	15	0.25–40
Proximal tephra fall	0.1	1100	>0.01

Note: Data are from Lipman and Mullineaux (1981).

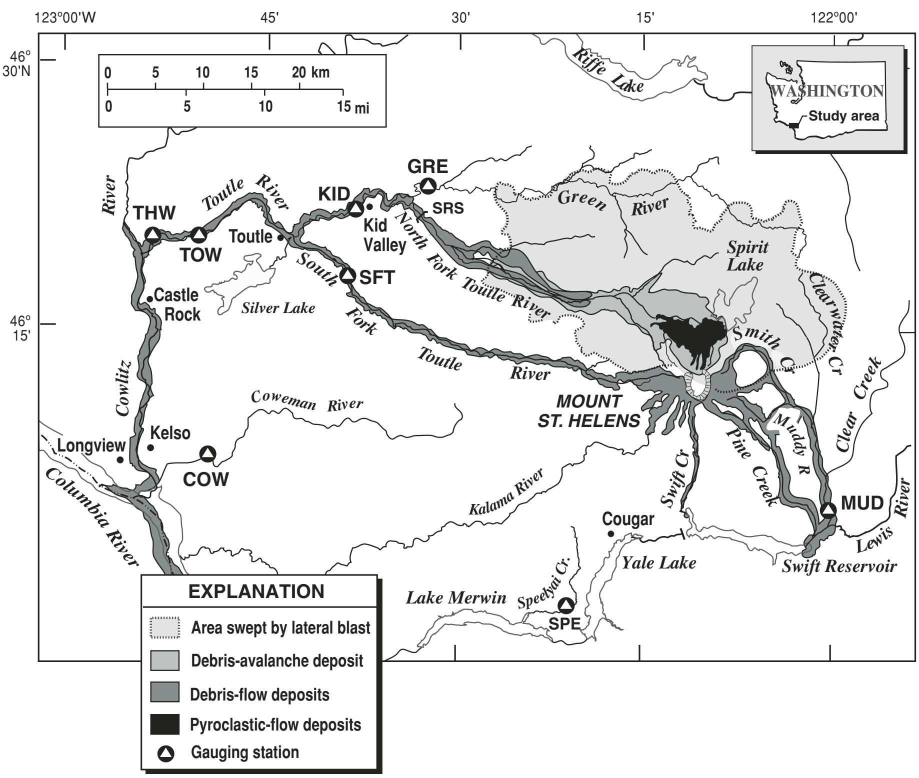


Figure 1. Distribution of major volcaniclastic deposits of the 1980 Mount St. Helens eruption and location of gauging stations (e.g., Toutle River, TOW). SRS identifies a sediment retention structure. Gauges for Tilton River (TIL), Cispus River (CIS), and East Fork Lewis River (EFK) are located 60 km northwest, 40 km northeast, and 45 km southwest of Mount St. Helens, respectively (see notes in Table 2).

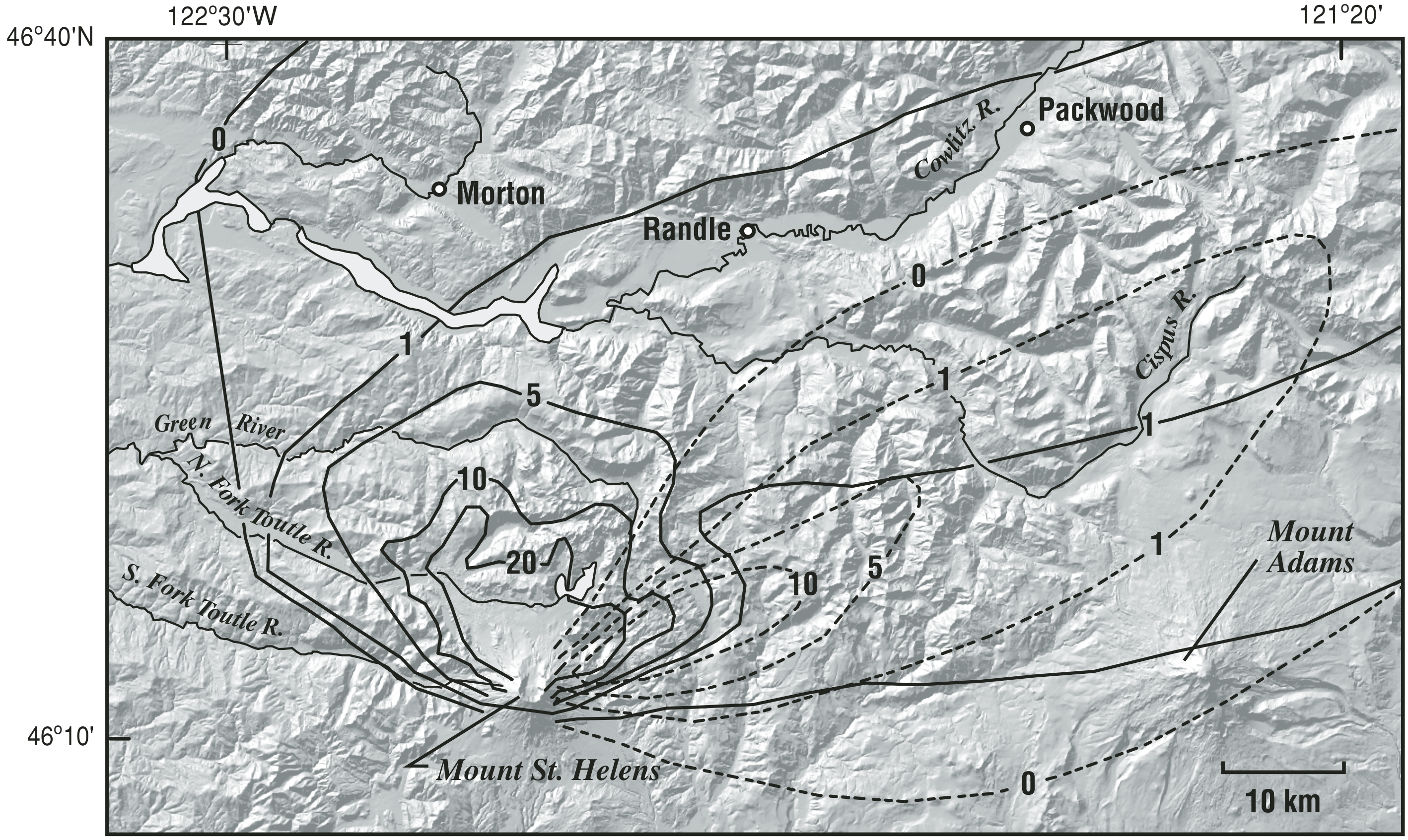


Figure 2. Isopach maps of 18 May 1980 blast deposit, including fall facies (solid lines), and proximal Plinian tephra fall (dashed lines); values are in centimeters. Map was modified from Waitt and Dzurisin (1981).

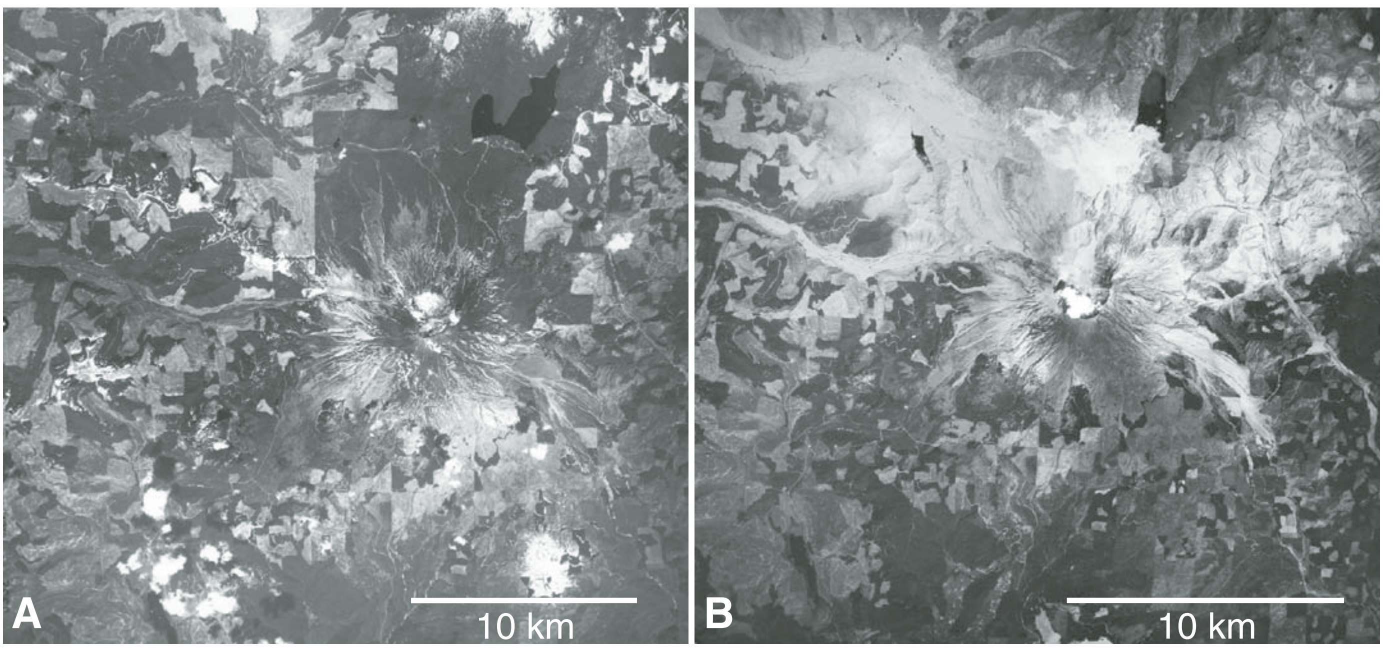


Figure 3. High-altitude vertical aerial photographs of Mount St. Helens. Note Spirit Lake in the upper-right quadrant of each photograph for reference. (A) 1975. (B) 19 June 1980.

debris avalanche) stripped vegetation and soil from the landscape. Locally, blast deposits on slopes remobilized spontaneously and generated secondary pyroclastic flows that deposited channel fill as thick as 10 m (Hoblitt et al., 1981; Brantley and Waitt, 1988). With increasing distance from the volcano, the blast flow toppled but did not remove trees (Fig. 4C; cf. Plate 1 in Lipman and Mullineaux, 1981). In basins of the Green River and upper Clearwater Creek (Fig. 1), the blast flow devastated hillslopes, but had relatively little impact on stream channels aside from locally toppling mature trees into channels (e.g., Lisle, 1995).

Extensive debris flows swept all major channels draining the volcano and deposited tens to hundreds of centimeters of gravelly sand on valley floors and floodplains. The destructive (10^8 m^3) North Fork Toutle River debris flow (Janda et al., 1981; Fairchild, 1987) traveled at least 100 km along the North Fork Toutle, Toutle, and Cowlitz Rivers (Fig. 1). On the volcano's western, southern, and eastern flanks, large but less voluminous (to 10^7 m^3) debris flows traveled up to tens of kilometers (Janda et al., 1981; Fink et al., 1981; Pierson, 1985; Major and Voight, 1986; Fairchild, 1987; Scott, 1988; Waitt, 1989). Notably large flows swept the channels of the South Fork Toutle and Muddy Rivers (Fig. 1). Overall, the debris flows reamed riparian corridors, straightened and smoothed river channels (Fig. 4F), and transformed them from sinuous, gravel-bedded, pool-riffle systems to streamlined, sand-bedded systems (e.g., Janda et al., 1981, 1984; Pierson, 1985; Scott, 1988; Major et al., 2005; Swanson et al., 2005).

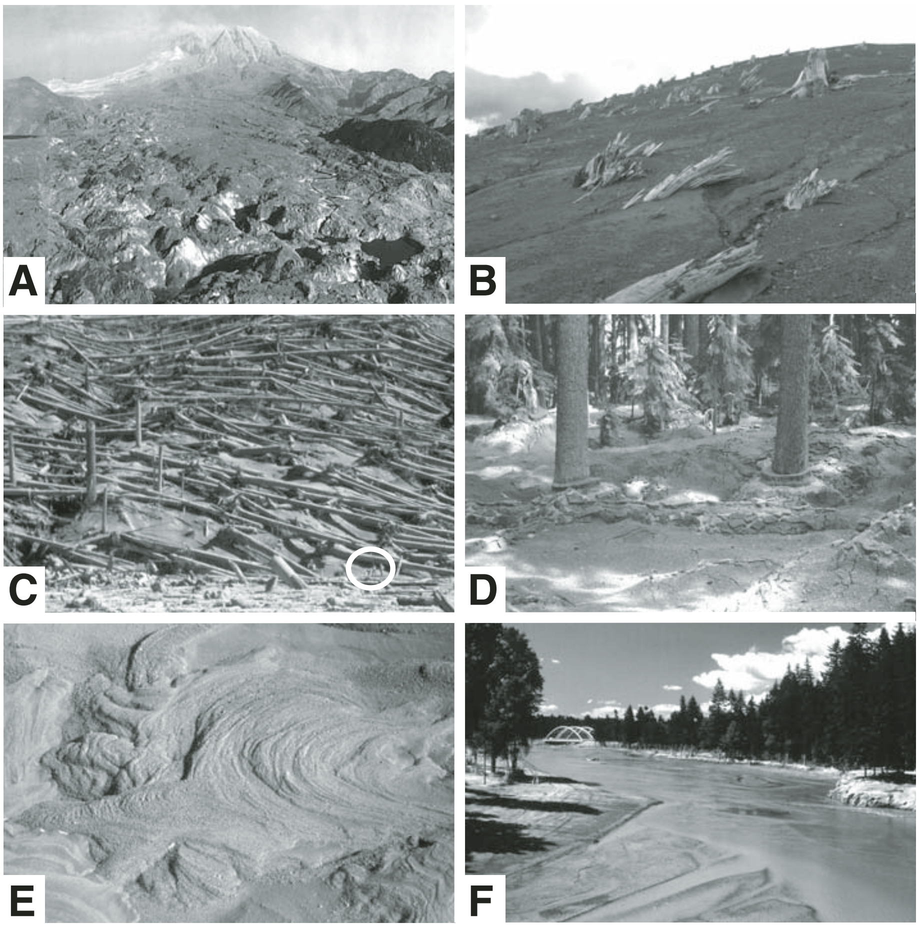


Figure 4. Landscape disturbances caused by the 1980 Mount St. Helens eruption. (A) Hummocky debris-avalanche deposit in upper North Fork Toutle River valley, June 1980. Field of view is ~4 km across. Photograph by H. Glicken. (B) Tephra-mantled hillslope within blow-down zone of lateral blast (cf. Fig. 1). Stumps are ~0.5 m wide. Photograph by U.S. Geological Survey (USGS), 1980. (C) Tephra-mantled hillslope within blow-down zone of lateral blast and thick valley fill in upper Smith Creek (cf. Fig. 1). Note person in circle for scale. Photograph by Lyn Topinka, USGS, 24 September 1980. (D) Tephra-mantled landscape east of volcano beyond zone of lateral blast, summer 1980. The tree trunks are ~0.5 m across. Photograph courtesy of Joe Antos, University of Victoria. (E) Pyroclastic flow deposits mantling debris-avalanche deposit in upper North Fork Toutle River valley. Individual flow lobes are ~30–50 m across. Photograph by Lyn Topinka, USGS, 30 September 1980. (F) Lower Toutle River valley, transformed from cobble-bedded channel to smooth, sand-bedded channel by large debris flow. Field of view is ~80 m across. Photograph by Lyn Topinka, USGS, 6 July 1980.

Fall from a billowing (Plinian) eruption column, which developed shortly after the onset of the eruption, blanketed proximal areas east-northeast of the volcano with gravelly to silty pumice fall as thick as tens of centimeters (Waitt and Dzurisin, 1981; Fig. 2); it also generated pumiceous pyroclastic flows that accumulated locally on the surface of the debris-avalanche deposit (Figs. 1 and 4E). Close to the volcano, tephra fall and pyroclastic flows augmented deposition on an already devastated landscape, but beyond 15 km east of the volcano, accumulations of tephra fall caused the primary disturbances in many watersheds (e.g., Cispus River basin; Tables 2 and 3; Figs. 2 and 4D; Sarna-Wojcicki et al., 1981; Waitt and Dzurisin, 1981). Tephra fall greater than ~5 cm thick significantly damaged forest understory (Antos and Zobel, 2005).

Deposits from several smaller eruptions in 1980 augmented the disturbances caused by the 18 May 1980 eruption. Eruptions from May to October 1980 deposited thick pyroclastic fill on the surface of the debris-avalanche deposit and veneers of tephra in neighboring watersheds (Rowley et al., 1981; Sarna-Wojcicki et al., 1981). However, these eruptions deposited sediment prior to the onset of the wet season. Thus, from a hydrological perspective, we consider the multiple 1980 eruptions as a single event. Minor eruptions from 1980 through 1986 caused a few snowmelt-induced debris flows and sediment-laden floods (Waitt et al., 1983; Pierson and Waitt, 1999; Pringle and Cameron, 1999), but those events had little significant landscape impact.

HYDROLOGICAL IMPACTS OF THE VOLCANIC DISTURBANCES

Explosive volcanism at Mount St. Helens caused ecological and geophysical perturbations that radically altered the landscape hydrology. The 1980 eruption destroyed mature forest over hundreds of square kilometers, broadly deposited tephra having a nearly impervious surface

TABLE 2. DRAINAGE BASIN CHARACTERISTICS FOR GAUGING STATIONS AT MOUNT ST. HELENS

Station	Station number	Record period (Water year)†	Basin area (km^2)‡	Primary disturbance zone§	Disturbance#	Basin area disturbed (%)
Disturbed basins						
Toutle River (TOW)††	14242580	1930–present	1285	C	DA, B, DF, T, PF	37
North Fork Toutle (KID)	14240500	1931–1933	320			
	14241100	1981–1994	735	H,C	DA, B, DF, T, PF	60
South Fork Toutle (SFT)‡‡	14241500	1940–1957	310			
		1981–present	300/310	C,(H)	B, DF	11
Green River (GRE)	14241000	1947–1950	340			
		1981–1994	335	H	B, T	53
Muddy River (MUD)§§	14216500	1950–1973	340			
		1982–present	350	H,C	B, DF, T	67
Speelyai Creek (SPE)	14219800	1960–present	33	H	T	100
Cispus River (CIS)	14232500	1930–1996	830	H	T	87
	14231900	1997–present	650			
Undisturbed basins						
Coweman River (COW)##	14245000	1951–1984	308		Undisturbed	0
Tilton River (TIL)†††	14236200	1942–present	365		Undisturbed	0
Control basin						
East Fork Lewis (EFK)	14222500	1930–present	325		Undisturbed	0

†Of the gauges currently in operation, only the record through water year 2000 was examined.
‡Upstream of gauge location.
§C—channel; H—hillslope.
#DA—debris avalanche; B—directed blast; DF—debris flow; T—tephra fall; PF—pyroclastic flow.
††Prior to 1980, the station was located near Silver Lake and had a drainage area of 1230 km^2. Data collection at TOW began in March 1981. In this analysis, we combine data collected in water years 1980 and 1981 at Highway 99 bridge, a temporary station (THW) located 9 km farther downstream, which has a drainage area of 1325 km^2, with that of TOW. There are no significant tributary inputs to the Toutle River between TOW and THW; thus, discharges measured at these two stations are considered to be equivalent. Posteruption drainage areas changed with time, however, as various lakes in the North Fork Toutle River valley eventually discharged water. Until July 1981, 103 km^2 of the upstream drainage area was noncontributing; from July 1981 to November 1982, 54 km^2 was noncontributing. After November 1982, the drainage area shown contributed to discharge.
‡‡The posteruption gauge for SFT was moved 4 km downstream following flood damage in 1996. The 1981–1996 gauge had a contributing area of 300 km^2; the new gauge has a contributing area of 310 km^2.
§§The posteruption gauge station was moved from above Clear Creek to below Clear Creek in 1984, which added an additional 134 km^2 of drainage area. The basin area shown is for the gauge below Clear Creek.
##This basin, west of Mount St. Helens, was unaffected by the eruption, but its short posteruption period of record precluded using it as the control basin. We used its record as a check on the veracity of an eruption-induced signal detected in the disturbed basins.
†††This basin was also unaffected by the 1980 eruption. The gauge is located 60 km northwest of Mount St. Helens, and we used its record as another check on the veracity of an eruption-induced signal detected in the disturbed basins.

TABLE 3. HYDROLOGICAL EFFECTS OF VOLCANIC DISTURBANCE PROCESSES

Disturbance process	Primary disturbance zone†	Basin	Basin area disturbed (%)	Deposit Thickness (m)	Hydrological effect
Debris avalanche	C	NF Toutle	15§	10–195	Channel blockage; temporary impoundment of surface/subsurface flow
Blast pyroclastic flow (vegetation damage; local removal of soil; tephra deposition by blast cloud)	H (C)	NF Toutle SF Toutle Green Muddy	59§ 6 53 28	0.5–1 0.01–0.2 0.01–0.5 0.01–1	Increased throughfall; reduced infiltration; reduced evapotranspiration; transient expansion of drainage density due to greater precipitation throughfall and increased runoff; rapid stabilization of drainage density within 2 yr of eruption; local channel aggradation
Debris flows	C	NF Toutle SF Toutle Muddy Toutle	2§ 5 5 3	0.1–3	Channel aggradation; hydraulic smoothing; simplification of channel form and structure
Pumiceous pyroclastic flows	C	NF Toutle	2§	0.25–40	Little additional impact beyond that of debris avalanche
Plinian tephra fall (>1 cm thick)	H	Muddy Cispus Speelyai	62 87# 100‡	0.2 ≤0.1 0.005	Reduced infiltration; transient expansion of drainage density due to greater precipitation throughfall and increased runoff; rapid stabilization of drainage density within 2 yr of eruption

Note: See Table 2 for basin characteristics.
†C—main channel; H—hillslope and headwater areas.
§Percentages are normalized to the area of only the North Fork Toutle River valley (~400 km^2). The basin area shown in Table 2 is the drainage area above the gauging station, which includes the North Fork Toutle valley and the Green River valley.
#Includes deposit of 18 May 1980 blast fall found beyond the boundary of the 1 cm Plinian fall isopach.
‡Affected by fall from eruptions on 25 May 1980 (0.003 m) and 12 June 1980 (0.002 m).

over more than 1000 km^2, and greatly altered the character of major channels that drained the volcano (Tables 1 and 3; Figs. 1 and 4). Infiltration capacities of slopes ravaged by the lateral blast were reduced from ~75–100 mm h^{-1}, typical of forested soils in the Cascade Range (Johnson and Beschta, 1980; Leavesley et al., 1989), to as little as 2 mm h^{-1} (Fiksdal, 1981; Leavesley et al., 1989). One year after the eruption, spatially averaged and plot-specific infiltration capacities within this disturbance zone remained <10 mm h^{-1} (Fiksdal, 1981; Swanson et al., 1983; Leavesley et al., 1989). After nearly 20 yr, plot-specific infiltration capacities remain 3–5 times lower than predisturbance capacities (Major and Yamakoshi, 2005). Infiltration capacities of proximal tephra-covered forest lands beyond the blast-affected area were not measured, but likely were significantly reduced. Antos and Zobel (1997, 2005) noted that even though tephra that fell through live canopy mixed with tree needles and became coated with litter, it had an impervious surface crust. Infiltration rates of 5–15 mm h^{-1} characterized fine distal tephra fall on agricultural land in eastern Washington for months after the eruption (e.g., Cook et al., 1981).

The volcanic impacts modified the typical modes of landscape water transfer (Fig. 5) and altered hillslope hydrology in the most heavily affected basins. Normally, hillslope storage and subsurface flow are the dominant components of forest hydrology in the Pacific Northwest. Vegetation loss and greatly reduced infiltration radically modified the amount of precipitation reaching the surface, the evaporative and infiltration losses, hillslope storage, subsurface flow, and the dynamics of snow accumulation and melt, which directed substantially more rainfall and snow melt to overland flow (cf. Fig. 5). However, enhanced depression storage owing to accumulations of downed trees and tephra-surface irregularities partly counteracted landscape changes that enhanced runoff.

Channel changes had variable hydrological impacts. Straightening and smoothing of channels by debris flows enhanced flow efficiency by reducing hydraulic roughness. In contrast, disruption of the upper North Fork Toutle River valley by the debris-avalanche deposit temporarily diminished channel flow. The debris-avalanche deposit blocked several channels tributary to the North Fork Toutle River (Janda et al., 1984), and because of its irregular surface of hummocks and closed depressions (Glicken, 1998), it disrupted through-going flow. Drainage development on the debris-avalanche deposit began shortly after emplacement when ponds that formed in depressions on the deposit breached (Janda et al., 1984). Channel development was augmented in several ways: by breakouts of lakes impounded adjacent to the avalanche deposit, by controlled releases from the largest lakes impounded along the deposit margin, by pumping water from Spirit Lake (cf. Fig. 1) across the deposit surface, by meltwater floods and debris flows issuing from the crater, and by runoff erosion (Dunne and Leopold, 1981; Rosenfeld and Beach, 1983; Janda et al., 1984; Paine et al., 1987; Simon, 1999). Lakes and ponds that formed adjacent to and on the surface of the deposit trapped and slowly released local runoff. It took nearly 3 yr to fully integrate a new drainage network across

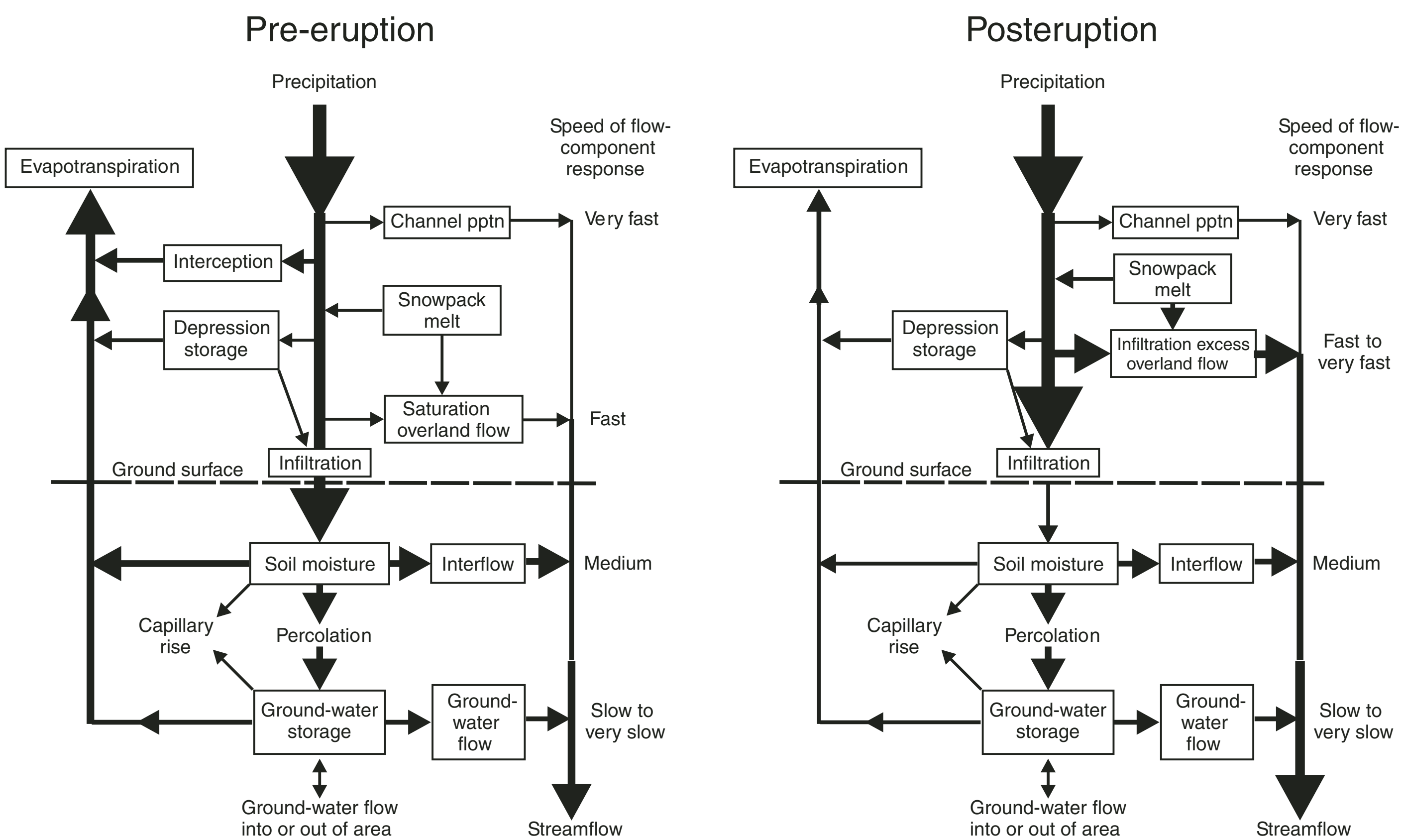


Figure 5. Schematic depiction of chief land phases of water transfer before and after the 1980 eruption. Width of arrows depicts relative magnitude of water transfer. In the posteruption diagram, note the loss of canopy interception, the greater amount of precipitation reaching the ground surface, reduced evapotranspiration, the reduction of surface infiltration, and more prominent overland flow. Also note the hypothesized changes in subsurface water transfer (adapted from Solomon and Cordery, 1984).

the deposit (Meyer, 1995; Simon, 1999), and much of that integration was accomplished by artificial means (Janda et al., 1984). Obliteration of the drainage network in the upper North Fork Toutle River valley partly counteracted other landscape changes that enhanced surface runoff. As of 2006, some lakes and ponds continue to trap runoff.

CLIMATE SETTING

The Mount St. Helens area is characterized by cool, wet winters and warm, dry summers (Fig. 6). Mean annual precipitation at Spirit Lake (elevation 977 m) (Fig. 2) was 2092 mm from 1933 through 1944 (U.S. Weather Bureau, 1932–1946), most of which fell from September through May (Fig. 6A). At a lower elevation (200 m) south of the volcano, mean annual precipitation at Peterson's Ranch near Cougar from 1953 through 2004 was 2923 mm (Washington Climate Summaries, 2004). Below ~600 m elevation, precipitation falls mainly as rain. Seasonal snowpack accumulates above 1000 m elevation, and above 1200 m elevation, snowpack of more than 3 m is common and can persist into July (Fig. 6B; Swanson et al., 2005). Elevations between 200 m and 1000 m are within the transient snow zone and are subject to rain-on-snow events that can trigger major flooding (Harr, 1981; Marks et al., 1998). Much of the proximally disturbed landscape, exclusive of channels, lies above 750 m elevation. Mean temperature at Spirit Lake in January is about −1.5 °C and in July is ~15 °C; at Peterson's Ranch, respective mean temperatures are 3.2 °C and 18.8 °C (U.S. Weather Bureau, 1932–1946; Washington Climate Summaries, 2004; Fig. 6A). Soils are wettest from December through May and driest in August and September (Fig. 6C). The seasonal hydrograph (Fig. 6D–E) is driven largely by prolonged, low-intensity autumn and winter rainfall augmented by spring melt of high-elevation snowpack; the greatest discharges occur when rains fall on thick snowpacks.

Climate patterns in the Pacific Northwest produced cyclic decadal-scale periods of wetter- and drier-than-average conditions for much of the twentieth century (Mantua et al., 1997; Biondi et al., 2001; Fig. 7A). Higher-frequency fluctuations, driven chiefly by El Niño–Southern Oscillation, punctuated each multidecadal period (e.g., McCabe and Dettinger, 1999). Relatively dry conditions characteristic of the 1980s were followed in the mid to late 1990s by wetter-than-average conditions (Fig. 7A). From 1995 to 2000, mean annual streamflow from basins near Mount St. Helens was ~40% to 50% greater than that during the period 1980–1994 (Fig. 7B). The long-term cyclic hydrologic patterns and shorter-term, higher-frequency fluctuations affected flows from the disturbed basins as well as the control basin. The paired-basin analysis used in this study minimized the influences of these climate patterns on our comparative flow analyses.

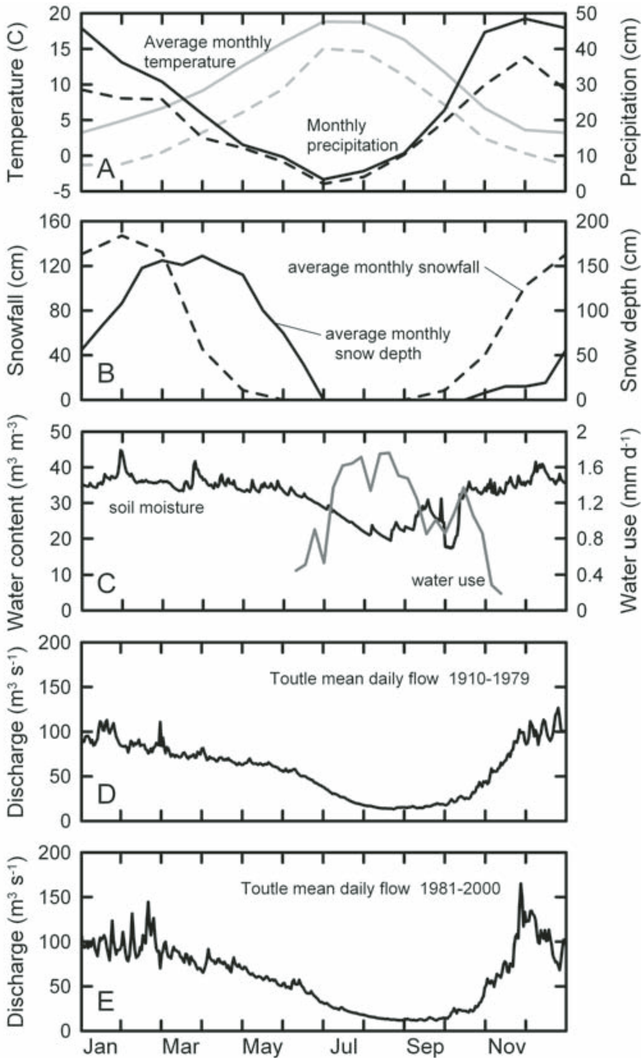


Figure 6. Regional climate and hydrology near Mount St. Helens. (A) Average monthly precipitation and temperature at Spirit Lake (elev. 977 m; 1932–1946; dashed lines) and Peterson's Ranch (elev. 200 m; 1953–2004; solid lines). (B) Average monthly snowfall and snow depth at Spirit Lake (1932–1946). (C) Average daily soil moisture in upper 0.3 m of soil profile in old-growth forest in Mount St. Helens region (2003–2005) (Wind River Canopy Crane Research Facility, 2005), and total water use from upper 2 m of soil profile, which serves as a proxy for evapotranspiration (from Warren et al., 2005). (D) Seasonal distribution of preeruption mean daily discharge along the lower Toutle River (1910–1979). (E) Seasonal distribution of posteruption mean daily discharge along the lower Toutle River (1981–2000).

HYPOTHESIZED HYDROLOGICAL RESPONSES TO THE 1980 ERUPTION

We hypothesize that ecological and geophysical impacts of the 1980 eruption caused distinct seasonal responses of peak flows at Mount St. Helens. In autumn (October–December)[1], soil moisture and air temperature are optimal for transpiration by conifers (Jones, 2000), vegetation water use is moderate (but declining), precipitation is plentiful, and snow accumulation generally is modest (Fig. 6A–C). Vegetation loss

[1]We define seasons by dividing the water year, which runs from 1 October through 30 September, into equal quarters. All of our temporal data refer to water years.

caused by the 1980 eruption increased precipitation throughfall, modified snow accumulation, distribution, and melt, and reduced seasonal transpiration. These factors alone can result in increased autumn runoff magnitudes and volumes (e.g., Jones, 2000; cf. Fig. 5). In addition, the meager infiltration capacity of the mantle of tephra fall led to atypically abundant overland flow from hillslopes that had predisturbance substrates characterized by heavily forested, highly permeable gravelly and sandy loams (e.g., Collins and Dunne, 1986). Enhanced overland flow would augment the influences of reduced transpiration and increased throughfall, and direct greater quantities of water to channels more rapidly than usual. Enhanced depression storage owing to accumulations of downed trees and tephra-surface irregularities could have modulated surface runoff, however. Nevertheless, we expect autumn storm flows to exhibit flashier hydrographs and have enhanced peaks that rise more rapidly.

Winter (January–March) is characterized by abundant precipitation, substantial snow accumulation at high elevations, and limited transpiration by vegetation (Fig. 6A–C). During winter, soils are at their wettest and typically have little capacity to store water. Thus, loss of transpiration should have a small effect on winter runoff, but enhanced throughfall and overland flow (within the transient snow zone) would augment seasonally high runoff. Modified snow accumulation, distribution, and melt resulting from vegetation loss would likely affect snowpack moisture storage and influence runoff. Such modifications would especially affect the dynamics of water release during rain-on-snow events. We expect winter storm flows to exhibit enhanced and more rapidly rising peaks, but to a lesser extent than expected for autumn flows.

In spring (April–June), air temperature is conducive to transpiration by conifers, precipitation is modest and declining, and accumulated snow is melting (Fig. 6A–C). However, because soils typically are wet (Fig. 6C), the amount of moisture transpired is small relative to the amount in storage (Jones, 2000). Thus, reduced transpiration owing to vegetation loss has less effect on spring runoff than autumn runoff. Reduced infiltration and accelerated snow melt would enhance overland flow, which would enhance the magnitude and rise of spring flows. We expect spring flows to exhibit peaks that are enhanced, but to a lesser extent than autumn flows.

Summer (July–September) peak flows are associated primarily with localized storm cells or transient large storms. Although transpiration is normally at its seasonal peak in summer (Fig. 6C), reduced transpiration owing to vegetation loss likely has little effect on runoff, owing to the normally low soil moisture (Fig. 6C). Lack of recharge during winter and spring owing to more abundant overland flow would further depress soil moisture and reduce the subsurface flow contribution to summer runoff. Enhanced overland flow across tephra deposits may enhance small flows, but in general, summer peak flows probably would be the least affected by the ecological and geophysical changes caused by the eruption.

Eruption-induced channel perturbations as well as modifications to hillslope hydrology likely affected peak flows. Denlinger et al. (2001) showed that channel form and structure are primary factors that affect flood stage, velocity, and power. Modifications of channel form and resistance at Mount St. Helens at least temporarily altered peak flow movement along disturbed channels. In several channels, diminished sinuosity, removal of large woody debris, and aggradation modified channel form and structure (loss of channel bars, pools, riffles; Fig. 4F), which reduced channel form, bed form, and grain (skin friction) roughness. Drastically reduced roughness, combined with abnormally high sediment concentrations (Dinehart et al., 1981; Dinehart, 1985; Major, 2004), likely permitted peak flows to travel rapidly with little attenuation.

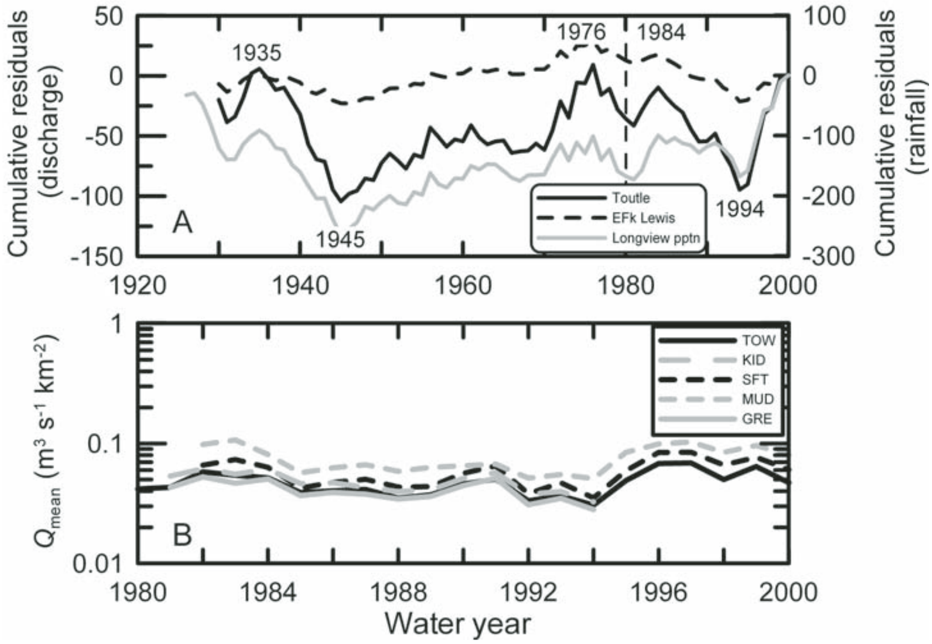


Figure 7. (A) Time series of cumulative sums of annual deviations from long-term mean flow of the lower Toutle River near Silver Lake and East Fork Lewis River near Heisson, and from long-term precipitation at Longview (cf. Fig. 1; after Hurst, 1951). The vertical dashed line indicates 1980 eruption of Mount St. Helens. A downward trend denotes periods of drier-than-average conditions; an upward trend denotes periods of wetter-than-average conditions. (B) Time series of normalized posteruption mean annual discharges for rivers at Mount St. Helens (from Major, 2004). See Figure 1 for gauging station locations (e.g., TOW).

METHODOLOGY

We used linear regression analyses of paired-basin responses to evaluate the styles, magnitudes, and durations of peak flow responses to the 1980 eruption. Paired-basin comparisons, in which one basin serves as a control, are used to evaluate hydrological responses to disturbance because they eliminate influences of climate trend on hydrological signals (cf. Fig. 7A–B). Paired basins in proximity are assumed to receive consistently distributed seasonal precipitation and to be affected equally by long-term increases or decreases in precipitation. Unlike areally restricted convective storm cells that commonly affect more continental climates in the United States, broad frontal systems affecting tens to hundreds of thousands of square kilometers are typical of the maritime climate of the Pacific Northwest during the wet season, when most peak flows occur.

Paired-basin analyses require a control basin to which the responses of altered basins can be compared. An ideal control basin is unaffected by deliberate anthropogenic impacts or other disturbances that might disproportionately influence runoff; it is a basin in which runoff is governed solely by hydrological processes in an undisturbed state. Control basins in experimental forests are preserved for this purpose, but they commonly have small drainage areas (~0.1–1 km^2) (e.g., Swank and Crossley, 1988; Jones and Grant, 1996; Hornbeck et al., 1997; Jones and Swanson, 2001). Few large basins are undisturbed, which complicates their utility as controls.

An ideal control basin near Mount St. Helens does not exist, but a reasonable proxy is the East Fork Lewis River basin, south of Mount St. Helens. The basin centroid is ~45 km southwest of the volcano, and the river flows unregulated from its headwaters to its outlet. The basin has a drainage area of ~560 km^2; ~325 km^2 lie upstream of the gauging station (EFK) near Heisson, Washington, which has operated continuously since 1930 (Table 2). The basin was unaffected by the 18 May 1980 eruption, but was dusted by trivial amounts of tephra fall (<0.002 m of fine to very fine sand) during subsequent eruptions in late May and June 1980 (Waitt et al., 1981). The basin below the gauging station is composed of rural residential and agricultural areas. Upstream, however, it lies within the heavily forested foothills of the Cascade Range; <2% of that area has been developed, and large proportions of the headwater basin lie within roadless areas. Development upstream of the gauging station centers chiefly around the small rural town of Yacolt, Washington (as of 2000, the town had an area of 1.3 km^2 and a population of 1055; U.S. Census Bureau, 2000).

Although development above the gauging station has been minimal, the basin has been affected by natural and anthropogenic disturbances. A massive wildfire in 1902, the Yacolt Burn, scorched ~1000 km^2 in the Cascade Range foothills (e.g., Colee, 1942). About 70% of the East Fork Lewis River basin above Heisson was burned. Severe stand-replacing fires also burned appreciable areas within the basin in 1927 (Rock Creek Fire), 1929 (Dole Fire), and 1952 (U.S. Forest Service, 1995). Hydrologic impacts of wildfires, even severe wildfires, are typically short-lived, however (e.g., Moody and Martin, 2001; Pierson et al., 2001; Cannon and Gartner, 2005), and by the time the gauging station started operating continuously in 1930, much of the hydrologic impact of the Yacolt, Rock Creek, and Dole burns had probably diminished.

Temporal changes in anthropogenic disturbance are more difficult to identify and quantify. Although the upper watershed above Heisson is generally undeveloped, it contains large public and private holdings actively managed for timber production (U.S. Forest Service, 1995). Recent analyses of Landsat imagery show, however, that there has been minimal stand replacement in the basin—either by logging or fire—from 1972 through 2000 (Sean Healy, 2005, personal commun.); aside from minor patches of stand-replacement harvests, land cover across the bulk of the basin above the gauge remained unchanged over that period. Runoff variations in the East Fork Lewis River basin owing to natural and anthropogenic disturbances contribute to the scatter in our paired-data comparisons and could have modulated response signals associated with the Mount St. Helens eruption, but overall the basin provided a robust control.

Hydrographs of substantive peak flows were extracted manually from original strip charts or digital records of river stage from 463 gauge-years of record (Table 2). To qualify as a substantive peak, the start of the hydrograph had to display a minimum rate of stage change of 1.5 cm h^{-1}, and the stage had to rise by a minimum of 10 cm. These criteria allowed us to pick distinct events and eliminate gradual fluctuations in seasonal base flow. If a storm generated a compound hydrograph, we extracted each peak if it met these criteria, otherwise we extracted only the maximum peak. Our descriptive parameters of a hydrograph included starting stage, peak stage, and the times of the start and peak of the hydrograph. Owing to complexities of definition, we did not define an end to a hydrograph. Therefore, we did not determine volumes of storm runoff, nor could we separate the relative contributions of storm flow and base flow. Peak stages meeting the screening criteria were converted to discharge using stage-discharge rating curves, with appropriate rating shifts, if noted.

Peak flows from the East Fork Lewis River were matched against those from nine gauges in regional basins near Mount St. Helens (Fig. 1; Tables 2 and 3), including five gauges from basins severely disturbed by the eruption (lower Toutle River, North and South Forks of the Toutle River, Green River, Muddy River), two basins affected by minor to moderate tephra fall (0.005–0.10 m) (Speelyai Creek, Cispus River) (Waitt and Dzurisin, 1981; Waitt et al., 1981), and two basins essentially unaffected by the eruption (Coweman River, Tilton River). The Coweman and Tilton River basins locally received trivial dustings of tephra fall from eruptions in the summer of 1980 (Coweman: <0.002–0.005 m, 25 May; Tilton: <0.002 m, 7 August) (Sarna-Wojcicki et al., 1981; Waitt et al., 1981). We matched discharges from the Coweman and Tilton Rivers with those of the East Fork Lewis River (i.e., matches between undisturbed basins) as a way to check the veracity of any eruption-induced signals detected in the disturbed basins. Matched peaks were assumed to register the same storm, and we considered discharges matched if they peaked within 12 h of each other. Overall, 68% (±8%) of the matched discharges peaked within 4 h, and 82% (±4%) peaked within 6 h, of each other.

We performed simple linear regression analyses of logarithms of unit-area peak discharges derived from matched seasonal discharges divided into pre- and posteruption periods. Most gauges used in this study had relatively long pre-eruption records (18–50 yr); however, two gauges (North Fork Toutle River and Green River) had only short (4 yr) pre-eruption records (Table 2). Posteruption data were subdivided into ~5 yr intervals to examine potential temporal evolution of peak flows. This interval was long enough to include a sufficient population of events to provide a meaningful analysis, yet short enough to examine fine-scale temporal variations. Unit-area discharges were used to remove the effects of variations in station locations over time (Table 2). Natural logarithms of unit-area discharges more closely met regression assumptions about normality and variance than did untransformed values (e.g., Helsel and Hirsch, 1992). We tested our data for normality and constant variance and found most temporal groupings met those criteria. For those that did not, we tried other transformations. We found, however, that logarithmic transformation provided the most uniformly distributed residuals. For a few groupings, we had to accept that discharges were not distributed normally, or that variances were not constant.

We compared pre- and posteruption regression models and tested the null hypothesis that they were coincident versus an alternative hypothesis that they were not coincident. We used a nested F-test to compare the regression models (Weisberg, 1985; Helsel and Hirsch, 1992), and applied Bonferroni's multiple comparison procedure (e.g., Thomas and Megahan, 1998) to maintain an overall significance level of $\alpha = 0.05$. Bonferroni's adjustment is a very conservative way of maintaining an overall significance level when conducting multiple comparisons. It is guaranteed to yield an overall level of type I error (rejecting the null hypothesis when it is true) that is less than or equal to α. However, it may yield an error level that is much less than α (D. Helsel, 2004, personal commun.), which increases the possibility of making a type II error—failing to reject the null hypothesis when it is not true. Our analyses are therefore very conservative and may not have detected all significant differences between pre- and posteruption peak flows.

To evaluate relative differences between pre- and posteruption discharges in the disturbed basins, we retransformed logarithms of unit-area discharges predicted by the regression models into original units and compared differences between discharges at benchmark flow magnitudes in the control basin. Duan's smearing procedure (e.g., Helsel and Hirsch, 1992) was used to correct for bias in the retransformation process. Bias corrections ranged from 4% to 26%; the median correction was 6.4%. Mean annual flow (Q_{mean}) and 1-, 2-, 5-, and 10-yr recurrence interval floods ($Q_{1\ yr}$, $Q_{2\ yr}$, $Q_{5\ yr}$, $Q_{10\ yr}$) in the control basin (Sumioka et al., 1998; D.L. Kresch, 2004, personal commun.) were selected as the benchmark discharges for which we determined relative changes in the disturbed basins. We selected these flow frequencies as evaluation benchmarks because they bracketed most of the posteruption peaks and they provided categorical discharges that usefully elucidated the distribution of the magnitudes of the matched discharges. In the following analyses, we restrict our discussions to discharges within these bounds.

RESULTS

Discharge Peaks

Eruption-induced landscape changes generally amplified peak flows from severely disturbed basins, whereas pre- and posteruption peak flows from basins unaffected by the eruption were not significantly different. However, peak flow responses to the eruption were complex, relatively short-lived, and they varied with respect to the nature of volcanic disturbance, season, discharge magnitude, and time since the eruption.

Response to Style of Volcanic Disturbance

Hydrological responses to the volcanically induced landscape disturbances were strongest from basins in which both hillslope hydrology and channel hydraulics were altered, and weakest from basins affected only by moderate to minor tephra fall. Comparisons of regression models show that posteruption peaks exceeded pre-eruption peaks for at least 5 yr in the following areas: in the North Fork Toutle River basin (KID; Fig. 1), which was affected by the large debris avalanche, a voluminous debris flow, and the devastating volcanic blast; in the South Fork Toutle (SFT) and Muddy (MUD) River basins, the headwaters of which were affected by the volcanic blast and channels of which were affected by large debris flows; in the Green River (GRE) basin, where the volcanic blast razed hundreds of square kilometers of forest and deposited several centimeters of tephra fall; and along the lower Toutle River (TOW), which amalgamated the responses from several basins (Tables DR1, DR2 and Fig. DR1[2]; Figs. 1 and 8). After ~5 yr, the response signals from those basins were largely attenuated. In basins affected only by minor to moderate Plinian tephra fall (Speelyai Creek [SPE] and Cispus River [CIS]), differences between pre- and posteruption regression models are not significant (with an exception noted in the following discussion; Table DR2, Fig. DR2 [see footnote 2]; Fig. 8). Furthermore, pre- and posteruption regression models for basins unaffected by the 1980 eruption are not significantly different, a finding that bolsters the significance of the results from basins severely impacted by the eruption.

Seasonal Response and Temporal Variation

Amplified peak flows from the heavily impacted basins occurred predominantly in autumn (Oct.–Dec.) (Table DR2; Fig. 8A–D), and they persisted chiefly through 1984. In the Toutle River and North Fork Toutle River basins (Fig. 1), however, amplified autumn peak flows persisted longer, through 1989 on the Toutle River and 1987 on the North Fork Toutle River (Table DR2; Fig. 8B). Winter peak flows were also amplified on Toutle River and North Fork Toutle River through 1984 (Table DR2; Fig. 8E), whereas they generally remained unchanged elsewhere. Locally peculiar amplifications of autumn and winter peaks, which may not be related to eruption-induced landscape changes, were detected on South Fork Toutle River and Speelyai Creek in the 1990s (Table DR2; Figs. 1, 8C, 8D, 8G, and 8H). In contrast to the wet-season responses, posteruption spring peaks on the Toutle River were generally depressed through 1984 (Table DR2; Fig. DR2A–D), whereas they remained unchanged in other basins (Table DR2; Fig. DR2A–D). Within the confines of our limited data, posteruption summer peaks remained unchanged in all basins (Table DR2; Fig. DR2E–I).

Response by Flow Magnitude

Small and large autumn peak flows from all heavily disturbed basins were amplified by several percent to many tens of percent through 1984 (Table 4; Fig. 8A–D); from 1985 through 1989, autumn flows were amplified to a lesser extent on the Toutle River (Table 4; Fig. 8B), and from 1985 through 1987, moderate to large autumn flows were depressed relative to smaller flows on the North Fork Toutle River (Table 4; Fig. 8B).

Although the peak flow responses to the eruption were distinctly seasonal, the nature of the responses varied with flow magnitude. In some basins (Toutle River, South Fork Toutle River, Muddy River), both small and large autumn peaks (predicted by regression models) were amplified nearly proportionately (Table 4; Fig. 8A), whereas in others (North Fork Toutle River, Green River), small to moderate peaks were amplified disproportionately relative to large peaks (Table 4; Fig. 8A–B).

Winter peaks were amplified less than autumn peaks. Small and large winter peaks on the Toutle River and North Fork Toutle River exceeded pre-eruption peaks by a few tens of percent through 1984 (Table 4; Fig. 8E). On the Toutle River, both small and large winter peaks (predicted by regression models) were amplified nearly proportionately, whereas small winter peaks were amplified more on the North Fork Toutle River (Table 4). Winter peaks were also amplified by a few tens of percent in the 1990s on the South Fork Toutle River (Table 4; Fig. 8G–H), but there was no consistent relationship between small and large flows.

Spring peaks were generally unchanged by the eruption. On the Toutle River, however, small ($<Q_{1\ yr}$) spring peaks were depressed through 1984, whereas predicted large peaks (matched to control-basin discharges $Q_{2\ yr}$ to $Q_{10\ yr}$) were amplified by a few to several tens of percent (Table 4; Fig. DR2A, see footnote 2). Overall, seasonal peak flows from basins affected by both severe channel and hillslope disturbances increased in magnitude more than did those from a basin (Green River) subjected solely to severe hillslope disturbance. Peaks on Toutle River, North Fork Toutle River, South Fork Toutle River, and Muddy River (Fig. 1; Table 4) matched to control-basin flows $\geq Q_{1\ yr}$ increased 7%–150%, whereas those on the Green River increased ~20%–30%.

The variations in discharges among different basins, seasons, and flow magnitudes show that the hydrological response to the Mount St. Helens eruption was complex and inconsistent, and they also suggest that changes to channel hydraulics, and not just to hillslope hydrology, played a prominent role in the hydrological response.

Hydrograph Rise Times

In several disturbed basins, hydrographs over a range of peak flows, but chiefly those matched with control-basin flows smaller than $Q_{2\ yr}$, displayed median values of rise times that were accelerated or retarded by tens of percent

[2]GSA Data Repository item 2006123, Tables DR1–DR4 providing summaries of statistical analyses, and Figures DR1 and DR2 showing plots comparing seasonally undifferentiated and spring and summer peak flows, respectively, is available on the Web at http://www.geosociety.org/pubs/ft2006.htm. Requests may also be sent to editing@geosociety.org.

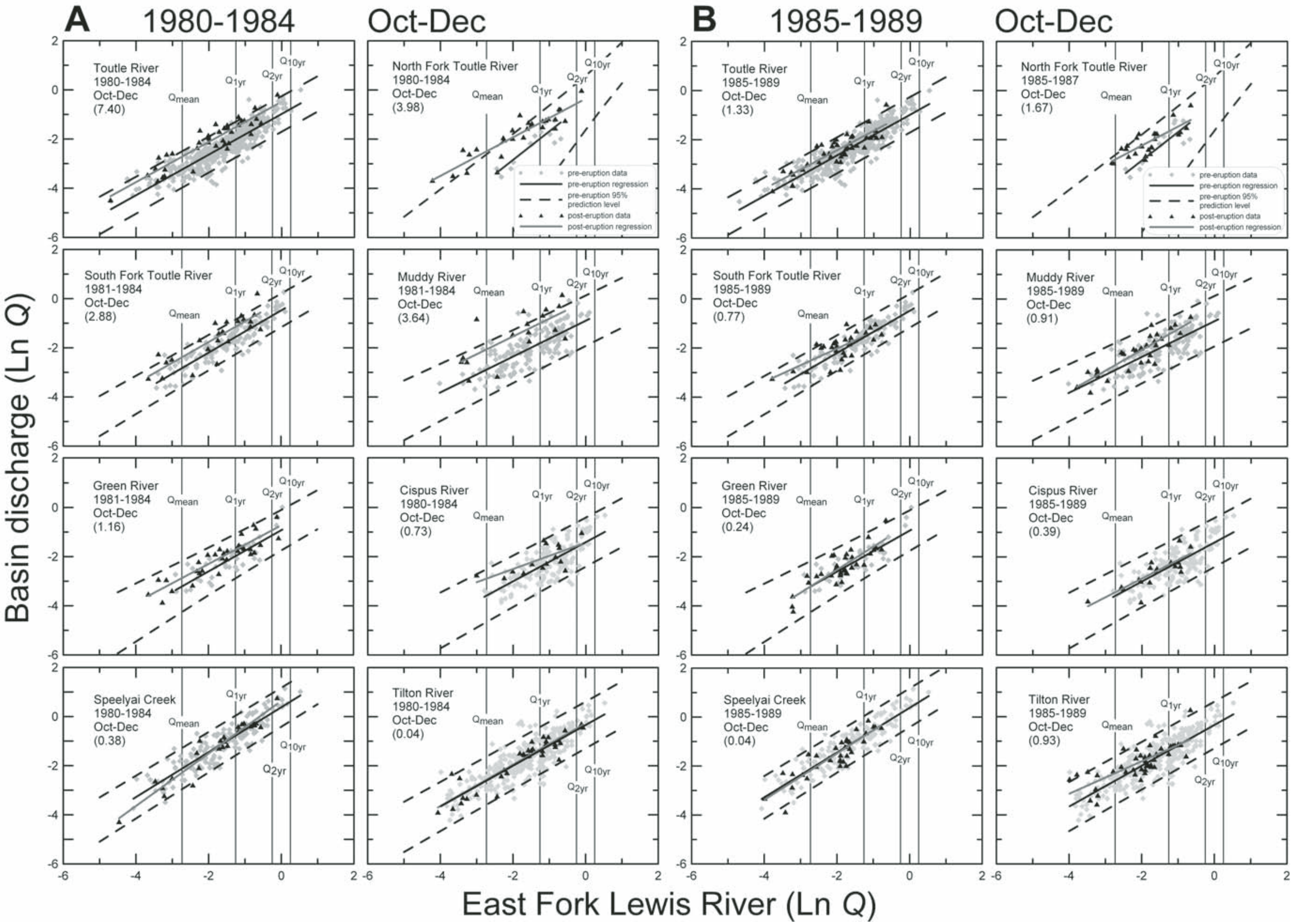
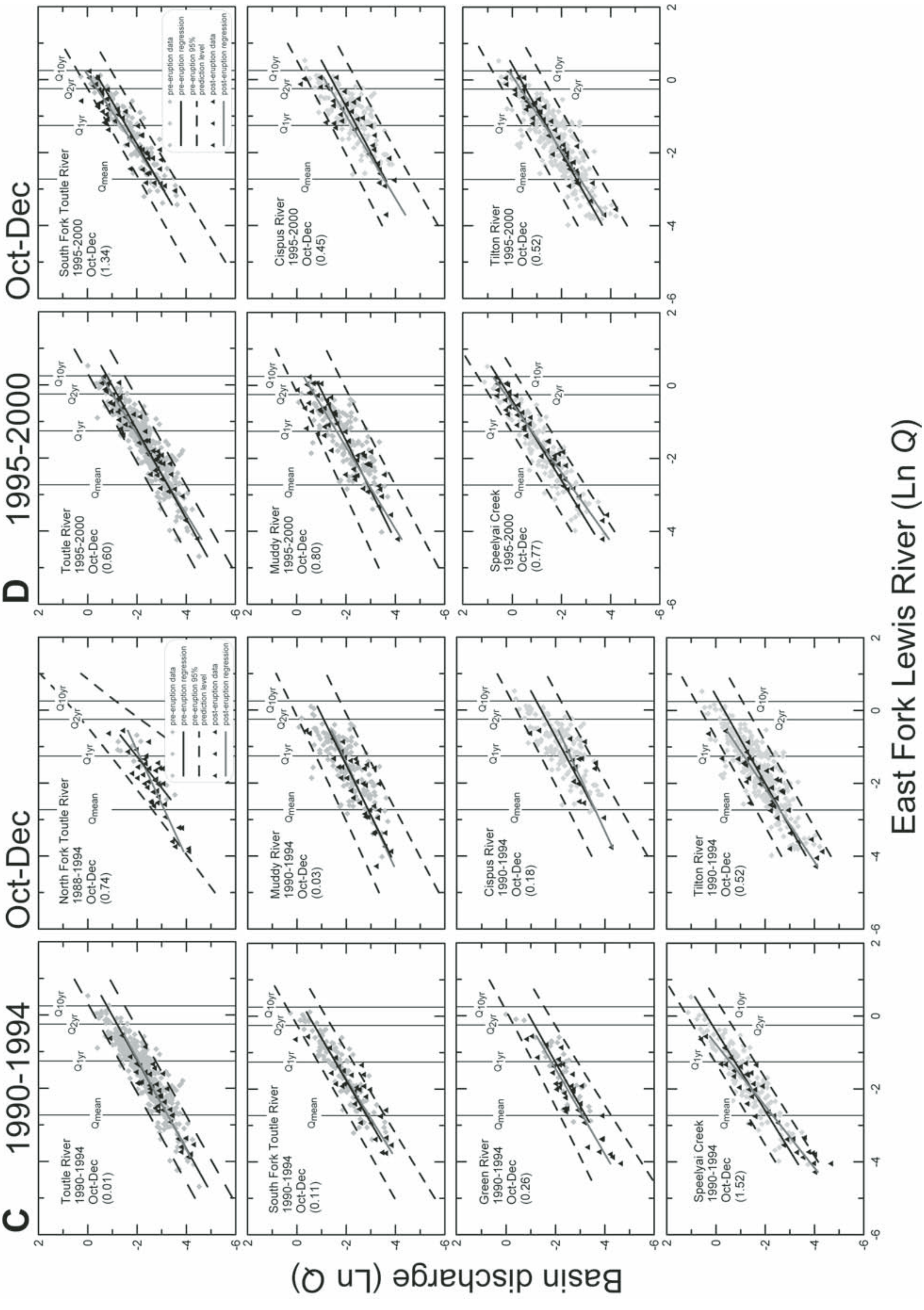


Figure 8. Plots of logarithms of unit-area discharges of autumn and winter peak flows from Mount St. Helens and nearby basins paired with peak flows from the East Fork Lewis River (control) basin (cf. Fig. 1; Table 2). Regression models are fit to pre- and posteruption paired discharges. In some basins, small and large posteruption peak flows increased roughly proportionately, whereas in others, small peak flows increased disproportionately (cf. Table 4). Values in parentheses give ratio of F statistic to critical F value (cf. Table DR2, see text footnote 2). After 1985, pre- and posteruption regression models in most basins were not significantly different. Units of discharge (Q) are ($m^3 s^{-1} km^{-2}$). (*Continued on the following three pages.*)



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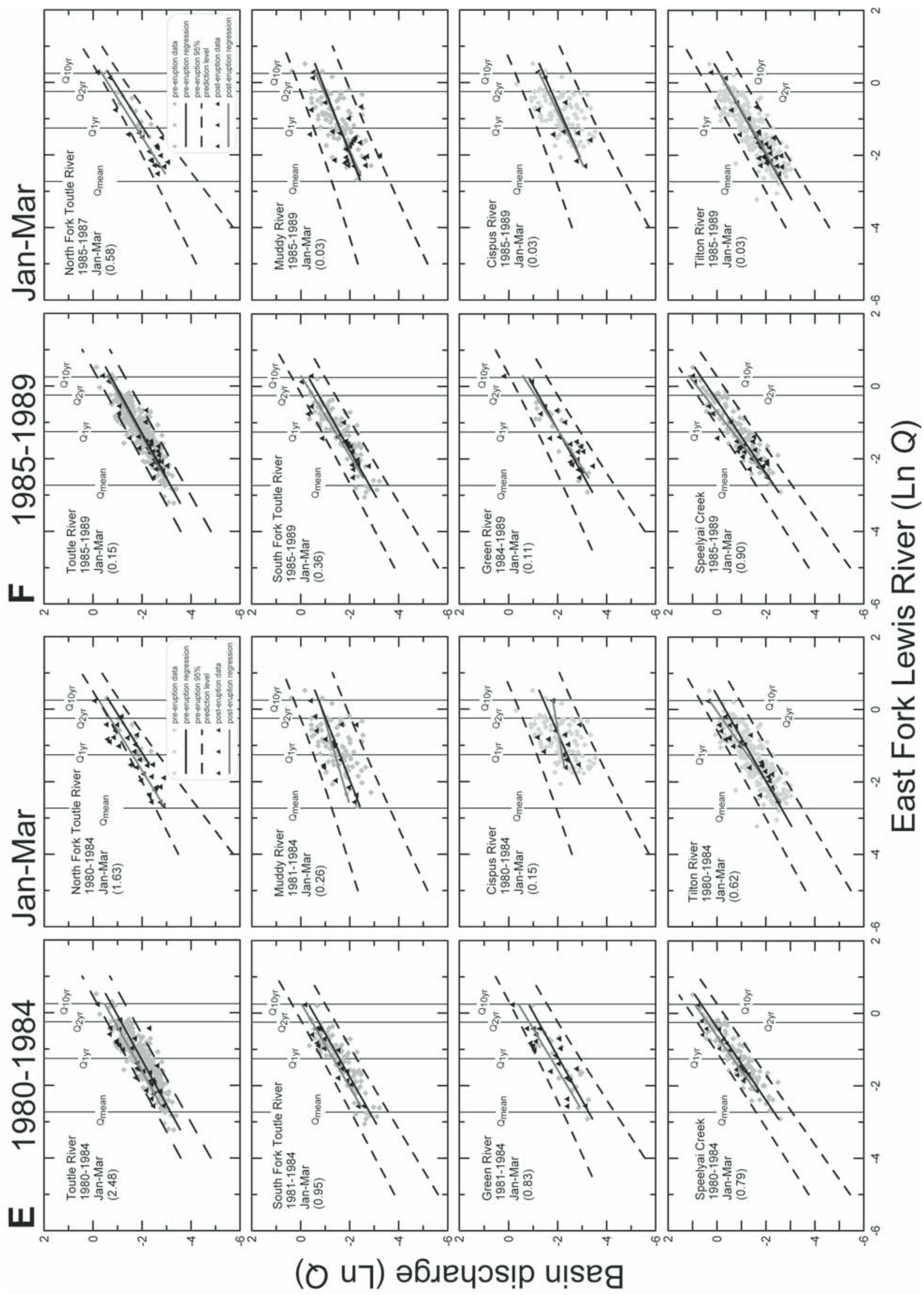

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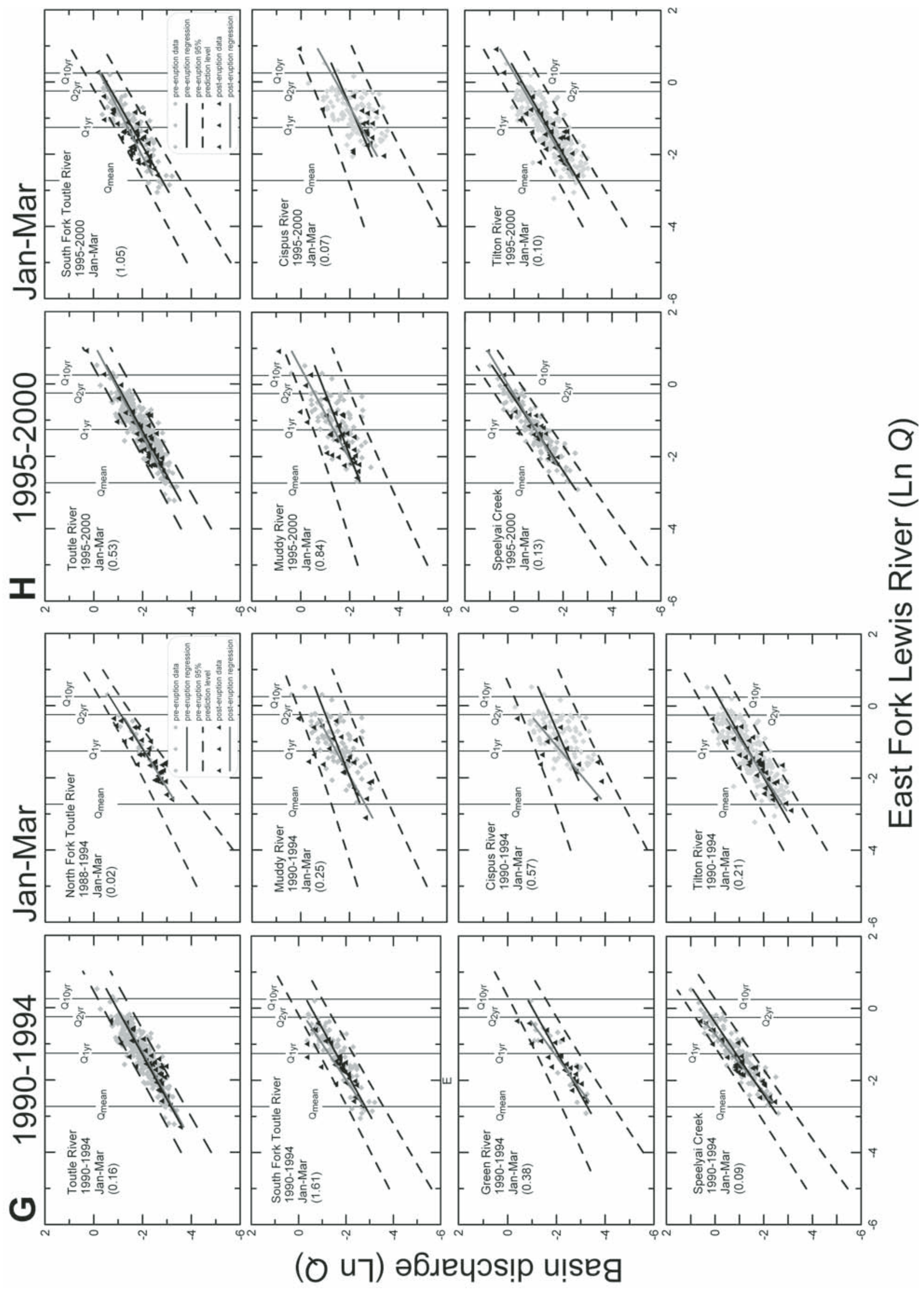

Figure 8 ***(continued)*****.**

relative to pre-eruption hydrographs. In the 1980s, seasonally undifferentiated posteruption hydrographs on Toutle River, North Fork Toutle River, Muddy River, and Green River (Fig. 1; Table DR3, see footnote 2) generally rose more rapidly than did pre-eruption hydrographs. After 1990, discharges on Toutle River, North Fork Toutle River, and Muddy River typically rose more slowly than did pre-eruption discharges. Such trends, however, were not universal among the disturbed basins (Table DR3). On the South Fork Toutle River (Fig. 1), for example, post-eruption hydrographs ubiquitously rose more slowly than did pre-eruption hydrographs, whereas on the Green River (Fig. 1) discharges typically rose more rapidly. Rise times of post-1980 hydrographs from less severely disturbed basins (Speelyai Creek, Cispus River; Fig. 1) remained unchanged or rose more slowly than did pre-eruption hydrographs (Table DR3).

Undifferentiated hydrographs for basins unaffected by the 1980 eruption also displayed median values of rise times that were accelerated or retarded by tens of percent relative to pre-1980 hydrographs. In the East Fork Lewis River basin (the control basin) (Fig. 1; Table 2), post-1980 hydrographs of small- and moderate-sized peak flows ($<Q_{2\ yr}$) rose more slowly than did pre-1980 hydrographs. In contrast, 1980s hydrographs of similarly sized flows on Coweman River and Tilton River (Fig. 1; Table 2) rose more rapidly than did pre-1980 hydrographs, similar to behavior documented in several basins disturbed by the eruption (Table DR3). Relative changes in post-1980 hydrograph rise times in basins unaffected by the eruption were, however, generally less than those in the disturbed basins (Table DR3).

Seasonal posteruption hydrographs also exhibited accelerated and retarded rise times. In several basins (Toutle River, North Fork Toutle River, Green River, Muddy River, but not South Fork Toutle River; cf. Fig. 1), chiefly moderate-magnitude ($Q_{mean} < Q < Q_{2\ yr}$) autumn and winter peaks rose more rapidly in the 1980s than did corresponding pre-eruption peaks (Table DR4, see footnote 2; Fig. 9). After 1990, seasonal peak flows rose more slowly than did pre-eruption peaks or they remained unchanged (Table DR4; Fig. 9). Posteruption rise times of seasonal flows in less-disturbed basins (Speelyai Creek, Cispus River) remained mostly unchanged or rose more slowly than pre-eruption flows (Table DR4; Fig. 9).

Seasonal flows in basins unaffected by the eruption likewise displayed accelerated or retarded rise times. Rise times of posteruption seasonal hydrographs on the East Fork Lewis River were typically retarded, whereas those on Coweman River and Tilton River remained unchanged or were somewhat accelerated (Table DR4; Fig. 9). Again, relative changes in rise times of posteruption seasonal hydrographs in these basins were generally less than those in basins heavily disturbed (Table DR4).

TABLE 4. RELATIVE CHANGES BETWEEN SEASONAL PRE- AND POSTERUPTION UNIT-AREA DISCHARGES AT MOUNT ST. HELENS THAT CORRESPOND TO VARIOUS FREQUENCY FLOWS IN THE EAST FORK LEWIS RIVER CONTROL BASIN

Basin	Period	Q_{mean}	$Q_{1\ yr}$	$Q_{2\ yr}$	$Q_{5\ yr}$	$Q_{10\ yr}$
				(%)		
Toutle River (TOW)						
Oct–Dec	1980–1984	68	66	65	64	64
	1985–1989	26	19	15	13	12
Jan–Mar	1980–1984	30	31	32	32	32
Apr–Jun	1980–1984	–26	20	67	87	98
North Fork Toutle (KID)						
Oct–Dec	1980–1984	230	90	30	15	7
	1985–1987	142	37	–7	–18	–24
Jan–Mar	1980–1984	68	62	59	58	57
South Fork Toutle (SFT)						
Oct–Dec	1981–1984	62	58	56	55	54
	1995–2000	3	24	40	46	50
Jan–Mar	1990–1994	10	42	68	78	84
	1995–2000	35	21	12	9	7
Muddy River (MUD)						
Oct–Dec	1981–1984	151	149	148	147	146
Green River (GRE)						
Oct–Dec	1980–1984	38	28	21	19	18
Speelyai Creek (SPE)						
Oct–Dec	1990–1994	–15	25	64	79	87

Note: Percentages have been rounded to the nearest whole number.

In general, differences among rise times between pre- and posteruption hydrographs are most prominent for autumn and winter peaks smaller than the 2 yr flood. Some signal is apparent in winter flows larger than the 2 yr flood on the Toutle River between 1980 and 1984 (Table DR4). Overall, however, posteruption rise times of flows larger than the 2 yr flood remained unchanged regardless of the nature of the disturbance (Tables DR3 and DR4).

DISCUSSION

Comparison of Predicted versus Measured Discharge Response

Hydrological models predicted that posteruption peak flows would be larger, rise faster, and occur more frequently than pre-eruption peak flows. Predicted peaks of posteruption unit hydrographs were 50% greater and rose ~25% faster than those of pre-eruption unit hydrographs (Orwig and Mathison, 1982). Predicted responses also varied with discharge magnitude, with changes greatest for small- to moderate-magnitude flows; the predicted 2 yr flood on the Toutle River increased as much as 60%, whereas the 50 yr flood increased as much as 25% (Lettenmaier and Burges, 1981). In contrast, Datta et al. (1983) predicted little response of daily discharges during periods of high flow (Oct.–Mar.) and depressed daily discharges during summer.

The predicted responses reflect various conceptual reasonings of the effects of volcanic disturbances on hillslope hydrology. In general, the various models assume an increase in areal distribution of impervious landscape (a proxy for reduced infiltration loss) resulting from tephra deposition. This assumption steered more precipitation to overland flow rather than to soil-moisture storage and more modulated throughflow, and the redirected runoff chiefly affected responses of small and moderate flows. Lettenmaier and Burges (1981) reasoned that prolonged or intense storms that generated large flows typically occurred during the wet season when pre-eruption soils were wet, had little storage capacity, and were rapidly responsive. At those times, even before the eruption, more precipitation was also directed to saturation overland flow or very shallow throughflow than to soil-moisture storage. Thus, an increased impervious area had only modest effect on the modeled response to large storms,

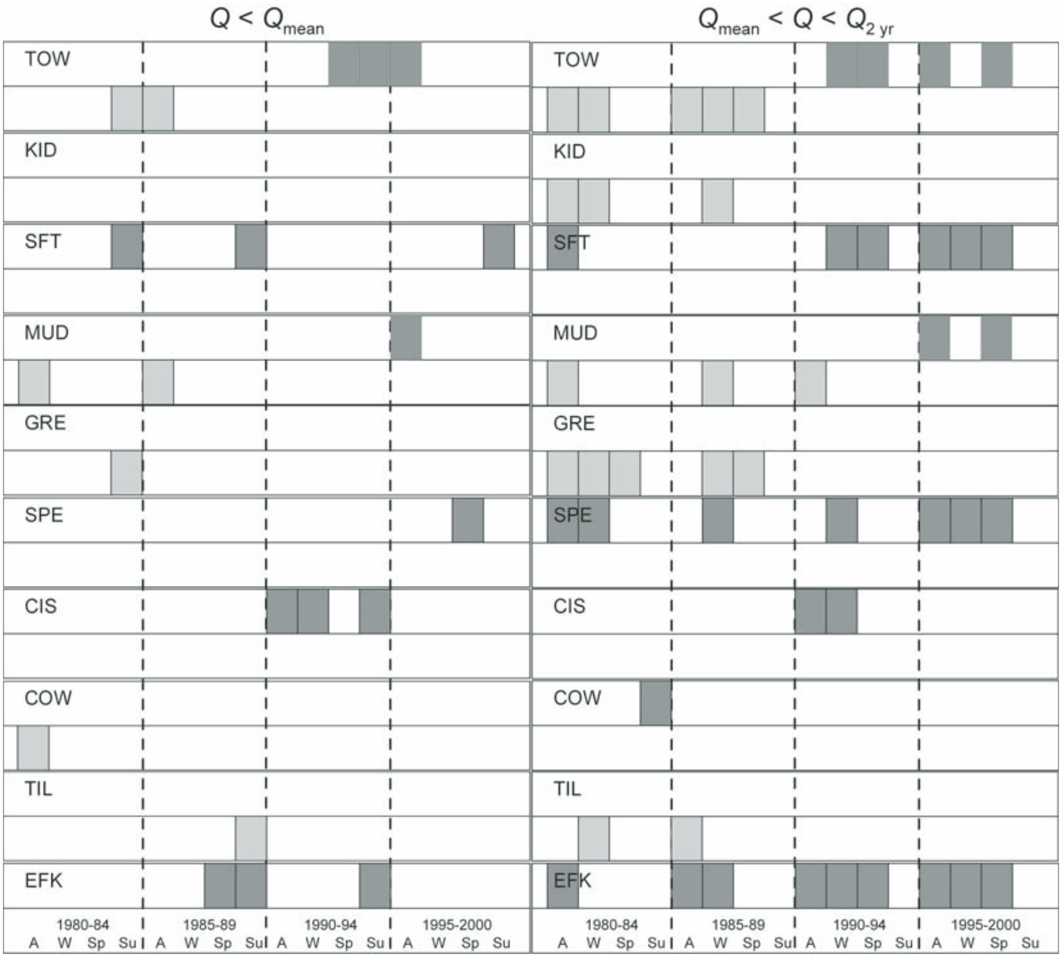

Figure 9. Bar plots showing seasons and time periods when posteruption hydrograph rise times were faster or slower than corresponding pre-eruption hydrograph rise times. Only periods exhibiting significant differences are shown. Negative bars (below central line) indicate faster posteruption rise times; positive bars (above central line) indicate slower posteruption rise times. Q_{mean} represents mean annual flow; $Q_{2\,yr}$ represents the 2 yr flood discharge (cf. Table DR4, see text footnote 2). See Figure 1 for gauging station locations (e.g., TOW).

and hence a modest effect on large peak flows. In contrast, less-intense storms that generated small to moderate flows were assumed to be more typical of times when evapotranspiration was active and pre-eruption soils had greater storage capacity. At those times, precipitation and snow melt contributed chiefly to soil-moisture recharge and more modulated throughflow. Under those conditions an increased impervious area had a greater effect on the modeled response of small to moderate peak flows. In another model (Datta et al., 1983), enhanced depression storage caused by downed trees and tephra-surface irregularities offset the effects of reduced infiltration. In that model, the competition between enhanced runoff and greater surface storage modulated the modeled wet-season response. Furthermore, surface storage inhibited soil-moisture recharge, reduced base flow, and consequently depressed modeled dry-season discharge. Implicitly, these models predicted substantially increased small to moderate peak flows in autumn, moderately increased peak flows in winter and spring, and depressed flow in summer.

Actual peak flow responses to the 1980 eruption were transient, strongly seasonal, and inconsistent with respect to flow magnitude. In heavily disturbed basins closest to the volcano, mean values of autumn peak flows increased by a few to several tens of percent for ~5 yr, and for nearly a decade on North Fork Toutle River and Toutle River. Mean values of winter peak flow at these locations also increased by tens of percent for ~5 yr. In some basins (Toutle River, Muddy River, South Fork Toutle River; Fig. 1; Table 4), mean values of seasonal peak flows over a range of magnitudes increased roughly proportionately through 1984. Elsewhere, or at other times, seasonal small flows increased disproportionately compared to large flows (North Fork Toutle River, Green River; Fig. 1; Table 4), and some even decreased relative to large flows (Toutle River; Table 4). After 1984, changes in peak flows were more subdued and less consistent.

The documented longer-term hydrological patterns generally reflect responses to evolving, posteruptive landscape adjustments. Discharges measured on the North Fork Toutle River after 1988 (Figs. 8C and 8G), however, followed construction upstream of a large sediment-retention structure (Fig. 1). Although that structure was designed to trap sediment but pass water, it clipped peak flows and affected large flows more than small flows. In the 1990s, mean values of autumn and winter peaks on the South Fork Toutle River and Speelyai Creek increased some tens of percent (Figs. 8C, 8D, 8G, and 8H), but they did not exhibit a consistent response between small and large flows. Those apparent changes may have reflected the influences of renewed logging and other land uses rather than responses to eruptive disturbance.

The transiently enhanced mean values of peak flows documented here (Table 4) are consistent with the predicted increases of unit-hydrograph peaks, although the extraordinarily enhanced small autumn flows on the North Fork Toutle River through 1987 and the more broadly enhanced autumn flows on the Muddy River through 1984 greatly exceeded predicted changes. However, the proportionately increased mean values of both small and large autumn flows on Toutle River, Muddy River, and South

Fork Toutle River, and of winter flows on Toutle River and North Fork Toutle River contrast with predicted disproportionate increases of small to moderate flows. In general, only the disproportionately increased small autumn flows on the Green River, where disturbances chiefly affected hillslope hydrology, on the North Fork Toutle River, below a zone of complex hillslope and channel disturbances, and on the Toutle River from 1985 through 1989 were consistent with predicted responses by flow sizes.

Another way of evaluating the posteruption hydrological response is to estimate the change in magnitude of a flow of a given frequency. Flood-frequency estimates made on the basis of analyses of annual maximum discharges through 1979 (pre-eruption) (Williams and Pearson, 1985) and updated through 1996 (posteruption) (Sumioka et al., 1998; D.L. Kresch, 2004, personal commun.) indicate substantial changes in flood-frequency relations in basins disturbed by the eruption, but the changes are contrary to predictions. Updated flood-frequency analyses suggest that large flows have experienced the greatest change. In the Toutle, South Fork Toutle, and Muddy River basins, the 2 yr flood increased <6%, the 50 yr flood increased 15%–40%, and the 100 yr flood increased 20%–45% (data are insufficient to estimate pre-1980 flood-frequency relations for the North Fork Toutle River and Green River). In basins less affected by the eruption (Speelyai Creek and Cispus River), only those discharges on the Cispus River larger than or equal to the 25 yr flood changed by more than 10%. In basins unaffected by the eruption (Coweman River, Tilton River, East Fork Lewis River), magnitudes of floods as large as the 100 yr event remained unchanged or changed by less than 10%. Some changes in the updated flood-frequency relations simply reflect analyses of longer time series and not disturbances caused by the 1980 eruption. However, the estimated magnitudes of change of the large flows in the disturbed basins cannot be explained simply as time-series artifacts. Those magnitudes of change, when compared to changes in basins little affected or unaffected by the eruption, clearly reflect hydrological responses to the eruption. Our analyses of an extensive set of discharges rather than annual maximum flows, however, show that the hydrological responses were more complex than suggested by a simple comparison of pre- and posteruption flood-frequency relations.

In several disturbed basins, hydrographs during the 1980s rose more rapidly than predicted. Posteruption unit hydrographs were predicted to rise ~25% more rapidly than pre-eruption unit hydrographs (Orwig and Mathison, 1982). Instead, on all but the South Fork Toutle River, median values of rise times of peak flows smaller than $Q_{2\,\mathrm{yr}}$ were as much as 60% faster than pre-eruption peaks, and those of discharges larger than $Q_{2\,\mathrm{yr}}$ were locally as much as 80% faster (Tables DR3, DR4, see footnote 2). The geomorphic significance of these changes in rise times is clouded, however, by similar changes in some basins unaffected by the eruption (e.g., Tilton River; Tables DR3, DR4). It is probable that the faster rise times of posteruption hydrographs at Mount St. Helens are related to hydrologic and geomorphic changes caused by eruptive disturbance, as a strong signal shows up in multiple basins. However, the data are insufficiently compelling to draw a firm conclusion.

Transient but Inconsistent Responses of Peak Flows

The 1980 eruption of Mount St. Helens caused a widespread, transformative landscape disturbance, yet the peak flow responses from affected basins were transient and inconsistent. Part of the response transience can be attributed to the sizes of upstream drainage areas (all but one in excess of 300 km^2) relative to the nature and percentage of basin disturbance (cf. Tables 2 and 3). To some degree, the spatial distributions of the disturbances, the generally high elevations of the most intensely disturbed zones (typically within or near the snow accumulation zones in basin headwaters), and the capacity of large watersheds to modulate peak flows had an effect on hydrological responses.

The nature and pace of secondary landscape modifications (e.g., Swanson and Major, 2005), however, exerted the greatest influence on the evolution of peak flow responses. Tephra-surface modifications by erosional, biogenic, and cryogenic processes rapidly modulated post-1980 infiltration characteristics, even in the absence of deliberate, mechanical land-management practices or extensive regrowth (e.g., Collins and Dunne, 1988; Major and Yamakoshi, 2005); hence, surface runoff was swiftly reduced (e.g., Swanson et al., 1983; Collins and Dunne, 1988; Leavesley et al., 1989; Major and Yamakoshi, 2005). Leavesley et al. (1989) noted significant surface coarsening within one year of the eruption. By 1981, surfaces on the blast deposit had coarsened from sandy silt to sandy loam, and infiltration capacities had increased roughly twofold. Dramatic surface coarsening also occurred east of the volcano in areas where a thin mantle of fine tephra draped pumice gravel (Swanson et al., 1983). In both areas, surface textures coarsened in the absence of significant revegetation. By 1982, introduced and native vegetation broadly covered slopes in the Green River basin (Collins and Dunne, 1988), although vegetation recovery was slower elsewhere (e.g., Lawrence, 2005). Vegetation development, especially grasses, rapidly modified soil structure and increased infiltration (e.g., Hino et al., 1987). Within 2 yr after the eruption, hillslope water transfer was significantly modified, and overland flow had been substantially reduced (e.g., Collins and Dunne, 1986; Swanson and Major, 2005).

Rates of stabilization varied greatly among channels affected by the debris avalanche and debris flows, but dramatic adjustments, and consequent extraordinary sediment transport, declined sharply within a few years of the eruption (Martinson et al., 1984; Meyer et al., 1986; Meyer and Martinson, 1989; Simon, 1999; Major et al., 2000; Hardison, 2000). Channel locations and geometries changed most rapidly through 1981 as channels incised and widened. Within 5 yr, dramatic channel changes were largely complete, although some reaches exhibited progressive longer-term change (D.R. Saunders, 2005, personal commun.). In general, channels across the debris-avalanche deposit widened hundreds of meters and incised tens of meters, and those affected by debris flows rapidly widened by tens of meters and incised up to ten meters (Meyer and Martinson, 1989). Since the 1980 eruption, channel beds have coarsened considerably. Immediately after the eruption, median (d_{50}) bed-material sizes in the Toutle River basin were largely coarse sand (0.5–1 mm diameter) (Simon, 1999). Within 2 yr of the eruption, median bed-material sizes in most basins had coarsened by a factor of two or more to very coarse sand, and within a decade had coarsened by two orders of magnitude to fine gravel (Simon, 1999). As channels widened and beds coarsened, flow resistance increased substantially. Discharge peaks diminished rapidly owing to increased flow resistance, relative channel stabilization, and reduced runoff.

The inconsistent nature of responses of seasonal peak flows among basins and different time periods reflects data limitations as well as physical modifications to the landscape. The nature of the responses detected using paired-basin regression analysis was influenced by the predominance of small to moderate peaks that populate the data set. In nearly all of the basins analyzed, regression models through 1994 were driven chiefly by flows paired with control-basin discharges smaller than $Q_{2\,\mathrm{yr}}$, and predominantly by those paired with discharges smaller than $Q_{1\,\mathrm{yr}}$ (Fig. 8). From 1995 through 2000, a period of wetter-than-average conditions in the Pacific Northwest, during which regional floods of record occurred, peaks that paired with control-basin flows larger than $Q_{1\,\mathrm{yr}}$ exerted a stronger influence on the regression models. Furthermore, in the North Fork Toutle and Green River basins, pre-eruption data span a limited period (cf. Table 2), and consequent regression models may not be truly representative of mean pre-eruption conditions. Likewise, matched pairs of spring and summer peaks were relatively rare compared to those of autumn and winter peaks, making it difficult to detect significant peak flow responses in spring and summer.

The inconsistent seasonal responses among basins and various time periods suggest that peak flows responded to more than just perturbed hillslope hydrology. Conceptual and numerical models discussed previously indicate that increased throughfall, reduced infiltration, reduced evapotranspiration, altered depression storage, and modified accumulation and melt of snowpack (cf. Fig. 5) are most likely to enhance small to moderate autumn peak flows, and less likely to substantially affect large winter peaks. The predominance of responses by small to moderate autumn and some winter peak flows is consistent with geophysical and ecological disturbances that modified precipitation losses on hillslopes and directed greater quantities of runoff more rapidly to channels (cf. Fig. 5). However, the relatively large increases in autumn and winter peak flows that paired with control-basin discharges larger than $Q_{2\,\mathrm{yr}}$ are not as readily explained by simple perturbations to hillslope hydrology. Larger peak flows of those magnitudes require mechanisms that disproportionately enhance large flows, or, alternatively, mechanisms that exert a proportionately greater resistance on small flows relative to large flows. Although it is difficult, or perhaps impossible, to completely disentangle the effects of perturbations to hillslope hydrology from altered channel hydraulics on the hydrological responses, we propose that the spatial extent of the eruption-induced channel disturbances, the nature and pace of secondary channel modifications, and associated changes in channel hydraulics can partly explain the inconsistencies among discharge responses and the proportionate increases of both small and large peak flows observed on Toutle River, South Fork Toutle River, and Muddy River.

Influence of Channel Changes and Sediment Transport on Peak Discharges

There are two basic ways to increase flow magnitude: direct greater quantities of runoff more rapidly from hillslopes into channels, or increase channel efficiency. If channel efficiency and sediment transport did not affect peak discharges, then posteruption peak flows would have been routed along channels in a manner similar to pre-eruption flows regardless of magnitude. Hence, changes only to hillslope runoff would have caused differences between pre- and posteruption discharges, which we and others argue would have disproportionately affected small to moderate flows. We propose that proportionately increased mean values of both large and small peak flows resulted chiefly from changes in channel efficiency and enhanced sediment transport, which differentially affected flows of various magnitudes.

Peak flow can be spatially enhanced by reducing valley storage or increasing channel efficiency in addition to modifying runoff. Valley storage occurs when a flood spills over bank. If a flow remains confined, valley storage is minimized and peak discharge is enhanced. Reducing valley storage disproportionately affects large flows, because small flows are normally confined.

Changes in valley storage likely had little effect on posteruption peak flows. Across the debris-avalanche deposit and along channels affected by large debris flows, channel responses followed complex cycles of incision, widening, and aggradation (Meyer and Martinson, 1989). Initially, the North Fork Toutle River incised the debris-avalanche deposit and maintained a constant width-to-depth ratio during the first year after the 18 May eruption. This incision was followed by substantial channel widening and bed aggradation, then further widening with little net change in bed elevation. Along channels affected by debris flows, steep upstream reaches were incised and widened during the first year after the eruption, then principally aggraded, whereas gentle downstream reaches were aggraded and widened, and later became incised. Changes in posteruption channel geometries thus varied considerably with the nature of disturbance and position along a valley. In general, any discharge enhancements that resulted from channel incision and minimization of valley storage were short-lived.

Discharge also can be enhanced by increasing channel efficiency, accomplished by reducing flow resistance. Flow resistance can be reduced by smoothing the channel bed, thereby reducing the skin friction of stationary particles. Along the Toutle, South Fork Toutle, lower North Fork Toutle, and Muddy Rivers, the passage of large debris flows straightened and smoothed channels, and changed them from sinuous, cobble-bedded, pool-riffle systems to streamlined, sand-bedded systems (Janda et al., 1984). But radically reducing skin friction affects transit of both large and small flows; other forms of enhancing channel efficiency must affect chiefly large flows if our hypothesis has merit. Flow resistance also can be reduced by minimizing channel sinuosity and form drag, and greatly increasing sediment transport.

Channel straightening affects large flows more than small flows. Even with reduced channel sinuosity, small peak flows typically follow sinuous paths influenced by the presence of bed forms that they cannot mobilize. This is especially true under the braid-like conditions that commonly develop in channels severely disturbed by eruptions (cf. Fig. 4F; e.g., Hayes et al., 2002; Gran and Montgomery, 2005). Hence, skin friction, form drag, and in-channel sinuosity offer resistance to small flows even in highly disturbed systems. In contrast, larger discharges moving along alluvial channels that have extremely mobile beds have the ability to mobilize bed forms and to follow more streamlined flow paths, especially in heavily sediment-laden, volcanically disturbed systems (e.g., Montgomery et al., 1999; Hayes et al., 2002; Gran and Montgomery, 2005).

Large sediment loads can affect flow resistance in multiple ways. Elevated loads of suspended sediment can partly damp turbulence and decrease flow resistance (e.g., Vanoni, 1946; Dinehart, 1987). Hydraulic roughness, however, exerts more influence on flow resistance than does suspended-sediment transport (Pitlick, 1992). Bed-material grain size (skin friction), bed forms (form drag), and channel form, all of which are greatly influenced by the quantity and caliber of bedload transport, affect hydraulic roughness. Under lower- and upper-stage plane-bed conditions (e.g., Middleton and Southard, 1984), the bulk of flow resistance is provided by grain roughness and channel form (e.g., Simons et al., 1965; Denlinger et al., 2001), and under intense bedload transport, the grain-size distributions of bedload and bed material are similar (e.g., Pitlick, 1992). Because smaller discharges have shallower flow depths than larger discharges, the length scale of grain roughness affects a greater proportion of the depth of smaller discharges than larger discharges. Therefore, under plane-bed conditions, larger flows "feel" disproportionately less frictional resistance than smaller flows. If bed sediment is mobilized into plane-bed conditions, common in sediment-laden volcanically disturbed systems (e.g., Montgomery et al., 1999; Gran and Montgomery, 2005), larger discharges are disproportionately less affected by skin friction and form drag than smaller discharges. In contrast, if high rates of bedload transport develop significant bed forms rather than plane-bed conditions, resistance can greatly increase as flows become depth-limited with respect to bed-form amplitude (Pitlick, 1992).

A combination of greater discharges, smoother and straighter channels, and efficient sediment transport commonly produced upper-stage and plane-bed conditions in channels during the first several years after the 1980 eruption (e.g., Dinehart, 1992). Repetitive formation and dissipation of breaking waves (antidunes to plane bed) and extraordinary sediment transport during large-magnitude ($Q > Q_{2\ yr}$) floods (Hammond, 1989; Dinehart, 1998; Childers, 1999; Major, 2004) in severely disturbed channels attest to extraordinary bed mobility and the predominance of grain roughness over form drag under such discharges (e.g., Simons et al., 1965). In contrast, a disproportionate increase in magnitudes of small to moderate flows in the Green River basin is consistent with a lack of significant channel disturbance. Complex disturbance, transient surface-water storage, gradual network reintegration, and a short pre-eruption discharge record make it difficult to interpret the significance of the peak flow response pattern on the North Fork Toutle River.

Comparison of Discharge Responses at Mount St. Helens with Other Disturbances

Although the eruption of Mount St. Helens triggered a broad hydrological response, some explosive eruptions trigger little response. For example, Strombolian eruptions, especially at cinder cones in dry climates, can deposit tephra fall that is sufficiently coarse grained such that no appreciable runoff occurs until enough eolian dust accumulates in the soil over hundreds to thousands of years and reduces infiltration (J.D. Pelletier, 2006, personal commun.). Explosive Plinian eruptions, such as at Mount St. Helens, however, commonly damage or destroy vegetation and deposit a few to several millimeters of fine, silty tephra, either of which can drastically alter surface hydrology. Thus, ecological impact, tephra-surface texture, and climate are important factors that influence posteruption hydrological responses.

Streamflow responses to the 1980 Mount St. Helens eruption are broadly similar to responses to other forms of landscape disturbances. Peak flows affected by forest and agricultural practices, wildfires, mining, and urbanization typically are flashier and have larger peaks than pre-disturbance peak flows, and the increased discharges are most pronounced for small, frequent flows (e.g., Guebert and Gardner, 1992; Jones and Grant, 1996; Jones, 2000; Bowling et al., 2000; Moody and Martin, 2001; Knox, 2001; Konrad et al., 2005; Iroumé et al., 2006). Streamflow responses to forest practices in small (up to a few km^2) watersheds can persist for decades, whereas responses to wildfire typically are transient (a few years) (e.g., Cannon and Gartner, 2005). At Mount St. Helens, peak flow responses in the Green River basin, which was subject chiefly to hillslope disturbance, closely match responses characteristic of many other forms of landscape disturbances. Such a result is not unexpected, because many of those other disturbances chiefly affect hillslope hydrology.

Peak flow responses from watersheds subjected to severe channel, as well as hillslope, disturbance at Mount St. Helens differ from responses to other forms of disturbances. The approximately proportionate increases of both small and large peak discharges on Toutle River, South Fork Toutle River, and Muddy River have no apparent analog, of which we are aware, in response to perturbations chiefly to hillslope hydrology. For that reason, we have proposed that the responses of those basins likely reflect the influences of channel modifications and extraordinary sediment transport. A chief difference between the volcanically induced disturbances at Mount St. Helens and other disturbances elsewhere is the extent to which major channels were significantly modified. Disturbances caused by mining can add significant amounts of sediment to channels, but the spatial extents of channel disturbances by mining typically are less than those at Mount St. Helens. Given the nature and spatial extents of the channel modifications at Mount St. Helens relative to the sizes of the basins, it seems highly probable that primary and secondary modifications to channel hydraulics played a prominent role in the hydrological responses.

CONCLUSIONS

Landscape disturbances by the cataclysmic 1980 eruption of Mount St. Helens induced a widespread, but transient, hydrological response. Posteruption peak flows from basins heavily impacted by the eruption typically had larger magnitudes and generally rose more rapidly than pre-eruption peak flows. However, peak flow responses were distinctly seasonal and short-lived; at the basin-scales examined (300–1300 km^2 drainage areas), the landscape disturbances caused by the eruption principally affected autumn and winter peaks for a period of ~5 yr. In the most heavily disturbed basins, a distinct response lasted as long as a decade. In general, magnitudes of the affected seasonal flows were greater, rise times faster, and large flows more substantively affected than predicted by early modeling efforts.

The strongly seasonal and transient responses to the eruption were driven chiefly by modifications to hillslope hydrology. Vegetation loss allowed more precipitation to reach the land surface, reduced transpiration, and likely altered seasonal soil moisture; tephra deposition greatly reduced surface infiltration. These changes radically altered hillslope hydrology and changed it from a regime of meager surface runoff to one that was transiently dominated by overland flow.

In addition to the signal imparted by the perturbations to hillslope hydrology, we conclude that modifications to channel efficiency imparted a clear imprint on discharge responses. Channel changes caused chiefly by massive sediment deposition altered the manner in which runoff moved downstream. In particular, channel smoothing, straightening, and enhanced sediment transport allowed development of upper-regime flow conditions that significantly reduced flow resistance. Variations in channel geometry and flow resistance, related to modifications of alluvial channels with very mobile beds, differentially affected flows of various magnitudes and likely played a prominent, and additional, role that further affected the hydrological response.

The Mount St. Helens landscape has proven to be extraordinarily hydrologically resilient. Rapid modifications of surface infiltration, owing principally to physical rather than biogenic effects, widening and stabilization of channel geometries, and coarsening of channel beds reduced runoff, increased flow resistance along channels, and swiftly diminished the magnitudes of posteruption peak flows, leading to an unexpectedly rapid hydrological recovery from the devastating landscape disturbances caused by the eruption.

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