3
Physical Events, Environments, and Geological–Ecological Interactions at Mount St. Helens: March 1980–2004

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3.1 Introduction

The diversity and intensity of volcanic processes during the 1980 eruption of Mount St. Helens affected a variety of ecosystems over a broad area and created an exceptional opportunity to study interactions of geophysical and ecological processes in dynamic landscapes. Within a few hours on the morning of May 18, 1980, a major explosive eruption of Mount St. Helens affected thousands of square kilometers by releasing a massive debris avalanche, a laterally directed volcanic blast, mudflows, pyroclastic flows, and widespread tephra fall (see Figure 1.1; Figures 3.1, 3.2; Table 3.1). These primary physical events killed organisms, removed or buried organic material and soil, and created new terrestrial and aquatic habitats. Despite these profound environmental changes, important legacies of predisturbance ecosystems, including live organisms, propagules, and organic and physical structures, persisted across much of the affected landscape. The physical characteristics of the volcanic processes (elevated temperature, impact force, abrasion, and depth of erosion and burial) in part determined the extent of mortality and the types and significance of biotic legacies in the posteruption landscape.

The primary volcanic events of May 18, 1980 triggered a succession of interacting biological, geological, hydrologic, and anthropogenic changes in the Mount St. Helens area (see Figure 1.2). The initial events altered landforms, watershed hydrology, sediment availability and delivery, and the roles of vegetation and animals in regulating physical and biological processes. Concurrently, hydrologic and geomorphic processes altered the paths and rates of ecological responses by persistently modifying habitats, thereby favoring some species and ecological processes while impeding others. Understanding the interactions between geological and ecological processes is critical for interpreting ecological change in the posteruption landscape.

To set the stage for examining the responses of plants, animals, fungi, microbes, and ecological processes to the major 1980 eruption, we summarize volcanic events immediately preceding the eruption, those associated with that major eruption and subsequent eruptions, and characteristics of the posteruption landscape. We also highlight ecological conditions in the weeks leading up to the major eruption because some conditions at this proximal time scale strongly influenced posteruption ecological responses. In addition, we discuss new environments created by the eruption, variations in climate and hydrology that affected both secondary disturbances and the potential for ecological responses, and the nature and pace of change imposed by secondary disturbances on ecological processes through the first quarter century after the eruption. We explore concepts linking ecological succession to the succession of geomorphic processes that occur in response to large-scale, severe disturbances. Broad geographical and historical contexts in geological and ecological terms leading up to 1980 are described by Swanson et al. (Chapter 2, this volume). Here, we summarize the recent eruptive activity at Mount St. Helens in terms of its relevance to ecological responses. More comprehensive technical discussions of the 1980 eruptive activity can be found elsewhere (Lipman and Mullineaux 1981; Foxworthy and Hill 1982; Major et al. 2005).

3.2 Events of March Through May 18, 1980

Intrusion of magma high into the edifice of Mount St. Helens culminated in the geophysical events that so dramatically affected the landscape on May 18, 1980. The first indication of volcanic unrest occurred only 2 months before the major eruption. After 123 years of quiescence, Mount St. Helens awoke, on March 15, 1980, with a series of small earthquakes (Endo et al. 1981). Earthquakes increased markedly in magnitude (to 4.2) and frequency (30 to 40 per day) on March 20, heralding 2 months of continuous earthquake activity, episodic steam-driven explosions from the summit, and progressive outward growth of the volcano's north flank as magma intruded into the cone (Table 3.2). The injected magma fractured, steepened, and weakened the north flank of the volcano.

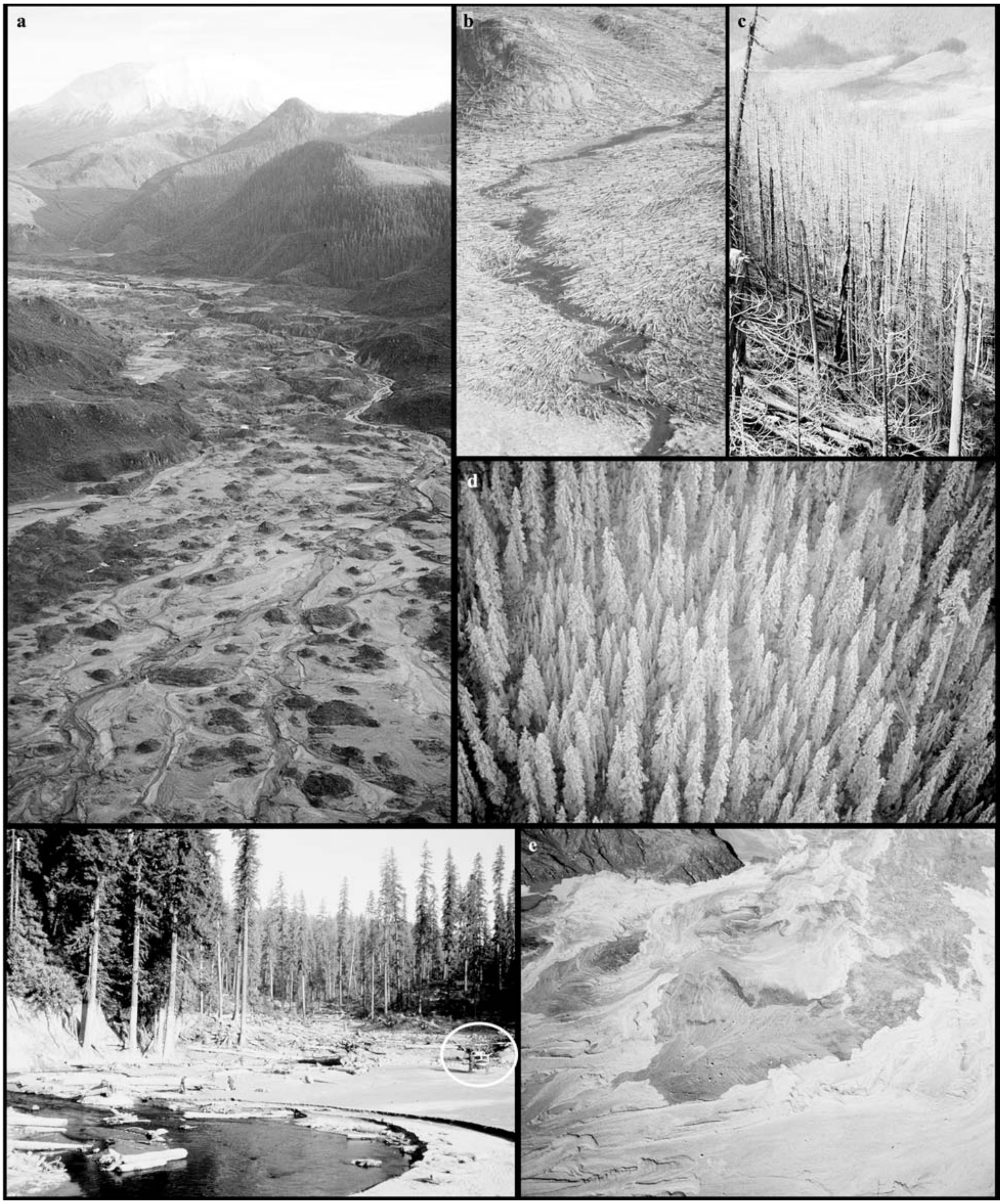

FIGURE 3.1. Photographs of the Mount St. Helens landscape showing conditions produced by the May 18, 1980, eruption: (a) debris-avalanche deposit (North Fork Toutle River valley looking east); (b) blowdown zone with extensive toppled trees (upper Green River valley looking south); (c) scorch zone with standing dead forest (upper Clearwater Creek valley); (d) tephra-fall zone; (e) pyroclastic-flow deposits on the Pumice Plain (light-colored features that flowed from left to right); (f) mudflow zone [Muddy River looking downstream; note helicopter (in *circle*) at right of photo for scale].

TABLE 3.1. Characteristics of volcanic events at Mount St. Helens, May 18, 1980.

Event	Volume of uncompacted deposit (km^3)	Area affected (km^2)	Deposit thickness (m)	Temperature (°C)	Organic matter
Debris avalanche	2.5	60	10–195	70–100	Rare
Blast	0.2	570			
Blowdown zone		370	0.01–1.0	100–300	Common
Scorch zone		110	0.01–0.1	50–250	Common
Mudflows		50	0.1–10	30	Common
Tephra fall	1.1	1000	>0.05	<50	Rare
Pyroclastic flows	0.3	15	0.25–40	300–850	None

Note that the total area of the lateral blast includes zones mapped as tree-removal, blowdown, scorch, debris-avalanche, and pyroclastic-flow zones in plate 1 of Lipman and Mullineaux (1981). The pyroclastic-flow zone occupies a part of the debris-avalanche zone. The area shown for the blowdown zone is for the area mapped as such by Lipman and Mullineaux (1981) and is included in the blowdown zone shown in Figure 3.2. We do not list characteristics of the tree-removal zone because it is a composite of multiple processes, so thickness, temperature, and organic matter are quite varied and difficult to characterize. The tephra-fall zone is considered only for the area outside the blast-affected area, where thickness exceded 5cm.
Source: Various sources in Lipman and Mullineaux (1981).

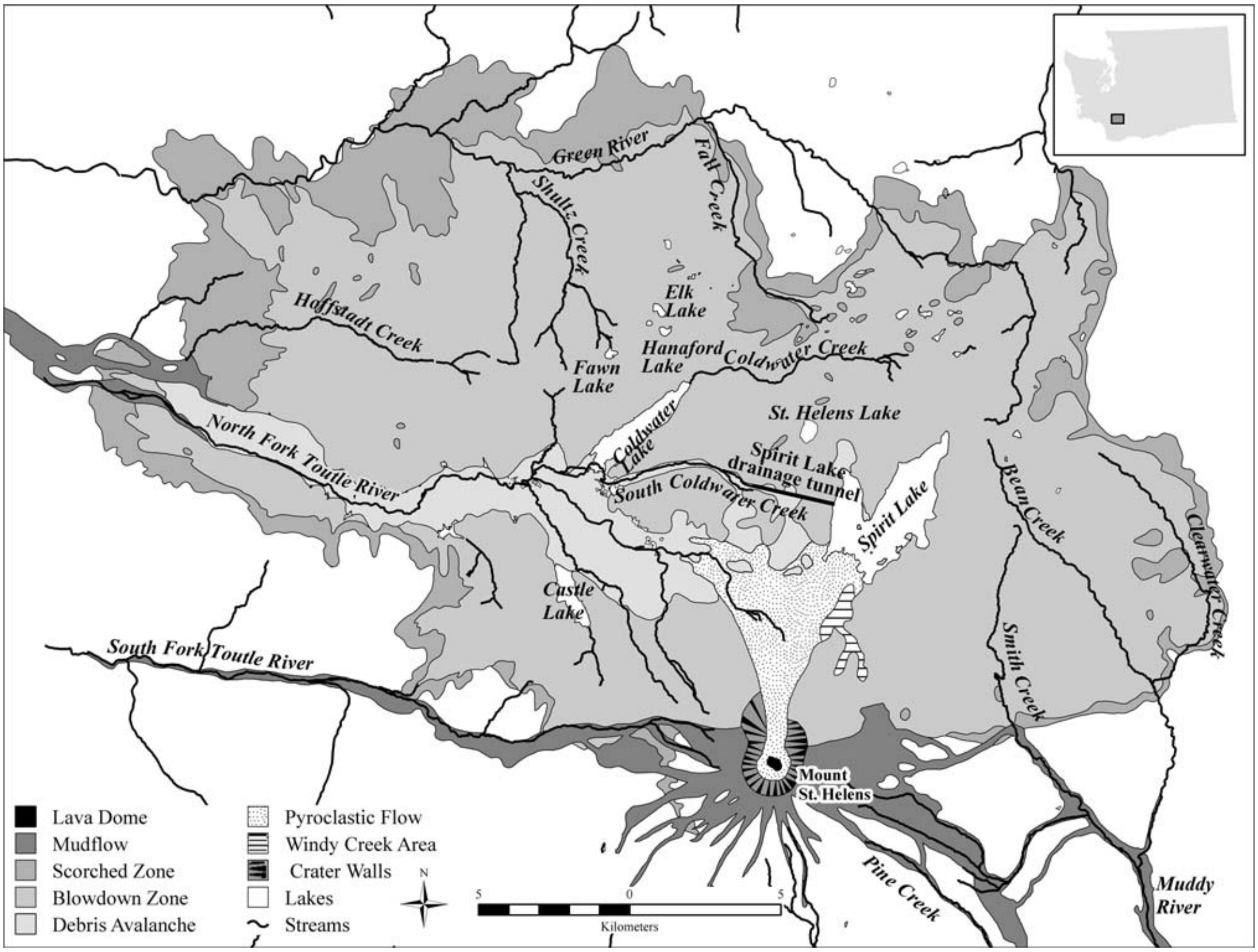

FIGURE 3.2. Distribution of primary volcanic deposits and disturbance zones of the 1980 Mount St. Helens eruptions. [Adapted from plate 1 in Lipman and Mullineaux (1981).]

TABLE 3.2. Chronology and characteristics of eruptive events at Mount St. Helens, 1980 to 2004.

Date	Events
March 27–May 18, 1980	Steam-blast explosions; deformation of north flank by magma injection; very little tephra fall beyond flanks of cone. $V = 0.0006$ km^3.
May 18, 1980	See Table 3.1.
May 25, 1980	Tephra fall greater than 1 cm thick over about 300 km^2 west and northwest of vent. $V = 0.031$ km^3. Pyroclastic flows reach Pumice Plain. $V = 0.001$ km^3.
June 12, 1980	Tephra fall greater than 1 cm thick over about 200 km^2 south. $V = 0.027$ km^3. Pyroclastic flows reach Pumice Plain. $V = 0.01$ km^3.
July to December 1980	Periodic growth of lava dome in crater. Some removal of dome during eruption of July 22.
July 22, 1980	Tephra fall less than 1 cm thick to east and northeast of vent. $V = \sim 0.004$ km^3. Pyroclastic flows reach Pumice Plain. $V = 0.006$ km^3. Partial removal of lava dome.
August 7, 1980	Tephra fall less than 1 cm thick to north and northeast. $V = 0.0008$ km^3. Pyroclastic flows travel more than 5.5 km from vent. $V = 0.004$ km^3.
October 16–18, 1980	Tephra fall less than 1 cm thick to southwest and southeast. $V = 0.0005$ km^3. Pyroclastic flows reach Pumice Plain. $V = 0.001$ km^3.
December 27, 1980 to October 21, 1986	Sixteen dome-growth events, two minor explosions, and three mudflows.
October 22, 1986 to September 2004	No magmatic eruptions; several minor phreatic explosions from dome.
September 2004	Lava dome growth; steam and ash eruptions

V, uncompacted volume of deposits.
Source: From Christiansen and Peterson (1981); Sarna-Wojcicki et al. (1981); Rowley et al. (1981); Swanson et al. (1983a); Brantley and Myers (2000).

Ecological and hydrologic conditions leading up to the morning of May 18 profoundly affected ecological responses to the eruption. For example, low- to mid-elevation (less than 1000 m above sea level) forests and meadows surrounding the volcano contained discontinuous patches of late-lying snow; but above about 1000 m, snow generally covered the terrain. That snow cover provided some protection from the eruption for terrestrial and aquatic organisms.

3.2.1 Debris Avalanche

The May 18, 1980, eruption commenced with a magnitude 5.1 earthquake and an associated collapse of the volcano's north flank at 8:32 A.M. (Voight et al. 1981, 1983). The resulting 2.5-km^3 debris avalanche, the largest landslide in recorded history, rushed northward in multiple pulses that broadly split into three lobes (Lipman and Mullineaux 1981; Voight et al. 1981; Glicken 1998) (see Figures 3.1a, 3.2). One lobe entered and passed through Spirit Lake, where it generated a seiche (an oscillating wave) that extended as high as 260 m above the pre-1980 lake level and raised the level of the lake by 60 m. A second lobe traveled northward 7 km at a velocity of 50 to 70 m s^{-1} (Voight 1981) and overtopped a 300- to 380-m-high ridge (later named Johnston Ridge). The bulk of the avalanche, however, moved 23 km westward down the North Fork Toutle River valley in about 10 minutes.

The resulting debris-avalanche deposit radically modified the upper North Fork Toutle River valley. It buried about 60 km^2 of the valley with hummocky, poorly sorted sand and gravel to a mean depth of 45 m (Voight et al. 1981; Glicken 1998; see Figures 3.1a, 3.2; see Table 3.1) and disrupted the drainage pattern (Lehre et al. 1983; Janda et al. 1984). The original locations of rock units and glaciers on the volcano and the flow paths of the various debris-avalanche pulses determined the distribution of rocks, soil, and ice in the heterogeneous avalanche deposit (Glicken 1998). The avalanche pulses that surged down the North Fork Toutle River valley smeared avalanche debris that contained little organic matter, a few toppled trees, and blocks of soil along the valley margin. The front of the debris avalanche scoured a forest from the valley floor and walls and left some of it as a tangled mass of vegetation at the downstream end of the debris-avalanche deposit. The margins of the upper half of the deposit chiefly comprise material from the first slide block that engulfed the outer flank of the volcano, which contained glacier ice and organic matter. That debris was emplaced at ambient temperature. In contrast, the central part, as well as the lower half, of the avalanche deposit comprises material that came mostly from deeper within the core of the volcano. That material contained negligible organic matter and a greater abundance of hot rock from within or near the intruded magma body. Temperatures were 70° to 100°C at depths of 1 to 1.5 m on the central part of the avalanche deposit 10 to 12 days after emplacement (Banks and Hoblitt 1981).

3.2.2 Directed Blast

Collapse of the volcano's north flank suddenly decompressed the shallow magma body within the volcano and the superheated groundwater circulating near it (similar to opening the vent on a hot pressure cooker) and triggered a devastating volcanic blast. The sudden decompression initiated both phreatic (steam-driven) and magmatic (exsolving gas-driven) explosions that propelled fragmented debris outward from the volcano until gravity caused the turbulent cloud of rock, ash, and gas to flow across the landscape (Hoblitt et al. 1981; Moore and Sisson 1981; Waitt 1981; Waitt and Dzurisin 1981). The blast followed the debris avalanche off the volcano but rapidly outran it. Along the shores of Spirit Lake and in the upper North Fork Toutle River valley, the blast raced ahead of the avalanche and toppled mature forest. The dense debris avalanche hugged the valley floor; but the lower-density, more-energetic blast flowed over the rugged topography it encountered and spread across a wide area.

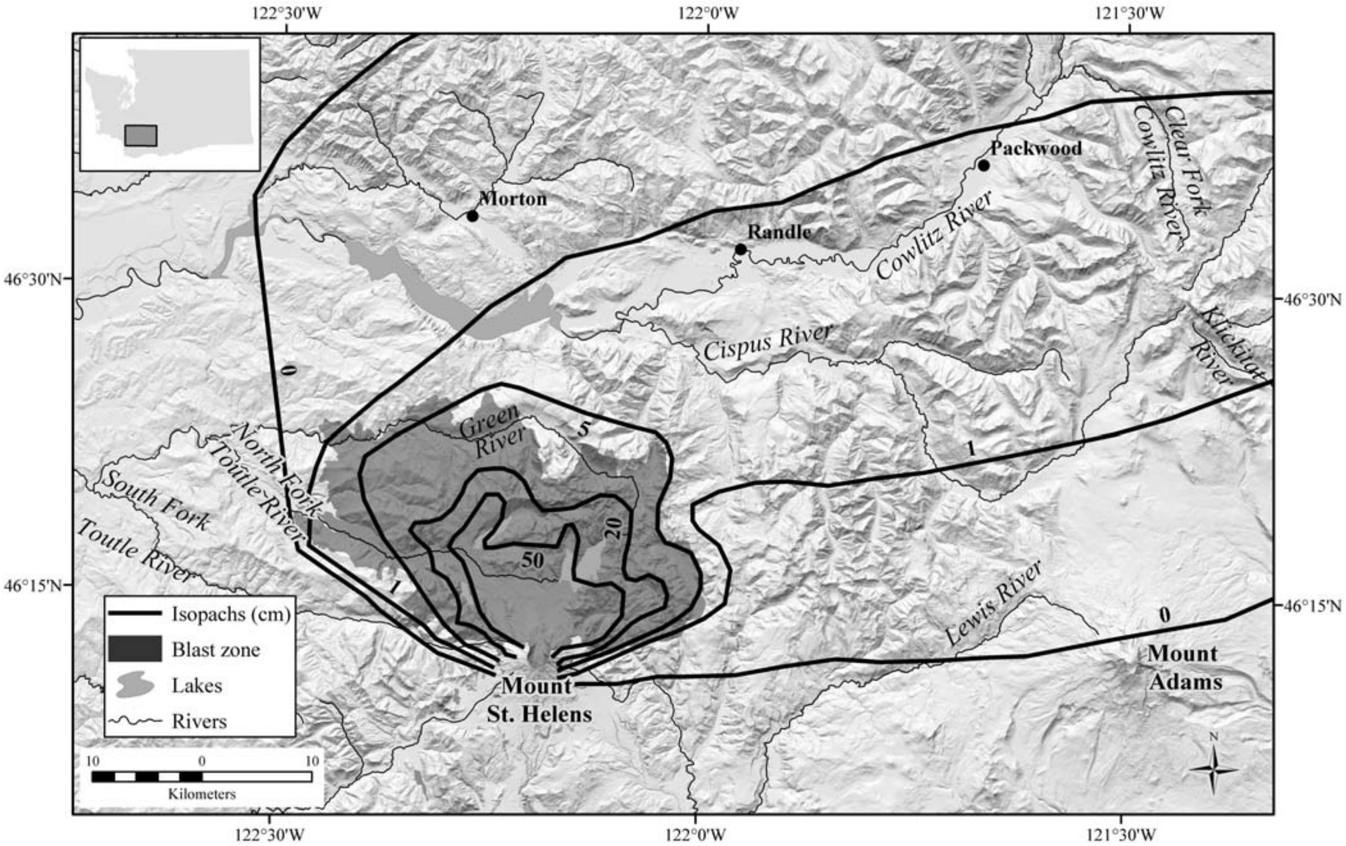

FIGURE 3.3. Thickness (in centimeters) map of blast deposit. (Isopachs are contours of equal deposit thickness.) Some of the blast deposit extends beyond the boundary of the blast area because fine material lofted high into the atmosphere drifted east with the wind. [Adapted from Figure 261 in Waitt (1981).]

The blast produced a hot cloud, charged with 0.2 km^3 of rock debris (see Table 3.1), that removed, toppled, or scorched most aboveground vegetation over an area of about 570 km^2 in a 180° arc north of the mountain (see Figure 3.2) (Moore and Sisson 1981; Waitt 1981). The area affected by the blast can be subdivided into three primary zones (see Table 3.1). Close to the volcano, the blast, the debris avalanche, and the avalanche-triggered seiche in Spirit Lake stripped most trees from the landscape. This area is called the tree-removal zone (Lipman and Mullineaux 1981, plate 1). The impact force of the blast diminished with increasing distance from the volcano, so it left progressively more of the trunk and limb structure of trees intact toward the margin of the zone. Beyond the tree-removal zone, the blast created a 370-km^2 blowdown zone of the blast-toppled forest (see Figures 3.1b, 3.2). Yet farther from the vent, the blast no longer had the force to topple trees, but instead left a 0.3- to 3.0-km-wide scorch zone, a 110-km^2 swath of standing dead forest killed by the heat of the blast cloud (see Figures 3.1c, 3.2). Our terminology of these zones differs slightly from that used by Lipman and Mullineaux (1981, plate 1).

The blast blanketed the landscape with a deposit that was progressively thinner and finer textured with increasing distance from the volcano (Figure 3.3). These deposits blanketed the tree-removal zone with 0.2 to 1.5 m of pebbly to sandy gravel overlain by sand, covered the blowdown zone with 0.1 to 1.0 m of sandy gravel to sandy silt, and draped 0.01 to 0.1 m of coarse sand to sandy silt over the scorch zone (Hoblitt et al. 1981; Waitt 1981). Within the tree-removal and blowdown zones, blast debris deposited on slopes in excess of 35° was unstable. On such steep slopes, the deposit commonly remobilized and formed secondary "blast-pyroclastic flows" that slid down hillsides and collected in stream valleys close to the volcano. The resulting secondary deposits are up to 10 m thick, and locally contain abundant organic matter stripped from the hill slopes. Emplacement temperature of these deposits ranged from 100° to 300°C (Banks and Hobblitt 1981).

The hot, rock-laden blast abraded, burned, and singed vegetation. Initial temperature of the blast deposit varied with the abundance of fragments of the hot magma, but it generally ranged from about 100° to 300°C (Banks and Hoblitt 1981; Moore and Sisson 1981). The hottest parts of the blast, typically northeast of the volcano, entrained and charred wood fragments (Moore and Sisson 1981). In the scorch zone, the blast

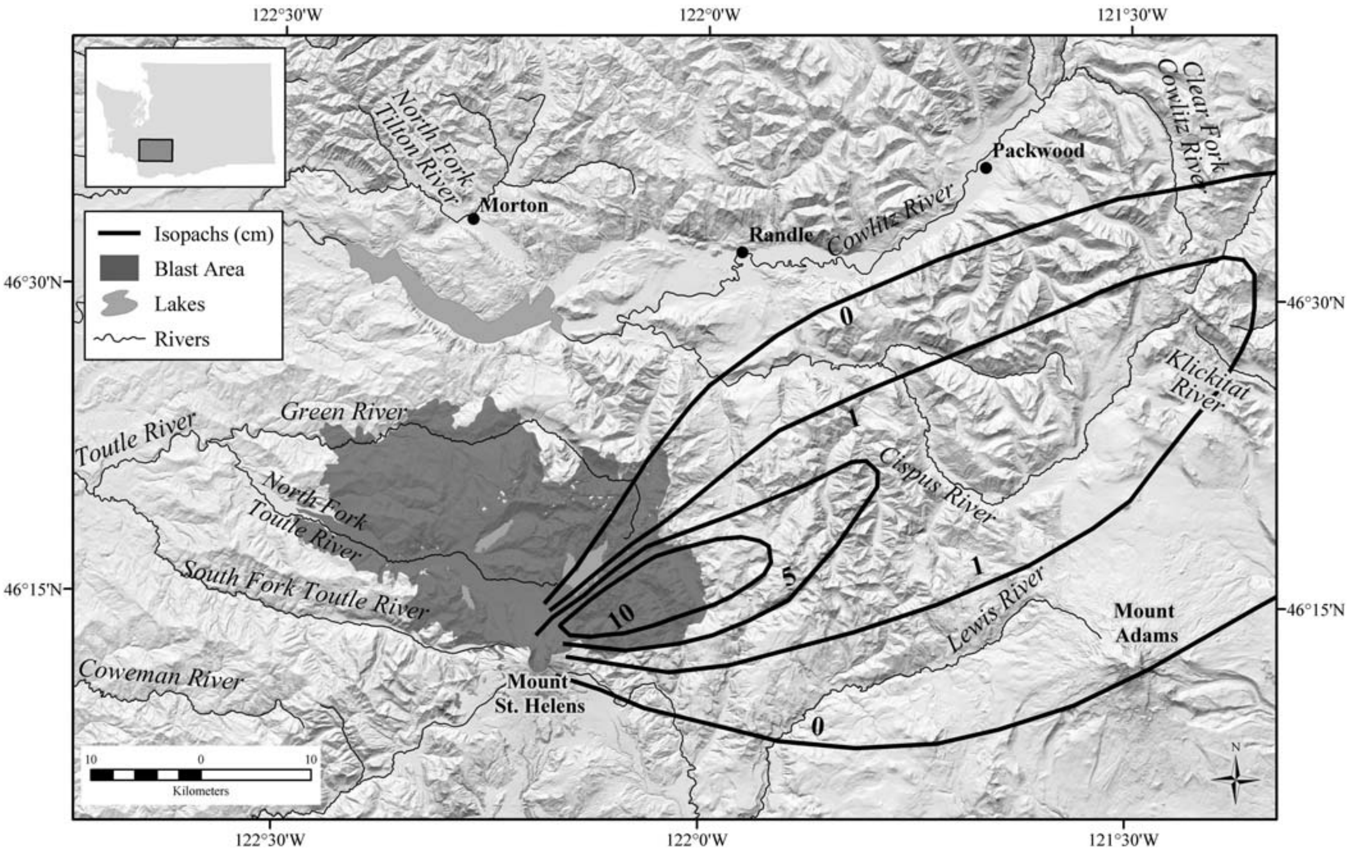

FIGURE 3.4. Thickness (in centimeters) map of May 18, 1980, tephra-fall deposit associated with vertical plume that developed after onset of eruption. This map does not include tephra falls associated with the blast or with post-May 18, 1980 eruptions. [Adapted from Waitt and Dzurisin (1981), Figures 358, 361.]

elevated air temperature to about 50°C (Winner and Casadevall 1981) and singed foliage, but it did not char the vegetation. Despite the high temperature of the blast, there is little evidence that it appreciably elevated the temperature of the buried preeruption soil.

3.2.3 Tephra Fall

Tephra fall caused the most widely distributed impact of the eruption. A billowing, vertical plume grew within minutes of the onset of the eruption, continued for about 9 hours, and ejected about 1.1 km^3 of tephra (calculated as uncompacted material) into the atmosphere (Sarna-Wojcicki et al. 1981) (see Figure 3.1d; see Table 3.1). Wind blew the tephra mainly to the east-northeast (Figure 3.4).

Tephra-fall texture and thickness vary greatly within the fallout zone (Figure 3.4). Pebble-sized pumice fell close to the mountain and formed a layer greater than 20 cm thick over an area of about 16 km^2, much of it within the area affected by the blast only hours earlier. Granule- to sand-sized tephra fell onto and through the forest canopy across hundreds of square kilometers northeast of the volcano to thicknesses of several centimeters. A silty layer as much as a few centimeters thick blanketed the entire area of tephra accumulation, including much of the blast area. Emplacement temperature of the tephra fall beyond the blast-affected area was apparently less than 50°C (Winner and Casadevall 1981). Vegetation beyond the limit of the blast area was covered by cool tephra fall.

3.2.4 Pyroclastic Flows

Subsequent to the debris avalanche and blast, pyroclastic flows inundated the upper North Fork Toutle River valley and Spirit Lake basin and locally spilled onto the flanks of the volcano. Pumice-rich pyroclastic flows spewed from the eruption vent for 5 hours, beginning about noon on May 18 (Rowley et al. 1981; Criswell 1987). These flows covered about 15 km^2 of the surface of the debris-avalanche deposit immediately north of the volcano and deposited approximately 0.3 km^3 of hot, loose, pumiceous sediment (see Figures 3.1e, 3.2; see Table 3.1). The inundated area is named the Pumice Plain to denote the composition and flat surface of these deposits. Individual pyroclastic-flow deposits ranged from 0.25 to 10 m thick; the total accumulation is as much as 40 m thick (Criswell 1987). Cobbles and boulders of pumice, and fragments of denser rock, cover the surfaces of these deposits, but their cores comprise

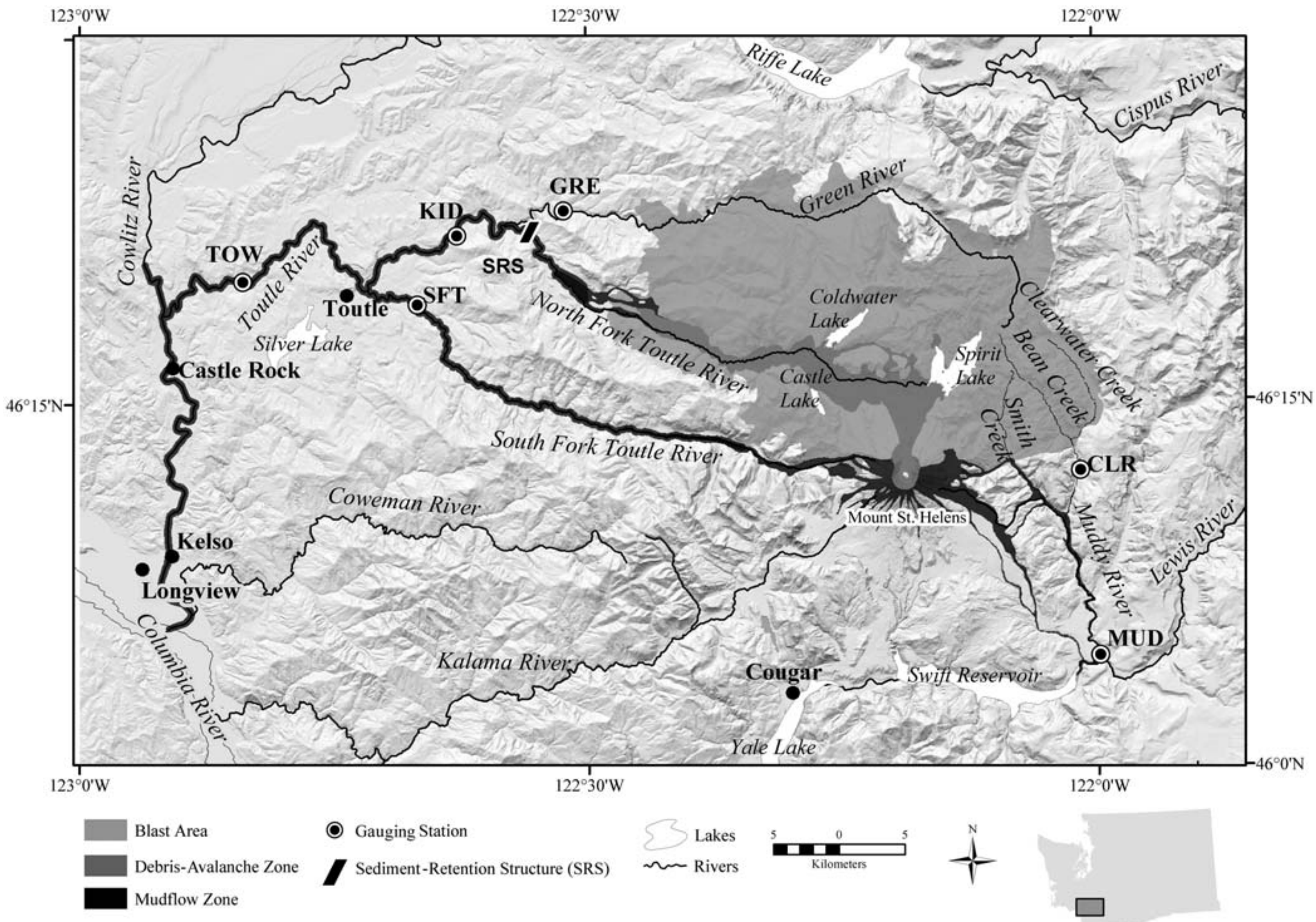

FIGURE 3.5. Locations of stream gauging stations along the lower Toutle River (*TOW*), North Fork Toutle River (*KID*), South Fork Toutle River (*SFT*), Muddy River (*MUD*), Green River (*GRE*), and Clearwater Creek (*CLR*). *SRS*, sediment-retention structure on the North Fork Toutle River.

gravelly sand (Kuntz et al. 1981). Emplacement temperatures of pyroclastic flows on the Pumice Plain ranged from about 300° to 730°C but were as much as 850°C close to the vent (Banks and Hoblitt 1981).

3.2.5 Mudflows

Eruption-triggered mudflows (commonly termed lahars) inundated the major channels that drained the volcano and flowed long distances downstream (see Figures 3.1f, 3.2; Figure 3.5). The debris-avalanche, blast, and pyroclastic flows generated the mudflows in several ways (Janda et al. 1981; Pierson 1985; Major and Voight 1986; Fairchild 1987; Scott 1988; Waitt 1989). The largest and most destructive mudflow on May 18 emanated from the debris-avalanche deposit and flowed down the North Fork Toutle River valley. That mudflow resulted from liquefaction induced by ground shaking (Fairchild 1987) and from muddy slurries produced by consolidation and slumping of water-saturated parts of the avalanche deposit (Janda et al. 1981; Glicken 1998). The combination of the great volume of this mudflow (1.4×10^8 m^3; Fairchild and Wigmosta 1983), more than 10 times larger than any of the other mudflows of May 18 (Major et al. 2005), and the relative confinement of the Toutle River valley permitted it to flow 120 km to the Columbia River (see Figure 3.5). Along its path, this mudflow extensively inundated the floodplain and entrained riparian vegetation and piles of logs from logging camps, which added to its volume and destructive force.

Parts of the blast cloud eroded and melted snow and ice on the west, south, and east flanks of the volcano, producing large (to 10^7 m^3), rapidly moving mudflows. Blast-triggered mudflows traveled tens of kilometers along the channels of Smith Creek, Muddy River, Pine Creek, and the South Fork Toutle River (see Figures 3.2, 3.5). Some of those flows gradually transformed into more-dilute, sediment-laden floods as they moved down the valley (Pierson 1985; Scott 1988). The blast also initiated small, thin, unchannelized mudflows on the broad southern side of the volcano (Figure 3.2), which inundated alpine and subalpine areas, such as Butte Camp and the upper Pine Creek fan (Fink et al. 1981; Major and Voight 1986).

Relatively small pumiceous pyroclastic flows that spilled onto the east and west flanks of the volcano during the afternoon of May 18 melted ice and snow and produced additional moderate-sized mudflows in the headwaters of Smith Creek, Muddy River, and South Fork Toutle River (Janda et al. 1981; Pierson 1985; Scott 1988). These pumice-rich mudflows were smaller (up to 10^5 m^3) than the blast- and debris-avalanche-triggered mudflows initiated in the morning, and they traveled along channels that had been extensively modified by the earlier, larger flows.

Dynamic characteristics of mudflows, such as volume, velocity, peak discharge, and impact force, generally varied among mudflows of different origin and decreased as flows traveled downstream and interacted with channel structures and riparian vegetation. The mudflows were anywhere from a few meters to more than 10 m deep as they flowed away from the volcano (see Figure 3.1f), but they left deposits of gravelly sand that generally were less than 1 m thick along most valleys. Some mudflows had slightly elevated temperatures and were described as having textures similar to "warm concrete" (Cummans 1981; Rosenbaum and Waitt 1981). Mudflows generally toppled and severely abraded vegetation close to the channel (see Figure 3.1d), but more tranquil passage through riparian forest farther from a channel axis resulted chiefly in deposition of a thin veneer of gravelly sand that caused little immediate tree mortality (Janda et al. 1981; Pierson 1985; Scott 1988; Frenzen et al., Chapter 6, this volume).

3.3 Volcanic Events Since May 18, 1980

Between May 18, 1980, and 1991, Mount St. Helens erupted an additional 21 times (Table 3.2) (Swanson et al. 1983a; Brantley and Myers 2000). These eruptions generally had minimal impact on the landscape and the biological response to the major eruption, except in the area immediately north of the crater. Five eruptions between May 25 and October 16, 1980, produced pyroclastic flows (Rowley et al. 1981) and tephra falls. None of these eruptions produced pyroclastic flows exceeding 0.01 km^3 (5% of the volume of the pyroclastic flows erupted on May 18), and the resulting deposits generally blanketed those of the May 18 eruption (Rowley et al. 1981). Post-May 18 tephra-fall deposits are generally less than 2 cm thick (Sarna-Wojcicki et al. 1981) and had little additional effect on the landscape. The volcano was largely snow free during these eruptions, and no major mudflows developed.

Eruptions between October 1980 and October 1986 involved predominantly nonexplosive growth of a lava dome within the volcano's crater, which by 1987 was about 270 m tall and had a volume of about 7.5×10^7 m^3 (Brantley and Myers 2000). Explosions during periods of dome growth between 1982 and 1986 commonly coincided with thick snowpacks in the crater, and some explosions led to rapid snowmelt and generation of mudflows, the most notable occurring on March 19, 1982 (Waitt et al. 1983; Cameron and Pringle 1990; Pierson 1999; Pringle and Cameron 1999). All these mudflows passed along the North Fork Toutle River and were gradually diluted to sediment-laden streamflow, but they were far less extensive than the mudflows of May 18, 1980.

The magnitude and intensity of eruptions at Mount St. Helens waned rapidly following the May 18, 1980, eruption. Significant magmatic explosions were largely complete by October 1980, and this phase of dome growth ended by October 1986. Between 1986 and 1991, the most significant eruptive activity involved minor steam explosions driven chiefly by rainfall seeping into fractures on the dome and contacting hot rock (Mastin 1994). Until dome growth commenced in 2004 no eruptions or explosions occurred after 1992. As eruptive activity declined in the mid 1980s, a glacier accumulated in the crater south of the lava dome (Anderson and Vining 1999; Schilling et al. 2004). As of 2001, the crater glacier had an ice volume (about 80×10^6 m^3) equivalent to 40% to 50% of the volume of glacier ice on the preeruption volcano.

3.4 Hydrology: Precipitation and Runoff

Temporal variation in climate and water runoff influenced both biological processes directly and the environments of posteruption biological succession. Such variations also affected geomorphic processes driven by rainfall and snowmelt runoff. Processes such as sheet and rill erosion, small landslides (debris slides), and lateral and vertical channel erosion dominated posteruption physical changes to the landscape, and they greatly affected the path and pace of biological succession.

Histories of regional precipitation and runoff provide a context for interpreting the physical and biological changes that have occurred on hill slopes and in channels between 1980 and 2000. Interdecadal periods of wetter- and drier-than-average conditions (Mantua et al. 1997; Biondi et al. 2001) were punctuated by higher-frequency El Niño–Southern Oscillation climate variability (McCabe and Dettinger 1999). Relatively dry conditions characteristic of the time of the 1980 eruption persisted for about 15 years (Mantua et al. 1997; Biondi et al. 2001), followed in the mid- to late 1990s by wetter-than-average conditions (Major et al. 2000; Major 2004). From 1995 to 2000, mean annual streamflow from basins near Mount St. Helens was 40% to 50% greater than during the period 1980 to 1994 (Figure 3.6b). This climatic variation in streamflow had a marked impact on erosion and sediment transport from basins affected by the 1980 eruption (Major et al. 2000; Major 2004).

Posteruption streamflow from the Mount St. Helens landscape reflects great changes in precipitation–runoff relationships, structural modifications to drainage networks, and reduction and regrowth of vegetation. Storms and snowmelt of similar magnitudes before and after May 18, 1980 produced substantially higher runoff after the eruption, as detected through comparisons of pre- and posteruption streamflow (Major et al. 2001). By comparing magnitudes and frequencies

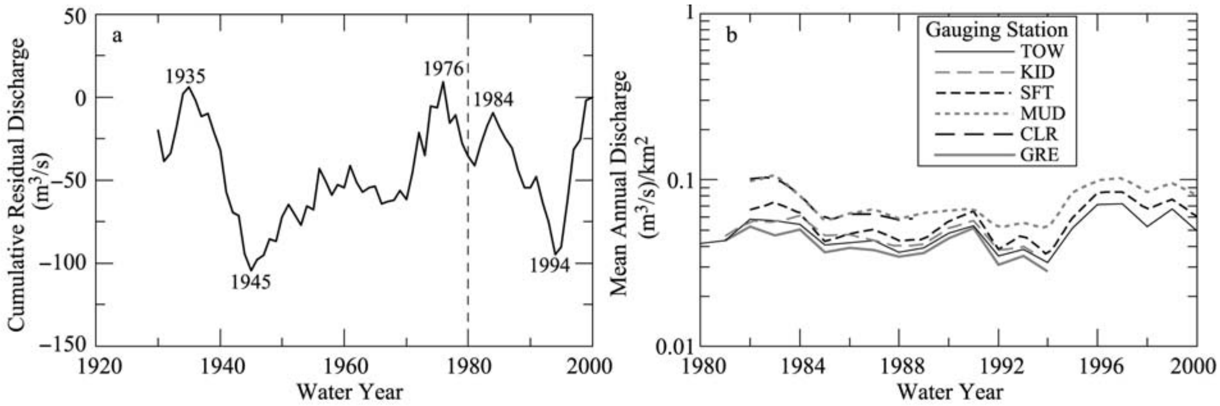

FIGURE 3.6. (a) Time series of the cumulative sum of annual deviations from long-term mean annual flow of Toutle River near TOW (see Figure 3.5 for station locations). A downward trend denotes periods of drier-than-average conditions; an upward trend denotes periods of wetter-than-average conditions. (b) Time series of posteruption mean annual unit area discharges for rivers at Mount St. Helens. [Adapted from Major (2004).]

of unit-area peak-flow discharges from basins affected by the 1980 eruptions with those of a nearby, relatively unaffected basin, Major et al. (2001) concluded:

- The major volcanic disturbances increased peak discharges over a broad range of magnitudes, on average by several tens of percent, for about 5 years;
- Streamflow from basins affected only by the blast or tephra fall (without mudflows) recovered to preeruption magnitudes within a few years, whereas basins having severe channel disturbances in addition to hill-slope disturbance recovered more slowly; and
- Streamflow from some basins that suffered severe channel disturbance displayed responses to the eruption even as late as 2000.

Dramatic increases in magnitude of peak flows immediately after the 1980 eruption resulted from changes to various conditions of the land. First, loss of canopy interception as a result of loss of the forest cover allowed more precipitation to reach the land surface and reduced evaporation. It also altered seasonal variation in soil moisture by reducing transpiration. Second, meager surface infiltration of the blast and tephra-fall deposits radically altered hill-slope hydrology and changed it from a regime that lacked overland flow to one dominated by it. Infiltration capacity of the ground surface decreased from about 75 to 100 mm h^{-1}, typical of forest soils in the region (Johnson and Beschta 1980), to less than 10 mm h^{-1} (Swanson et al. 1983b; Leavesley et al. 1989). In the blowdown zone about 20 km northwest of the volcano, the infiltration capacity of a silty crust developed on the blast deposit was as little as 2 to 5 mm h^{-1} in August 1980, and it only roughly doubled by the following summer (Leavesley et al. 1989). Salvage-logging operations, burrowing animals, trampling by large mammals, formation and melting of soil ice, and growth of vegetation partly mixed the tephra profile in the first few years after deposition (Collins and Dunne 1988). These processes, combined with local erosion, increased infiltration capacity (Swanson et al. 1983b; Collins and Dunne 1986, 1988). Parts of the blowdown zone deliberately disturbed by scarification and salvage logging had infiltration capacities as great as 28 mm h^{-1} by 1981 (Fiksdal 1981). As of 1998, measured infiltration rates of slopes in the blowdown zone northwest of Mount St. Helens that were not deliberately disturbed were locally as much as 25 mm h^{-1} (Major and Yamakoshi in press), suggesting that posteruption infiltration rates after nearly two decades have increased greatly without intervention but are still much less than likely preeruption rates. Consequently, common rainfall intensities (e.g., 10 mm h^{-1}, a 1-hour-duration storm having an approximately 2-year return interval) produce little surface runoff; but locally, lower-frequency, higher-magnitude storms can still induce appreciable surface runoff. Third, forest toppling by the blast altered the timing and rate of snow accumulation and melt during rain-on-snow events (Dunne and Leopold 1981; Orwig and Mathison 1981; Lettenmaier and Burges 1981; Janda et al. 1984; Simon 1999), similar to alterations observed following forest cutting (Harr 1981; Marks et al. 1998). Fourth, channel changes caused by the eruption altered the manner in which runoff moved downstream. Channel smoothing and straightening temporarily reduced flow resistance and allowed stormflow to travel faster and be attenuated less (Janda et al. 1984; Simon 1999). The alterations to both hill-slope and channel geomorphology, combined with enhanced sediment transport that increased flow volumes and damped turbulence (Janda et al. 1984), enhanced peak-flow discharges from many parts of the disturbed landscape.

In some parts of the landscape, the 1980 disturbances temporarily diminished water runoff. The Pumice Plain, for example, comprises highly permeable pyroclastic-flow deposits that produce virtually no surface runoff. In the upper North Fork Toutle River valley, the debris-avalanche deposit blocked several tributary channels and had an extensive, irregular surface

Table 3.3. Characteristics of river and riparian environments affected by the 1980 activity of Mount St. Helens.

Disturbance type[a] (example)	Initial disturbance of riparian vegetation	Initial change in channel complexity	Stability of channel location	Change in sediment load	
				Magnitude	Duration
Debris avalanche (North. Fork Toutle River)	Complete	Complete	Low	Very high	Many decades
Mudflow below debris avalanche (North Fork Toutle River)	Extensive; some sprouting of residual trees	Roughness reduced by scour and deposition	Low	Very high	Many decades
Mudflow (Pine Creek)	Extensive; some sprouting of residual trees	Roughness reduced by scour and deposition	Moderate	Moderate	Years
Blast area (Upper Green River; Bean Creek)	Extensive; some sprouting of residual shrubs and herbs	Increased by fallen trees; decreased by debris-flow scour and mass-movement sediment deposition	High	Moderate	Years
Downstream of blast area (Lower Clearwater Creek)	Minor	Minor	High	Moderate	Years
Tephra fall (Upper Clear Creek)	Minor	None	High	Low	Months
Downstream of tephra-fall zone (Lower Clear Creek)	Minor	None	High	Low	Months

Magnitude of "change in sediment load" ranges from low (less than 100% increase) to very high (more than an order of magnitude increase) (see Figure 3.8).
[a]Disturbance types are described in text.

composed of closed depressions that contained ponds. Lakes and ponds that formed adjacent to, and on the surface of, the debris-avalanche deposit trapped local runoff. Some lakes and ponds continue to trap runoff, whereas others filled and breached their impoundments and helped reconnect a drainage network across the deposit. It took more than 2 years before a drainage network reconnected headwater tributaries to the main channel of the upper North Fork Toutle River (Meyer 1995; Simon 1999). Obliteration of the stream network in the upper North Fork Toutle River valley by the debris-avalanche deposit partly counteracted landscape changes that enhanced surface runoff. Once the drainage network reintegrated sufficiently, posteruption peak-flow discharges of the North Fork Toutle River ranged from a few to several tens of percent larger than preeruption discharges of comparable frequency.

3.5 Environments Resulting from the 1980 Activity

The primary volcanic deposits created distinctive environments that posteruption processes modified at various rates (Collins and Dunne 1986; Smith and Swanson 1987; Meyer and Martinson 1989). Ecological responses to the 1980 eruptions depended, in part, on the character of these environments and on the type, intensity, and extent of modification of the primary deposits by secondary hydrologic and geomorphic events. The following discussion focuses on environments grouped as terrestrial, riparian, and riverine; lake and lakeshore; and hydrothermal.

3.5.1 Terrestrial, Riparian, and Riverine Environments

The degrees of primary and secondary landform modification and their impacts on biological legacies of the pre-1980 ecosystems varied among terrestrial, riparian, and riverine environments. Primary volcanic disturbances affected areas that are large relative to pathlengths of water and sediment transport down hill slopes, so physical processes in upland terrestrial environments respond almost exclusively to a specific primary volcanic disturbance. Rivers, on the other hand, transport water and sediment long distances and may flow through a variety of disturbance zones (see Figure 3.2). Thus, riverine and riparian environments respond to local disturbances as well as to primary and secondary disturbances farther upstream (Table 3.3).

3.5.1.1 Tephra-Fall Zone Outside the Blast-Affected Area

Biological legacies in the forms of residual organisms and organic structures, such as large logs on the ground, remained as significant factors influencing hydrology, erosion, and revegetation on tephra-mantled hill slopes beyond the blast area principally because of the limited thickness of tephra-fall deposits (see Figure 3.4), the low intensity of initial disturbance, and the modest intensity and short duration of secondary erosion. The thin tephra deposit did not obscure small-scale landforms, such as decomposed logs and root-throw pit-and-mound topography, but it damaged herbaceous and other ground cover while having little effect on trees and shrubs (see Antos and Zobel, Chapter 4, this volume). Minor sheet and rill erosion occurred on slopes steeper than approximately 30° (Swanson et al. 1983b), but the rate of erosion there was less than that observed within the blast area because litter derived from the still-living forest canopy mixed with tephra during deposition and also quickly covered the ground surface after deposition. This organic litter stabilized surfaces by reducing raindrop impact and increasing surface roughness, which fostered infiltration and reduced surface runoff. Rates of surface erosion within areas affected by tephra fall probably peaked and declined quickly over the first wet season, similar to rates of erosion documented in the blast-affected area (Swanson et al. 1983b; Collins and Dunne 1986; Antos and Zobel, Chapter 4, this volume).

Tephra fall had only minor effects on channel and riparian zones (see Table 3.3). Within a few years after the 1980 eruption, scattered patches of pumice gravel that had been deposited on bars and in the lee of obstructions provided the only obvious evidence for tephra fall within stream channels (Smith and Swanson 1987; Lisle 1995).

3.5.1.2 Blast Area: Blowdown and Scorch Zones

Posteruption environments in the blast area vary in the amount and condition of biological legacies, but many secondary erosion processes were common across the three zones affected by the blast. Many biological legacies survived in the blowdown and scorch zones because the blast deposit was relatively thin (less than 1 m). In the tree-removal zone, however, the blast removed most soil and vegetation.

The rugged topography of the preeruption landscape and the abundant trees toppled by the blast strongly influenced patterns of posteruption erosion in the blowdown zone. Sheet, rill, and gully erosion reworked the 1980 deposits in this zone (Swanson et al. 1983b; Collins and Dunne 1986; Smith and Swanson 1987). Unchannelized, shallow, linear depressions were gullied during the first significant posteruption rains in late 1980. Toppled trees created complex surface topography that obstructed runoff, encouraged infiltration, and reduced surface erosion. Posteruption vegetation development did not affect the rapid decline of hill-slope erosion that ensued within months of the eruption, even where artificial seeding was implemented (Stroh and Oyler 1981; Dale et al., Chapter 19, this volume). Surface erosion peaked and declined in response to physical factors before significant vegetation cover established on the blast deposit (Collins and Dunne 1986).

The blowdown and scorch zones experienced a common suite of secondary erosion processes whose rates diminished over time and differed between the two zones. In the blowdown zone, sheet and rill erosion were greater on steeper slopes and on slopes blanketed with deposits having fine-textured surface layers (Swanson et al. 1983b; Collins and Dunne 1986; Smith and Swanson 1987). Fresh tephra of blast and fall origins slid or was washed quickly from slopes steeper than about 35°. Collins and Dunne (1986) measured as much as 26 mm of erosion (computed as average landscape lowering) by 1981, but only an additional 1.8 mm occurred by 1982. The dramatic decline in sheet and rill erosion resulted from development of an armor layer of coarse particles, which followed disruption and removal of the surficial silty tephra. This rough armor layer increased water infiltration and reduced the magnitude and frequency of overland flow (Swanson et al. 1983b; Collins and Dunne 1986). Less than 15% of the tephra deposited in 1980 was removed before hill-slope erosion in the blowdown zone stabilized and geomorphic processes became more like those typical of forested areas (Collins and Dunne 1986). The rate of surface erosion in the scorch zone was generally less than 10% of that measured in the blowdown zone because litter from the scorched forest formed a protective layer and the thin tephra deposits retained forest-floor topography that disrupted surface erosion (Collins and Dunne 1986).

Temporal changes in a clear-cut area east of upper Smith Creek illustrate the general pattern of hill-slope erosion and vegetation development characteristic of natural recovery processes in the blowdown zone (Figure 3.7). Before the eruption, the site bore only stumps and a few logs left from cutting. After the eruption, the site was buried under 50 cm of blast and subsequent tephra-fall deposits. Between May 18 and early fall of 1980, minimal rainfall caused only minor erosion (Figure 3.7a). However, by January 1981 storm runoff had cut gullies into topographic depressions (Figure 3.7b). By then, however, gully erosion had ceased because the tephra surface had coarsened and reduced runoff, and the gullies had eroded down to the preeruption soil. Within 4 years, vegetation that included willow (*Salix* spp.) and perennial herbs colonized or resprouted from perennial rootstocks in the gully floors (Figure 3.7c); within 15 years, vegetation flourished, and Douglas-fir (*Pseudotsuga menziesii*) and several other species of trees and shrubs had colonized areas between gullies (Figure 3.7d).

In addition to sheet and rill erosion, small landslides also stripped tephra and soil from posteruption hill slopes. Numerous small, rapid landslides (100 to 100,000 m^3) occurred in steep areas of the blowdown and scorch zones, mainly within 5 years of the 1980 eruption and again in 1996. Field investigations and analyses of aerial photographs documented 278 slides that occurred between 1980 and 1984 over a 117-km^2 area within the watersheds of Smith Creek, Bean Creek, Clearwater Creek, and upper Green River (Swanson et al. 1983b; Swanson 1986). Landslide scars occupied less than 1% of the study area, but locally they covered up to 10% of the hill slopes and scoured as much as 30% of the channel lengths in these basins. Of the landslides inventoried in this area, 70% occurred in blast-toppled forest; the remainder occurred in areas roaded or clear-cut before 1980 or salvage logged after the eruption. An important cause of sliding may have been the loss of soil anchoring provided by tree roots, roots that had been pulled from the ground during the blast. Between 1980 and 1984, the frequency of slides in the blowdown zone was 35 times greater than that in undisturbed, forested areas and 9 times greater than that in areas clear-cut elsewhere in the Cascade Range (Swanson et al. 1981; Sidle et al. 1985). Many of these

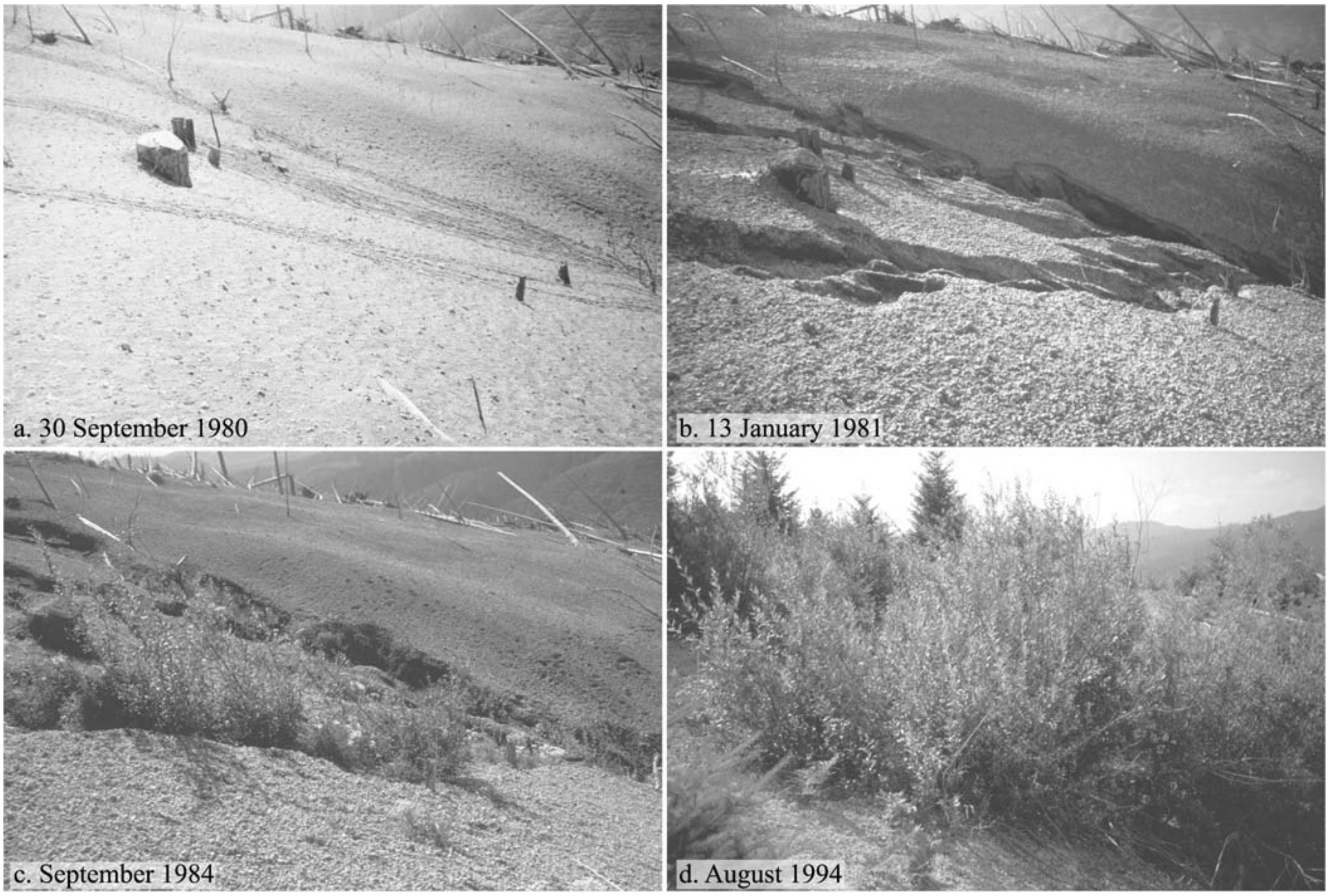

FIGURE 3.7. Chronological sequence of photographs of gully and vegetation development at site east of Smith Creek, 9 km east-northeast of the crater. This preeruption clear-cut facing the volcano received about 50 cm of blast and tephra-fall deposits [Photographer: F.J. Swanson.]

early slides in the blowdown zone occurred despite the absence of major storms (e.g., recurrence intervals exceeding 10 years).

Intense regional storms in November 1995 and February 1996 triggered locally extensive sliding in parts of the blast area. In the basins of the upper Green River, Clearwater Creek, Bean Creek, and Smith Creek, numerous landslides mobilized (1) tephra deposited in 1980, (2) newly established forest, and (3) older tephra deposited by prehistoric eruptions of Mount St. Helens. Each of those basins is underlain by thick accumulations of prehistoric tephra deposits (Crandell and Mullineaux 1978; Mullineaux 1996; see Figure 2.4). Many of the landslides occurred at sites where planted 12- to 14-year-old Douglas-fir forests were well established. The landslide deposits spread onto valley floors and locally entered channels and riparian zones. Floods associated with these storms mobilized some of these landslide deposits and wood in the channels in the eastern part of the blowdown zone, causing severe damage to channel and riparian areas.

Channel impacts and persistence of riparian biological legacies within the blast-affected area varied greatly with distance from the volcano. Within 10 km of the volcano, but outside the upper North Fork Toutle River valley, the blast, mudflows, and secondary blast-pyroclastic flows locally buried valley floors with up to 15 m of hot debris (Hoblitt et al. 1981;Brantley and Waitt 1988), thick enough to obliterate preeruption channel form and to destroy streamside vegetation. Farther from the volcano, the blast deposit was thinner, and secondary blast-pyroclastic flows were less common, so that many manifestations of preeruption channel morphology and riparian vegetation survived. Channels within the blowdown zone unaffected by mudflows or secondary blast-pyroclastic flows (see Figure 3.2) remained relatively stable, in part because tree roots protected their banks. In general, channels within the blowdown zone widened by less than 10 m and incised less than 1 m (Meyer and Martinson 1989). Where the blast toppled mature forest into streams, channel complexity increased greatly, and the trees dissipated streamflow energy and reduced bed and bank scour. Such channel stability aided posteruption survival and establishment of riparian vegetation and development of aquatic habitat. Channel complexity decreased where wood was removed (Lisle 1995) and where debris flows scoured or sediment aggraded channel beds (e.g., lower Bean Creek and upper Clearwater Creek). Development

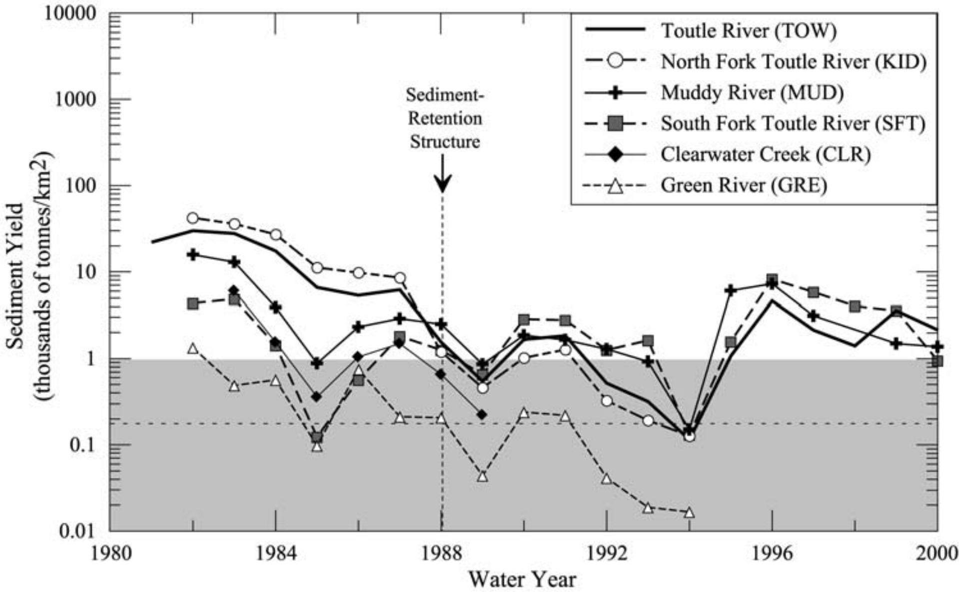

FIGURE 3.8. Annual suspended-sediment yields at gauging stations on rivers draining the Mount St. Helens area. (See Figure 3.2 for disturbance zones and Figure 3.5 for station locations.) *SRS* denotes date when the sediment-retention structure began to trap sediment, affecting sediment yield sampled at KID and TOW. *Shaded region* depicts the range of, and the *horizontal dashed line* depicts median value of, mean annual suspended-sediment yields from several rivers in the western Cascade Range. The yield value depicted by the horizontal line is comparable to average Cascade Range basin yields determined by Beschta (1978), Grant and Wolff (1991), and Ambers (2001). [Adapted from Major et al. (2000).]

of riparian and aquatic habitat along those channels was greatly delayed.

Sediment yield from watersheds where the blast severely disturbed hill slopes but minimally impacted channels was smaller and declined more rapidly than did the yield from basins that experienced severe channel disturbance. In the Green River basin, for example, sediment yield peaked as much as an order of magnitude above background level shortly after the eruption but returned to background level within 5 years (Figure 3.8; Major et al. 2000). This rapid decline was caused, in part, by the generally low streamflow during this period, but it occurred mainly because of the stabilization of rills and gullies that developed on hill slopes and the consequent reduction of sediment delivery to channels.

3.5.1.3 Debris-Avalanche Zone

The debris-avalanche deposit created a distinctive 60-km^2 landscape with complex, actively eroding terrain containing few biological legacies in the forms of soil blocks and vegetation fragments scattered along the deposit margin (see Figures 3.1a, 3.2). The topography of the central part of the deposit was a mosaic of steep-sided hummocks up to 30 m tall and closed depressions (Voight et al. 1981; Glicken 1998). Except for pond and wetland areas, which are common features of the deposit, this nutrient-deficient, erodible terrain presented an inhospitable environment for plant colonization until site conditions were ameliorated by weathering and other processes (Dale et al., Chapter 5, this volume).

Significant establishment of vegetation on the debris-avalanche deposit required development of a channel network across the deposit and a pervasive groundwater table within the deposit. Channel formation began at different times and progressed at varied rates in different areas (Rosenfeld and Beach 1983). On May 18, 1980, the North Fork Toutle River mudflow traveled across the lower half of the deposit and initiated channel formation (Janda et al. 1981, 1984; Meyer and Martinson 1989). Subsequent fill and spill of ponds on the avalanche-deposit surface and breaching of lakes along its margin triggered floods that carved new channels or modified existing ones. These breaches, combined with seasonal runoff and pumping of water from Spirit Lake across the deposit and into the North Fork Toutle River between 1982 and 1985 (Janda et al. 1984; Glicken et al. 1989), produced additional channel change (Rosenfeld and Beach 1983; Janda et al. 1984; Paine et al. 1987; Meyer and Martinson 1989).

Erosion and aggradation of channels across the debris-avalanche deposit created a complex mosaic of surfaces having varying ages, origins, and textures, and they frequently disturbed riparian vegetation. Within a decade of the 1980 eruption, about 35% of the avalanche-deposit surface had been reworked and replaced by braided channels and terraces (Meyer and Martinson 1989). In general, posteruption channels across the debris-avalanche deposit widened by hundreds of meters, incised tens of meters, and were locally aggraded by several meters (Meyer and Martinson 1989).

The groundwater system within the debris-avalanche deposit developed over several years after the eruption and strongly influenced the aquatic and riparian environments. The debris avalanche entrained a large proportion of hydrothermal groundwater that circulated within the volcano on May 18, 1980. That groundwater saturated parts of the debris-avalanche deposit and initially filled many of the depressions that formed on the deposit surface. Precipitation and streamflow contributed gradually to subsequent growth of the groundwater system. Groundwater feeds perennial and ephemeral ponds on the deposit, seep areas, and spring-fed channels. These hydrologic features, having persistently available water and no scouring flows, became hotspots of vegetation establishment and fostered development of plant and animal communities in pondshore, seep, and streamside environments (see Crisafulli et al., Chapters 13 and 20, this volume).

3.5.1.4 Pyroclastic-Flow Zone

The hot, thick, pyroclastic-flow deposits that blanketed 15 km^2 of the upper debris-avalanche deposit eliminated all components of previous ecosystems that may have survived the avalanche and blast and created a harsh environment for colonization. Erosion extensively modified the pyroclastic-flow deposits, carving deep, wide, steep-walled channels into the deposits (Meyer et al. 1986; Meyer and Dodge 1988), and several post-1980 mudflows enlarged the channels and locally coated the deposits with bouldery debris (Cameron and Pringle 1990; Pierson 1999; Pringle and Cameron 1999). In general, erosion dominated modification of the pyroclastic-flow deposits close to the volcano, and deposition predominated in more distant areas, especially where channels approach Spirit Lake. Dramatic channel incision into the pyroclastic-flow deposits followed a substantial increase in discharge of the North Fork Toutle River and an abrupt lowering of the local base level of the stream network in November 1982, when water was pumped out of Spirit Lake and into a channel that traversed the northern edge of the deposits (Janda et al. 1984; Paine et al. 1987; Glicken et al. 1989; Simon 1999; Dale et al., Chapter 19, this volume).

3.5.1.5 Mudflow Zone

Environmental modification by mudflows and effects of biological legacies of the pre-1980 floodplain ecosystem varied longitudinally with distance from the volcano and laterally with distance from a channel axis. On the flanks of the volcano, mudflows scoured substrates or destroyed vegetation and left deposits generally less than 20 cm thick across convex surfaces of alpine meadows and subalpine forest. Many biotic and abiotic remnants of the pre-1980 ecosystem survived in these environments, primarily along flow margins (Frenzen et al., Chapter 6, this volume). Farther from the volcano, channelized mudflows, mostly 3 to 10 m deep, swept river valleys and substantially modified landforms and aquatic and riparian ecosystems. Gross preeruption landforms generally remained intact, but the mudflows locally removed vegetation along river corridors and displaced, straightened, and smoothed river channels (Janda et al. 1981, 1984; Pierson 1985; Scott 1988). The mudflows moved large wood pieces and boulders onto floodplains and transformed rough, cobble- to boulder-bedded, forest-lined channels into smooth, sand-bedded channels having little riparian vegetation. Downstream of the blast area, mudflows completely removed riparian forest adjacent to channels, but commonly placed thin deposits on delicate understory plants and forest litter on floodplains (Frenzen et al., Chapter 6, this volume). Impact forces and abrasion by the wet-concrete-like fluid were greatest near the channel where the mudflows transported the coarsest sediment at the greatest velocities, but they diminished rapidly away from the channel (Scott 1988). Biological legacies persisting along channels included sprouts of standing but abraded hardwood trees and transported plant fragments (Frenzen et al., Chapter 6, this volume).

Posteruption sediment supply and transport affected the geomorphic response and stability of mudflow-affected channels. The greatest and most persistent changes to mudflow-affected channels occurred along the lower Toutle River and North Fork Toutle River below the debris-avalanche deposit (Meyer et al. 1986; Meyer and Dodge 1988). Initially, suspended-sediment yield along the North Fork Toutle River was as much as 500 times greater than yield typical of basins in the western Cascade Range (see Figures 3.8, 3.9; Major et al. 2000). Channel reaches downstream of the debris-avalanche deposit generally stabilized after sediment yield plummeted following closure of a large sediment-retention structure in 1988 (see Figures 3.5, 3.9). Secondary channel changes are evident, but less dramatic, along other mudflow-affected channels (Meyer et al. 1986; Meyer and Dodge 1988; Martinson et al. 1984, 1986) because of lower magnitude of sediment transport. The Muddy River and South Fork Toutle River, for example, had suspended-sediment yields that were initially about 20 to 100 times above typical background level (see Figure 3.8). Those high yields declined rapidly within 5 years but, even after 20 years, remained about 10 times greater than background (Figure 3.8; Major et al. 2000). Despite such high rates of sediment transport, the Muddy River and South Fork Toutle River each discharged only about one-tenth of the posteruption sediment load discharged by the lower Toutle River between 1980 and 1987 and about half the load discharged between 1988 and 2000 after closure of the sediment-retention structure (Major et al. 2000). Mudflow-affected channels, in turn, had suspended-sediment yields about 5 to 10 times greater than those from basins having headwaters located predominantly within the blowdown and scorch zones (e.g., Green River and Clearwater Creek).

Rates of stabilization varied greatly among reaches of mudflow-affected channels; but in general changes in channel geometry declined dramatically within a few years of the 1980 eruption (Meyer et al. 1986; Martinson et al. 1984, 1986; Meyer and Martinson 1989; Simon 1999; Hardison 2000). Channel locations and cross-sectional geometry changed most

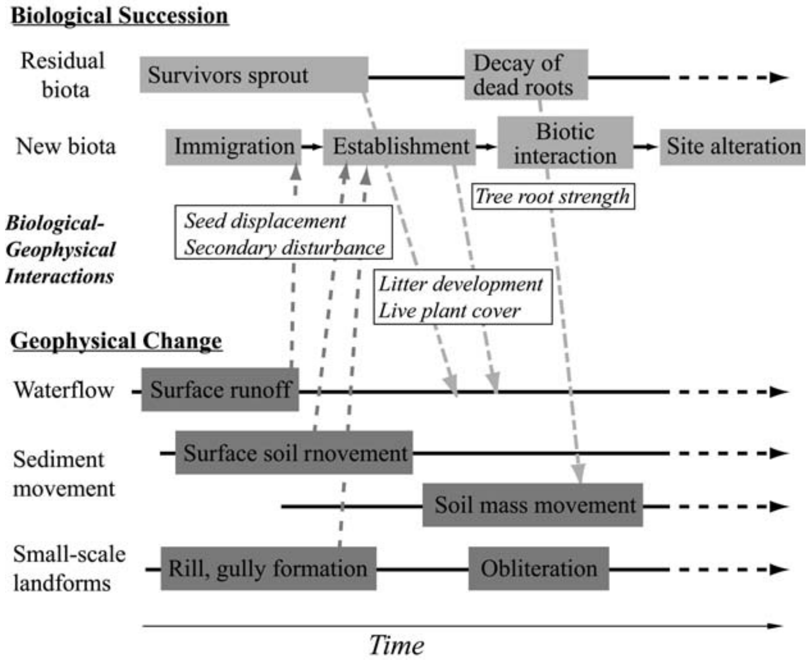

FIGURE 3.9. Schematic depiction of biotic succession, a sequence of geomorphic processes, and interactions of biotic and geomorphic processes at Mount St. Helens following severe landscape disturbance by the May 18, 1980 eruption.

dramatically through 1981 as rivers rapidly incised, widened, and transported the most easily eroded sediment. As channels widened and beds coarsened, rates of change slowed sharply, but some reaches exhibited progressive change through 2000. Mudflow-affected channels generally widened by tens of meters, incised up to 10 m, and were aggraded about 1 m (Meyer and Martinson 1989). Contrasting behaviors of stream reaches are broadly correlated to some extent with overall valley-floor gradient, floodplain width, and the sediment production capability of upstream and streamside areas (Hardison 2000). Although biological legacies persisted, riparian vegetation development along many reaches was delayed until channel geometry stabilized (Frenzen et al., Chapter 6, this volume). Even after channel stabilization, major flooding in 1996 severely disturbed riparian and aquatic systems along several channel reaches (Frenzen et al., Chapter 6, this volume).

3.5.2 Lakes and Lakeshores

The numerous lakes in the Mount St. Helens area after the major 1980 eruption included a mix of those that existed before the eruption (Swanson et al., Chapter 2, this volume) and newly created lakes. Ecological responses to these altered lake environments were similarly diverse (Bisson et al., Chapter 12, this volume; Dahm et al., Chapter 18, this volume).

Lakes existing within about a 30-km radius of the volcano before the eruption (see Figure 2.4; Swanson et al., Chapter 2, this volume) experienced impacts ranging from a light dusting of tephra to profound alteration of basin form and ecology. Of the existing lakes, Spirit Lake was modified most dramatically by the eruption. The debris avalanche raised its outlet, deposited 0.43 km^3 of sediment and woody debris in its basin, and raised its water level by 60 m (Voight et al. 1981; Glicken 1998). A large quantity of wood was also swept from hill slopes into the lake by the seiche generated when the avalanche slammed into the lake. Much of the wood in the lake formed a large, floating mat that has persisted to the present (2004). After 1980, the level of Spirit Lake rose gradually for 2 years until it was controlled by pumping and tunnel construction to form a stable outlet (Janda et al. 1984; Glicken et al. 1989; Meyer 1995). Lakeshore vegetation has recovered slowly because of the generally unstable nature of the lakeshore substrate and the (initially) fluctuating lake level. In contrast, the gross morphology of cirque lakes within the blast area and tephra-fall zone changed little because these lakes are large relative to the amount of sediment that they received. The amount of organic matter delivered to these lakes varied principally with the volcanic process that affected them. Twenty cirque-basin lakes within the blast-affected area received only a few tens of centimeters of blast deposit, including organic matter (Waitt and Dzurisin 1981). In the tephra-fall zone, little organic matter was delivered to lakes by the eruption. Vegetation generally recovered quickly along the shores of these cirque-basin lakes, especially where lake shores had been protected under snow.

Scores of new lakes formed during the May 18, 1980 eruption, and their character and longevity depended greatly upon their mode of formation (see Figure 18.1; Dahm et al., Chapter 18, this volume). Many of these lakes, however, were short lived; six of the nine largest lakes impounded along the margins of the avalanche deposit, for example, filled, overtopped, incised their impoundments, and emptied between 1980 and 1982 (Janda et al. 1984; Simon 1999). The U.S. Army Corps of Engineers constructed stable outlet channels at Spirit, Castle, and Coldwater to prevent catastrophic breaching of their blockages (Janda et al. 1984; Simon 1999).

New ponds formed on the surface of the avalanche deposit when groundwater seepage and surface runoff filled depressions among the hummocks. These ponds initially contained little organic matter because the avalanche deposit was nearly free of organic material. Similar to several lakes formed at the deposit margins, many ponds on the deposit surface were short lived. Within the first few years of the eruption, several of these ponds breached their outlets and emptied. As of 2000, approximately 130 perennial and ephemeral ponds still existed on the avalanche-deposit surface, and about half of them dry out by the end of each summer (Crisafulli et al., Chapter 13, this volume).

3.5.3 Hydrothermal Environments

The broad spectrum of thermal conditions of the new crater and the 1980 deposits created diverse hydrothermal environments that differed in initial temperature, longevity, chemical composition, state (liquid or gas) of emitted water, and discharge (Keith et al. 1981). Locally, some hydrothermal systems created special habitats for posteruption biological activity; such habitats were absent or extremely rare in the preeruption landscape (Phillips 1941). The new hydrothermal environments ranged from hissing fumaroles on the lava dome within the crater to tepid seeps and springs that emerged on the debris-avalanche deposit, pyroclastic-flow deposits, and along the shoreline of Spirit Lake. Lava-dome and crater-floor fumaroles are "rooted" hydrothermal features; that is, they derive their heat from magma. The longevity (perhaps millennial duration) of magmatically supplied heat can potentially maintain long-lived fumaroles. "Rootless" hydrothermal features, for which volcanic deposits themselves provide the source of heat, developed in the pyroclastic-flow, blast, and debris-avalanche deposits. The pyroclastic-flow deposits, particularly in the western half of the Pumice Plain, produced several steam explosions, fumaroles, hot springs, and seeps (Moyer and Swanson 1987). Warm-water seeps occurred locally within secondary blast-pyroclastic-flow deposits and along channels cut into the debris-avalanche deposit. From 1980 to 1984, hydrothermal features formed in some valleys close to the volcano (e.g., Coldwater Creek, South Coldwater Creek, and upper Smith Creek), where secondary blast–pyroclastic-flow deposits accumulated. Unlike most of the other volcanic deposits in which hydrothermal environments developed, these secondary pyroclastic-flow deposits contained abundant organic matter stripped from adjacent hill slopes. Cooking of this organic matter produced gases and liquids having distinctive, short-lived organic chemical characteristics (Baross et al. 1982). Hydrothermal features formed on the debris-avalanche deposit only where channels were cut in the vicinity of isolated pockets of hot rock. Warm-water seeps also occurred along the south shore of Spirit Lake since 1980, but by 2000 these were barely warmer than adjacent lake water. Within only a few years after the 1980 eruption, all known rootless hydrothermal systems had depleted their heat sources.

3.6 Interactions Among Geophysical Processes: A Geomorphic Succession Perspective

Geomorphic responses to profound landscape disturbances follow trajectories somewhat analogous to those followed by ecological responses to volcanic, wildfire (Swanson 1981), and perhaps other disturbance types. Ecologists find it useful to consider biotic responses to major disturbance in terms of succession of biota, biotic processes, and their alteration of site properties by processes such as soil development. Geomorphic processes also interact in ways that resemble interactions among species and biotic processes during the course of ecological succession (see Figure 3.9). In both biotic and geomorphic cases, some processes change sites in ways that favor occurrence of other processes. Furthermore, trajectories of geomorphic and biotic successions interact with one another in important ways.

The primary disturbances caused by the eruption on May 18, 1980, altered landscape hydrology and sediment delivery, which greatly affected the nature and pace of secondary processes that responded to those disturbances. For example, fresh tephra that blanketed hill slopes initially had very low infiltration capacities. Hence, the first substantial rainfalls on tephra-covered slopes produced abundant overland flow that concentrated in hill-slope depressions and carved shallow gullies (see Figure 3.7). Overland flow, freeze-thaw cycles, bioturbation by plants and animals, and deliberate mechanical disturbance of posteruption surfaces (Collins and Dunne 1988) also removed or disrupted the surficial layer of silty tephra and exposed coarser tephra. Exposure of the coarser tephra increased the infiltration capacity of the ground surface and reduced runoff. In many areas affected by the blast, gullies stopped forming by early 1981 and began refilling as their walls collapsed (Collins and Dunne 1986). Increased infiltration through the tephra profile and gully floors also delivered more water to the preeruption soil, which facilitated the occurrence of many small landslides. Thus, successional sequences of processes affected the geomorphic responses of tephra-mantled hill slopes, and some processes had to occur before others could follow.

Another type of sequential interaction of geomorphic processes involves disturbance-triggered movement and storage of sediment on hill slopes and through channel networks. Propagating sediment pulses can cause transient or long-lasting sediment storage in localized depositional sites. Sediment storage in channels causes aggradation, which commonly produces bank erosion and consequent lateral channel migration. Bank erosion in turn feeds sediment to channels and increases stream sediment load, which induces local aggradation. Thus, interacting geomorphic processes can establish feedbacks that cause extensive secondary modification of disturbed landscapes. Interactions of this type were widespread on the debris-avalanche deposit and along mudflow-affected channels (Janda et al. 1984; Meyer et al. 1986; Meyer and Janda 1986; Meyer and Dodge 1988; Martinson et al. 1984, 1986; Hardison 2000). As a result, sequences of geomorphic processes and landforms developed over time and space and greatly affected biotic response.

The persistence of specific geomorphic processes following severe landscape disturbance is varied and depends on local hydrology, properties of soil and biota, and the timing of weather events that trigger episodic processes. Some geomorphic processes diminish within months, whereas others persist for years, decades, or possibly centuries. Such variability reflects the rates at which individual processes operate, the lingering effects of hydrologic perturbations following disturbance, and the timing of sediment transport through disturbed watersheds (Nicholas et al. 1995; Madej and Ozaki 1996; Major 2004). Rates of sheet and rill erosion, for example, were most vigorous in the first few storms after the eruptions but diminished quickly because of changes in the texture and hydrology of tephra-deposit surfaces, even before significant establishment of vegetation. Rates of change of surface infiltration (Leavesley et al. 1989; Major and Yamakoshi in press), the timing of intense storms, and gradual changes in the decay and cohesion of tree roots affected the timing of small-scale landslides from hill slopes in the blast area. Rates of posteruption channel change were most dramatic within the first few years but ebbed as channels widened, streambeds coarsened, and sediment loads diminished. Nevertheless, some reaches of channels that continue to move extraordinary amounts of sediment (Major et al. 2000) remain highly dynamic, especially along the North Fork Toutle River where it traverses the debris-avalanche deposit (Bart 1999).

Secondary geomorphic processes that act as disturbances interact with establishing biota and affect the trajectory of biotic succession (see Figure 3.9). In general, geomorphic systems must achieve a critical level of stability before vegetation can establish sufficiently to flourish and retard further erosion. The example of sheet and rill erosion in the blast area is most conspicuous. Rates of sheet and rill erosion subsided sharply within a year of the major eruption, and thereafter hill-slope erosion stabilized (Collins and Dunne 1986). In some areas, the first substantial vegetation developed along the floors of eroded gullies within a few years of the eruption. In those sites, removal of the sterile blast and tephra-fall deposits fostered natural revegetation processes. Elsewhere, grass and legume seeds applied to control surface erosion in the summer of 1980 washed off steep slopes before they germinated, or they failed to establish or reduce erosion until biological, physical, and chemical conditions of the soil improved and stabilized (Stroh and Oyler 1981; Franklin et al. 1988; Dale et al., Chapters 5 and 19, this volume). Frequent posteruption debris slides occurred in the blowdown and scorch zones before trees and their root systems reestablished and stabilized the new volcanic deposits. Yet, even in well-established, although young, planted Douglas-fir forest, extensive sliding occurred in 1996 when older tephra deposits below the rooting zone of the young forest mobilized. Lateral channel change along some river corridors, even after 20 years, persisted at a pace that suppressed establishment of extensive riparian vegetation (Frenzen et al., Chapter 6, this volume). Biotic stabilization along the banks of some unstable river reaches, especially on the debris-avalanche deposit, appears to be years in the future.

The unifying theme of this collection of observations of the Mount St. Helens landscape is that primary and secondary geomorphic processes change site conditions in ways that enhance or limit the occurrence of other geomorphic processes, much as biotic processes and species change site conditions and affect subsequent biotic processes and communities. Some degree of physical stabilization of a disturbed landscape must precede establishment of vegetation; as vegetation develops, the land surface is further stabilized, and soil development accelerates. In this way, geomorphic and biotic processes interact and affect the path and pace of geomorphic and biotic responses to disturbances.

3.7 Outlook

The combined effects of primary and secondary physical processes triggered by the May 18, 1980, eruption of Mount St. Helens have created an array of environments in which posteruption ecological systems are developing. Initial posteruption conditions ranged from entirely new environments (such as the debris-avalanche deposit, the Pumice Plain, lakes, ponds, and hot springs where conifer forest once stood) to forest, stream, and lake environments only subtly modified by a light dusting of tephra. Dramatic and continuing physical changes to some environments by various forms of hill-slope and channel erosion have regulated the paths and rates of ecological responses to the primary disturbances.

Ecological studies of severely disturbed landscapes can benefit by beginning with a broad stratification of the landscape with a general approach that includes the following:

- Distinguishing between disturbance mechanisms (e.g., erosion/deposition, impact force, and heat) and types of disturbance processes to give greater predictive power concerning ecological responses of both surviving and invading organisms;
- Considering successions of secondary hydrologic and geomorphic processes and geology–ecology interactions in an effort to anticipate how the landscape will change physically and how such change may affect biological legacies, ecological succession, and development of biotic landscape patterns; and
- Designing studies that examine the effects of landscape structure on processes, patterns, and rates of ecological responses.

The May 18, 1980, eruption dramatically altered the landscape, and secondary processes that responded to that disturbance continue to evolve. Future climate change, rare storms and floods, wildfires, and eruptive activity will adjust the course of geomorphic and ecological change. The stage set by the 1980 eruptions and subsequent geomorphic, hydrologic, and ecological responses may influence the geological and ecological responses to the next eruption. Although the importance of biological legacies of the 1980 eruptions to responses to future eruptions will diminish with time, the natural longevity of trees and the slow pace of some ecological processes (such as wood decomposition, soil formation, and development of complex forest structure) may extend those biological legacies for many centuries, a time frame that may encompass the next significant eruption of the volcano.

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