Erosion of tephra from the 1980 eruption of Mount St. Helens

BRIAN D. COLLINS
THOMAS DUNNE
Department of Geological Sciences and *Quaternary Research Center, University of Washington, Seattle, Washington 98195*

ABSTRACT

Erosion of tephra on hillslopes north of Mount St. Helens was measured at intervals of several months between 1980 and 1983 using large arrays of stakes. Rill erosion was initially greater than sheetwash, but its importance decreased comparatively quickly. Both rill erosion and sheetwash decreased by one to two orders of magnitude over the three-year period, because a stable rill network developed and more permeable and less erodible substrates were created or exposed. The decrease occurred prior to recovery of plants. The causes for the rapid decline in erosion at Mount St. Helens explain patterns of erosion following several other recent tephra eruptions.

Basin-wide annual rates of erosion were computed by mapping the surface cover, the hillslope gradient, and the texture and thickness of tephra and by using statistical relations developed between them and erosion rates. Annual erosion decreased from 26 mm between May 1980 and May 1981 to 1.8 mm between May 1982 and May 1983. Projecting this rate of decline into the future indicated that only about one-sixth of the tephra on hillslopes will be removed by water erosion before soil creep and other forms of mass wasting again dominate hillslope evolution.

INTRODUCTION

The 1980 eruption of Mount St. Helens, a Quaternary volcano in the south Cascade Range of Washington, destroyed most above-ground vegetation and deposited a tephra mantle in the upper catchment of the Toutle River, north of Mount St. Helens. Sheetwash and rillwash immediately began to erode the tephra into the Toutle River. We undertook to monitor the forms and rates of erosion of the tephra on steep hillslopes in this humid climate, conditions that are common after eruptions of tephra around the Pacific Rim.

Recent tephra erosion has been studied at several volcanoes, including Parícutin (Segerstrom, 1950, 1960, 1961, 1966); Barcena (Richards, 1965); Irazú (Waldron, 1967); Vulcan (Ollier and Brown, 1971); Fuego (Davies and others, 1978); Sakurajima (Shimokawa and Taniguchi, 1983); and Usu (Kadomura and others, 1983). Most of the results, however, have been qualitative and hard to interpret quantitatively. Several ecological studies have also included qualitative descriptions of the erosion of tephra, most notably at Fernandina (Hendrix, 1981); Parícutin (Eggler, 1963); and Katmai (Griggs, 1918, 1919).

In order to identify individual erosion processes and their controls and spatial patterns, we measured erosion at large arrays of stakes and along transects on hillslopes that represent a range of the major controlling factors. Collins and others (1983) described the spatial patterns of erosion during the first posteruption year. The effects of revegetation and land management are being documented by B. D. Collins and T. Dunne (unpub. data). Here we emphasize changes in erosion over a three-year period, the process interactions that brought about the changes, and similarities between these interactions at Mount St. Helens and at other volcanoes.

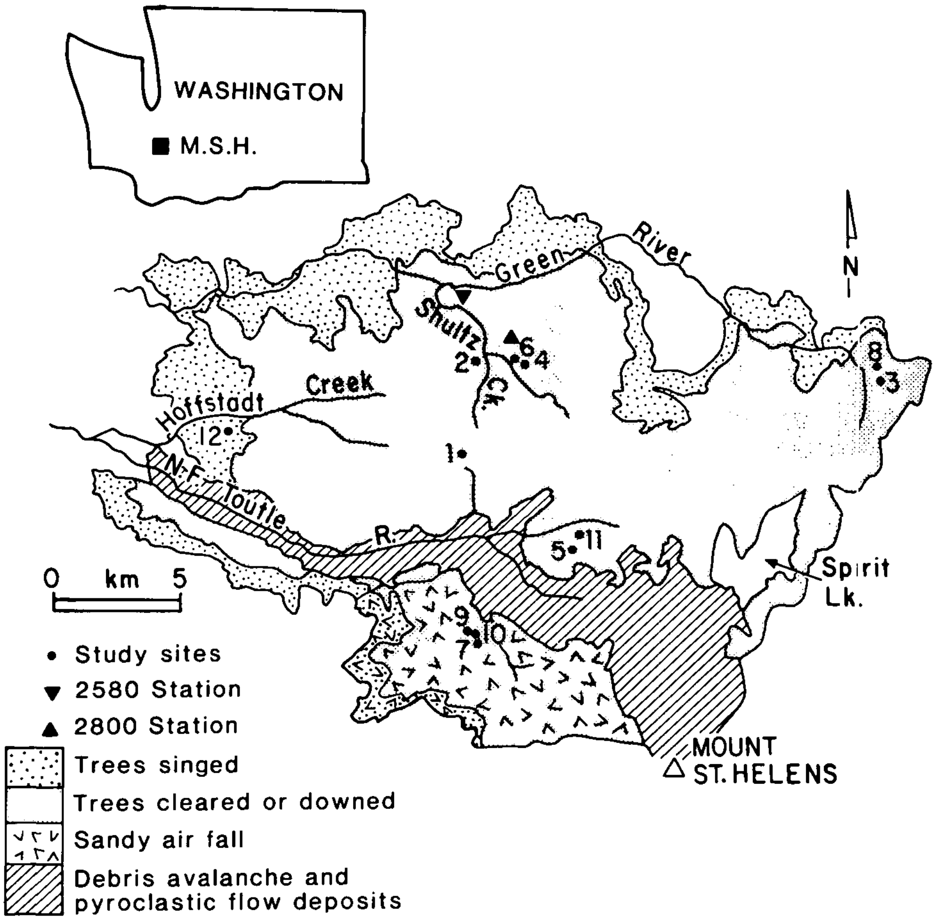

Figure 1. Location of tephra-covered hillslopes studied. Numbered locations refer to Table 1.

Geological Society of America Bulletin, v. 97, p. 896–905, 10 figs., 1 table, July 1986.

ERUPTIONS OF 1980

On the morning of May 18, a magnitude-5 earthquake triggered the detachment of a rockslide from the north flank of Mount St. Helens (Voight and others, 1981). A second rockslide followed immediately, triggering a lateral blast that swept through a 180° arc north of the volcano. The blast cloud traveled as a several-kilometres-thick, ground-hugging, turbulent flow from which the finer particles rose in dense clouds.

Effects on Vegetation

The blast killed the above-ground portions of nearly all plants within a 550-km^2 area. Trees were uprooted and removed from many south-facing hillslopes within 13 km of the volcano. On hillslopes sheltered from the blast and at distances as great as 28 km from its source, trees were uprooted or broken, but not removed (Fig. 1). The blast cloud expanded with increasing distance from the vent until it lifted from the ground (Kieffer, 1981). The distance at which this occurred is marked by the limit of downed trees, beyond which trees were singed and killed but remained standing in a band 0–4 km wide.

These effects of the blast and the pre-eruption patterns of forest management produced a variety of conditions of surface cover (see Collins and others, 1983, their Fig. 2 and photo 1). We mapped three types of cover from aerial photographs: forest cleared by the blast or clear-cut prior to the eruption, downed forest, and singed forest.

On nearby hillslopes facing the vent, forest litter and part or all of the mineral soil were scoured by the blast cloud. Throughout most of the zones of downed and singed trees, however, the litter either was unaffected by the blast or was mixed with the lower several centimetres of the directed blast deposits.

Nearly all merchantable timber on state and private land had been removed by winter 1982. Downed trees were preserved on much of the land within the National Volcanic Monument created in 1982. The effects on erosion of tree removal and other types of land management following the eruption are discussed elsewhere (B. D. Collins and T. Dunne, unpub. data).

Deposits

Deposits of the blast were investigated by us and by three U.S. Geological Survey groups. The stratigraphy and sedimentology and interpretations of the mechanics of deposition are described in detail by Hoblitt and others (1981), Moore and Sisson (1981), and Waitt (1981). In this report, the deposits of the blast are classified into "pyroclastic-surge deposits" and overlying air-fall deposits. We did not study the "blast-pyroclastic flow deposit" of Hoblitt and others (1981), which fills some valleys to anomalous thicknesses (Lipman and Mullineaux, 1981, Pl. 1) because it is being eroded dominantly by stream channels (Lehre and others, 1983) rather than by hillslope processes.

The median grain size of the pyroclastic-surge deposits decreased with distance from the vent from ~1 mm at 10 km to 0.2 mm at 25 km (Fig. 2a). The deposit thickness decreased with distance from the vent, from 1.0 m at 10 km to 0.02 m at 25 km, and varied with the topography, thinning to zero on the steepest slopes at all distances from the volcano. This effect was greatest near the volcano, where the tephra layer was thick on moderately to gently sloping ridges and hillsides and thin or absent on the glacially eroded, steep lower slopes. The samples shown in Figure 2a do not represent these very steep slopes.

Overlying the pyroclastic-surge deposits, there is an air-fall layer resulting from the ash-laden clouds that rose from the turbulent pyroclastic surge. The grain size of this air-fall layer is relatively constant throughout the area affected by the blast (Fig. 2b); its thickness immediately after the eruption ranged from ~1 to 10 cm and averaged 3–4 cm.

Air-fall tephra from numerous pyroclastic flows and from vertical, magmatic eruptions later on May 18 and subsequent dates fell atop the deposits of the blast. Only the eruption on May 25, however, deposited an appreciable thickness of tephra over much of the Toutle River catchment (Sarna-Wojcicki and others, 1981), covering the southwestern part of the blast-affected area (Fig. 1) with a sandy layer 2 to 10 cm thick (Fig. 2b).

The volume and weight of tephra were estimated by mapping the thickness, grain size, and bulk density of each tephra layer (Collins and others, 1983; Collins, 1984), excluding thick deposits in valley bottoms, at the margins of and within the debris-avalanche deposit in the valley of the North Toutle River, on the flanks of Mount St. Helens itself, and in catchments east of the Toutle River basin. The total was estimated as 0.07 km^3, or 9×10^7 t, consisting of 22% silt and clay, 63% sand, and 15% gravel. The total volume of blast deposits was estimated by Moore and Sisson (1981) to be 0.19 km^3.

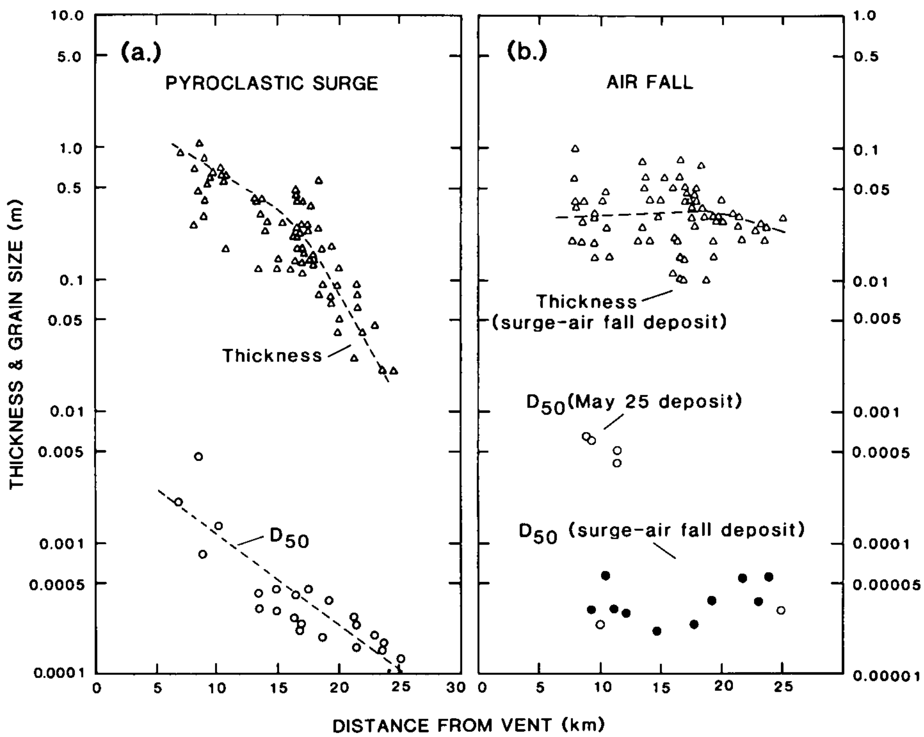

Figure 2. (a) Variation of median grain size (D_{50}) and thickness of the pyroclastic-surge deposit (excluding the air-fall unit) with distance from the volcano. (b) Grain size and thickness of the air-fall layer associated with the pyroclastic surge and grain size of the May 25 air-fall layer. Thickness of the May 25 layer (not shown) plots within the field defined by the surge-related air-fall layer (data indicated by filled symbols are from Hoblitt and others, 1981).

THE UPPER TOUTLE RIVER CATCHMENT

Dissection of Tertiary volcanics has given rise in the western part of the basin to generally straight to convexo-concave hillslopes with smooth longitudinal profiles, average gradients of 0.30, and lengths of 560 m. Elevations in the basin range from 350 to 1,786 m. The eastern portion of the basin is higher and more rugged, displaying the effects of alpine glaciation. Hillslopes there have more irregular profiles, with an average gradient of 0.45. Several metres of Holocene tephra that augment colluvium in the east pinch out to the west, where the colluvial cover averages 1–1.5 m in thickness.

Before the blast, forests below an elevation of ~900 m were dominated by Douglas fir (*Pseudotsuga menziesii*), Western hemlock (*Tsuga heterophylla*), and Western red cedar (*Thuja plicata*) and at higher elevation by true firs (*Abies* sp.) (Franklin and Dyrness, 1973). Intensive logging during the preceding 40 yr created a patchwork of old-growth forest, tree farms, and recently clear-cut forests.

TABLE 1. DESCRIPTION OF HILLSLOPES STUDIED

Site	Cover type*	Gradient	Elevation (m)	Horizontal length (m)	Average distance from vent (km)
(1) Maratta	c	0.16	930– 960	180	14.5
(2) Shultz 1	c	0.25	710– 890	625	17.6
(3) Green River 2	c	0.25	910–1,010	650	17.7
(4) Shultz 3	c	0.32	1,010–1,200	345	16.3
(5) Coldwater 1	c	0.36	1,100–1,200	345	8.6
(6) Shultz 2	c	0.56	955–1,025	245	16.6
(7) Castle 1	c	0.66	932–1,242	645	9.0
(8) Green River 1	d	0.25	915–1,065	725	17.8
(9) Castle 3	d	0.39†	1,100–1,200	552	9.3
(10) Castle 2	d	0.45	1,045–1,195	300	9.2
(11) Coldwater 2	d	0.54	945–1,095	240	8.9
(12) Hoffstadt 1	s	0.45	660– 690	225	21.6

Note: number in first column refers to locations shown in Figure 1.
*c, cleared forest; d, downed forest; s, standing singed forest.
†Average of two facing hillslopes.

Climate

Annual precipitation between 1930 and 1957 followed a strong west-northwest to east-southeast orographic gradient and ranged from 1,650 to 3,050 mm/yr (U.S. Department of Agriculture Soil Conservation Service and U.S. Weather Bureau, 1965). Approximately 75% of annual precipitation occurred during a 6-mo period beginning in October and reaching a monthly maximum in December. Storms were generally long and of low intensity; for example, the estimated average intensity of the 2-yr, 1-hr and 2-yr, 24-hr storms was 14 and 4 mm/hr in the Shultz Creek catchment (Miller and others, 1973).

The Weyerhaeuser Company recorded precipitation at 1-hr intervals during the study at 2 gages in the Shultz Creek catchment (Fig. 1), placing the 2580 Road gage at an elevation of 525 m in July 1980 and the 2800 Road gage at an elevation of 900 m in May 1981. Annual precipitation at the 2580 station averaged 1,400 mm between October 1980 and October 1982. The record of precipitation at the 2800 station was incomplete but averaged 8% more than at the 2580 station.

The maximum 1-hr precipitation intensities at the 2580 station between October 1980 and September 1981 and between October 1981 and September 1982 were only 9.9 and 9.4 mm/hr, respectively (Fig. 3). Intensities at the 2800 gage averaged ~10% more than those at the 2580 gage.

The minimum daily air temperature at the 2800 gage dropped below 0 °C on 54 days in the 1981–1982 year, as did the mean daily tephra temperature at a 5-cm depth for 6 days. Soil at 30-cm depth did not freeze. Although these data indicate the tephra layer rarely froze as deep as 5 cm, the diurnal range in air temperature and our own field observations indicate that the upper several centimetres of tephra underwent frequent freezing and thawing during the winter. The surface of the tephra at elevations above 1,100 m remained frozen for much of the winter.

Elevations between ~700–1,000 m were snow-covered intermittently during several weeks in each of the first two posteruption winters, and higher elevations were snow-covered for several months. During the third winter, snow cover was greater and persisted longer in the spring.

Hillslope Hydrology

Before the eruption of May 18, infiltration capacities of the heavily forested, gravelly and sandy loams greatly exceeded rainfall intensities and snowmelt rates. Storm runoff was generated mainly by subsurface stormflow and by saturation overland flow from small areas of the landscape. In logged areas, overland flow also occurred on road surfaces and margins. The major erosion processes were various forms of mass wasting, and the average sediment yield of the Toutle River basin was probably 100–200 t/km^2/yr (Reid and others, 1981; Swanson and others, 1982; Larson and Sidle, 1980).

The eruption radically altered the hillslope hydrology by removing the forest vegetation from large areas and by depositing a blanket of tephra over the permeable topsoil (Fig. 1). In August 1980, Herkelrath and Leavesley (1981) measured infiltration capacities of 1–5 mm/hr on silty air-fall tephra in the basin of Shultz Creek. Rainfall intensities measured during the first posteruption year (Fig. 3) indicate that these infiltration capacities were exceeded frequently. We often observed overland flow on the tephra surface between September and November 1980 while installing erosion stakes.

Under singed trees, falling needles were mixed with tephra as it fell and during later rains, creating a protective mat that inhibited runoff and erosion. Beyond the zone of singed trees, the canopy of standing trees temporarily intercepted the tephra, which fell later as dry dust, covering the forest floor only as a thin, discontinuous, and permeable coating. The dusty coating of tephra did not lower infiltration capacities or generate overland flow and surface erosion.

STUDY SITES AND METHODS

Hillslopes selected for study represent each of the three types of cover and within each type a range of gradient and length (Table 1). The South Castle Creek sites are within the area covered by the sandy May 25 tephra layer.

In September and October 1980, steel stakes 75 cm long and 1.3 cm in diameter were installed vertically on each hillslope, spaced at intervals of 2 or 5 m along contour transects 80 or 200 m long. The transects were spaced at intervals of 10%, 20%, 40%, 60%, and 80% of the distance from ridge crest to valley bottom. In addition, on 7 of the sites, 2 transects of 50 stakes each were set out at equal spacings down the general gradient of the hillslope. Each stake was driven into the underlying soil so that ~15 cm of the stake was exposed above the tephra surface. A ruler was used to measure the distance from the top of each stake to the surface at the time of installation and at several times during the following three years. The locality of each stake was categorized at the time of measurement as "rilled" or "unrilled," and the change in surface elevation between succeeding measurements was attributed to either rill ero-

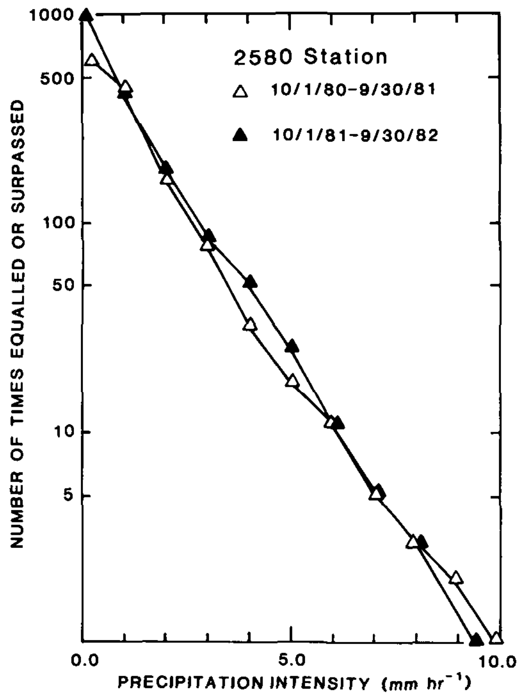

Figure 3. Intensity and frequency of precipitation in 1-hr intervals at the 2580 station (from data provided by Weyerhaeuser Company).

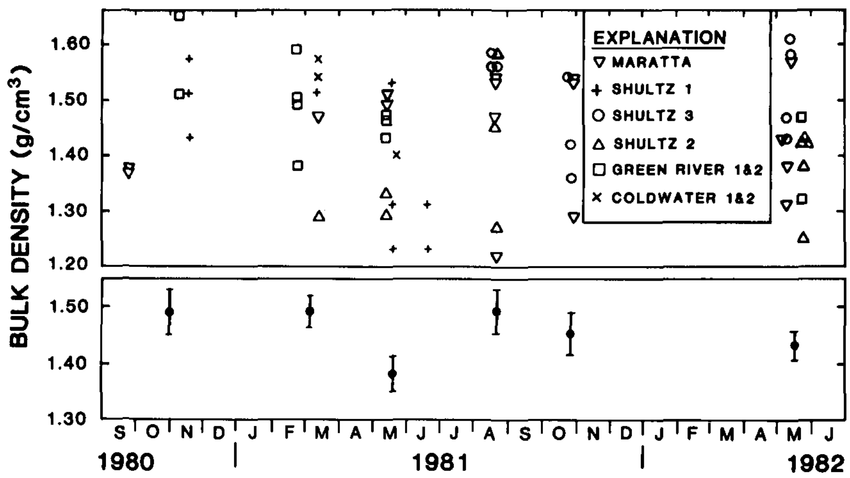

Figure 4. Bulk density of the upper 108 mm of undisturbed tephra at individual sites and mean and standard error of data from all sites. Samples were collected in a 73-mm-diameter cylinder with a 1-mm-wide lip.

sion or interrill erosion as appropriate. The rill cross sections originally were rectangular.

The top width and the average depth of each rill crossed by the transects were also measured, along with the depth to the buried soil in those rills that had exposed the soil. All field measurements were made during eight periods of several weeks in September-November and November-December 1980; February-March, May-June, August, and October 1981; May-June 1982; and May 1983.

Erosion along each transect was taken as the average change of elevation at all stakes. The average lowering by rill and interrill erosion on a transect was calculated from stakes at rilled and unrilled locations. The respective rates of erosion were calculated from those lowering rates and the relative proportions of stakes. Rill and sheetwash erosion were also calculated on the down-gradient transects, but on several hillslopes, certain down-gradient transects overestimated rill erosion, because they tended to follow swales in which larger rills were concentrated. For this reason, we computed the mean change in surface elevation as the spatially weighted average of the contour transects only.

The accuracy of the rill-erosion measurements was checked by subtracting the cross sectional area of all rills along a contour transect from the area of rills at the succeeding measurement and adding a correction for lowering between the rills. When computed in this way, the average rill erosion for all sites between September-October 1980 and May 1981 was within 5% of the rill-erosion rate indicated by stake measurements. In the two subsequent years, rill channels were no longer well approximated by a rectangular cross section, for reasons discussed below, and comparison of the two measures of erosion was no longer valid.

The first stake measurements were made four months after the eruption and after numerous cycles of wetting and drying. Waitt (1981) observed that the air-fall layer compacted during the first several days after the eruption but not in the following months, whereas the sandy pyroclastic-surge layers did not compact at all. To document possible changes in density of tephra, we sampled the bulk density each time we measured the stakes, taking random samples from relatively uneroded tephra profiles (Fig. 4). Densities at the Castle Creek sites were not routinely sampled. The data show no discernible, systematic change in density, nor do separate samples of the air-fall layer alone (Collins, 1984).

The elevations of stakes were not measured by spirit leveling, but the stakes almost surely have not changed in elevation, except for a small number that were undermined by rills. Soil temperatures indicate that the buried soil in which the stakes were anchored has remained unfrozen, and so the stakes have not been subject to frost heave.

Distances from stake top to ground surface were measured repeatedly on two transects, including one on a steep, sandy surface under a thick cover of blown-down trees. The average measurements differed by <0.5 mm.

CHANGES IN PROCESSES AND RATES

Figure 5a shows the cumulative mean change of surface elevation on the study sites. The changes represent the combined erosion of tephra and underlying colluvium. The temporal patterns of interrill and rill erosion are discussed separately below.

Interrill Erosion

The cumulative mean lowering of interrill surfaces is shown in Figure 5b. The figure displays the same over-all changes in rate of lowering shown by the rills (Fig. 5c, discussed below), although the decline is less rapid. This decreasing rate of change is due in part to an increase in the infiltration rate of the interrill surface. Infiltration experiments conducted in 1980 by Herkelrath and Leavesley (1981) were repeated in 1981 on 2 plots within the Shultz 1 site and showed that the infiltration rate had increased from 1–5 mm/hr to 7–9 mm/hr (G. H. Leavesley, personal commun.). The experiments were done on large plots that averaged the infiltration rates in rills and divides. Our measurements along one stake transect on the steeper plot

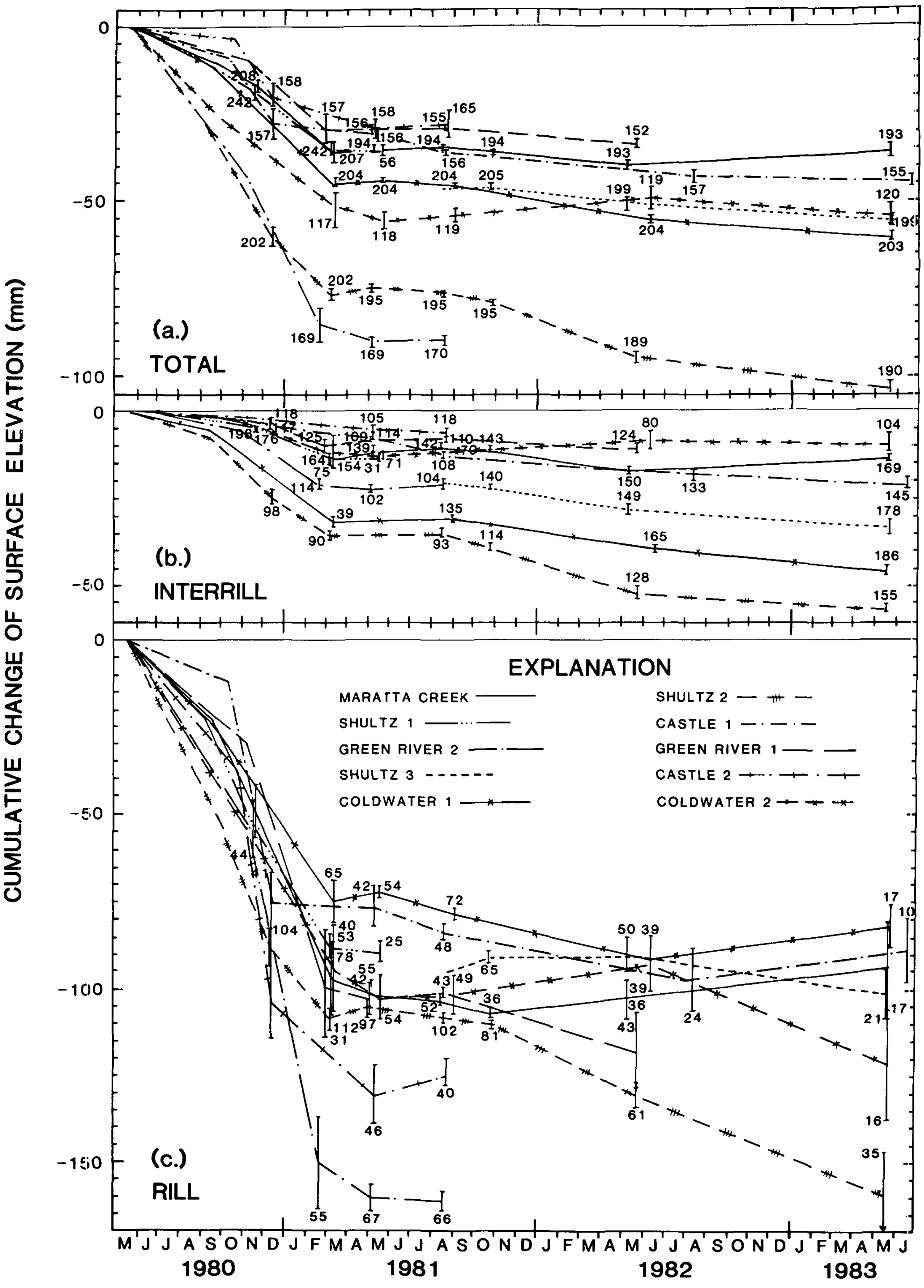
CUMULATIVE CHANGE OF SURFACE ELEVATION (mm)
(a.)
TOTAL
(b.)
INTERRILL
(c.)
RILL
0
-50
-100
-150
EXPLANATION
MARATTA CREEK
SHULTZ 1
GREEN RIVER 2
SHULTZ 3
COLDWATER 1
SHULTZ 2
CASTLE 1
GREEN RIVER 1
CASTLE 2
COLDWATER 2
M J J A S O N D J F M A M J J A S O N D J F M A M J J A S O N D J F M A M J
1980
1981
1982
1983

Figure 5. (a) Cumulative mean change in surface elevation of study sites. Error bars show one standard error of the mean. Numerals are the number of stakes in the sample, which changes on some sites because of loss or disturbance of stakes. Lowering prior to stake installation was taken as the sum of rill cross-sectional areas plus an estimate of sheetwash erosion from the regression analysis of data by Collins (1984). The standard errors range between 0.5 mm and 5.9 mm and average 2.0 mm. The Castle 3 site (not shown) exhibited mean lowering of 19 ± 6.3 mm (n = 121) between 25 October 1980 and 25 August 1981. (b) Cumulative mean change in interrill surface elevation with one standard error of the mean and sample size at each site. Lowering on the Castle 3 site between 25 October 1980 and 25 August 1981 was 0.2 ± 2.5 mm (n = 86). (c) Cumulative mean change of surface elevation in rills, showing one standard error of the mean and sample size. Lowering on the Castle 3 site was 66 ± 19 mm (n = 35) between 25 October 1980 and 25 August 1981.

showed that there was no change in the area of rills that incised through the silty air-fall layer between the two sets of experiments. The stability of the rills suggests that part of the increased infiltration was due to increased interrill permeability. This inference is confirmed by the infiltration experiments conducted by Fiksdal (1981) on 1-m^2 plots of unrilled tephra in March 1981. His results showed a range from 2 mm/hr on plots that had been relatively uneroded to 11 mm/hr on plots on which the silty layer had been more eroded. One reason for the increased interrill permeability in these experiments was exposure of the underlying pyroclastic-surge deposits by stripping of the silty air-fall layer. Field inspection and Figure 5b indicate, however, that the complete stripping of the 40- to 60-mm-thick air-fall layer occurred only locally. Although this local stripping of the air-fall layer was an important factor in increasing the interrill infiltration rate, changes in the permeability of the air-fall layer also may have been important. Where the air-fall layer had been eroded only slightly, visual observation of runoff during storms indicated that there was less runoff during the winter of 1981–1982 and hence an implied increase in infiltration, despite a similar or slightly more intense rainfall regime. The inferred increase in the infiltration rate of the interrill surface could have been brought about by growth of needle ice, which we commonly observed during the winter of 1980–1981, and which would have replaced the original structure of unconnected vesicles with a more open and connected network of pores.

The lowering rate also decreased because the interrill surface became less erodible due to the concentration of sand and granules by erosional sorting.

Rill Erosion

In September 1980, rill densities on the silty May 18 tephra ranged from 0.6 to 5.0/m of contour, whereas densities on the sandy May 25 tephra ranged between 0.2 and 1.7/m. Such a difference is to be expected from differences in infiltration capacity. Dunne (1980) argued that conditions favoring a greater frequency and amount of sheetwash relative to rainsplash would promote rill formation at shorter distances from local divides. The relative intensity of sheetwash and rainsplash on silty and sandy tephra is further indicated by the near-vertical sides of rills in silt, which contrast with the convex, gently sloping sides of rills in sand. Mosley (1973) demonstrated the efficacy of rainsplash in developing convex divides. It is also possible that the sand had a greater critical tractive force than that of the silt, but we do not know this, because the silty tephra was cohesive.

The low-order rill network began to form on the silty and sandy tephra within several centimetres of local drainage divides and at all distances downslope. The rill networks reflected microtopography created by tree roots and fallen wood. The higher-order rills were oriented along the general gradient of the hillslopes, with tributary junction angles being approximately inverse to the gradient. Many rills were anastomosing or discontinuous, either because water overflowed the channel banks, or because low gradients caused deposition of alluvial fans and spreading of flow.

Figure 5c shows that rills incised rapidly during the first ten months after the eruption (until February 1981) and at a sharply decreasing rate thereafter. Many rills stopped incising after reaching the buried soil, because the highly permeable, dense root mat on the soil surface absorbed water and reinforced the soil surface. Some of the larger rills penetrated this mat and incised the underlying soil (for example, Fig. 6). The depths of the largest rills were eventually limited on the steepest slopes by bedrock at depths of 1–1.5 m, except on thick, Holocene tephra layers where rills as deep as 2.5 m reached weathered, cohesive horizons.

As the beds of the rills were being lowered, important changes took place in the channel network. Figure 6 shows the frequency distribution of cross-sectional areas of rills along five transects on a typical hillslope. At the beginning of measurements, many small rills crossed each transect. In September 1980, the average density of rills with cross-sectional areas less than 1 cm^2 was 1.6–4.0/m, and this density declined rapidly to zero over the following year. Three years after the eruption, all rills with cross-sectional areas less than 31.6 cm^2 had been eliminated. Apart from these smallest rills, however, the modal area of rills in tephra remained between 18 and 100 cm^2 throughout the measurement period at all distances downslope, except at the time of the last measurement, when only some of the largest rills remained.

The rills incised into colluvium were generally larger than those confined to the tephra, and their size increased gradually and irregularly downslope and through time. During the 3 yr after the eruption, the modal cross-sectional area increased from a range of 56–562 cm^2 (increasing downslope) to a range of 562–1,778 cm^2. By the end of this period, all rills incised into colluvium with cross-sectional areas of <56–100 cm^2 were obliterated. The density of all rills decreased monotonically throughout the observation period, whereas their total cross-sectional area increased during the first winter and decreased thereafter (see Fig. 7 for the results from a typical hillslope). We interpret these changes in the rill networks in the following manner.

High drainage densities developed soon after the eruption during low-intensity rains with small drops, which are typical of the Pacific Northwest. Such rainstorms on steep surfaces that had exceedingly low infiltration capacities would generate high shear stress for sheetwash but relatively low amounts of rainsplash. Dunne (1980) proposed, and Dunne and Aubry (1986) confirmed experimentally, that rills should form under such conditions in a manner close to that described by Horton (1945) for sheetwash acting alone. In the early stages of network development, larger storms could cause both an increase in the sizes of rills (Fig. 7) and overtopping of drainage divides between rills, leading eventually to the cross-grading and micropiracy proposed by Horton (1945, p. 333) as a means of reducing drainage density (Fig. 7).

Frost action and erosion of silty tephra meanwhile would have increased the infiltration capacity, as documented previously. Such a change would have reduced the frequency and amount of sheetwash without altering the raindrop energy available for rainsplash. It is even possible that disruption of the surface made the tephra more susceptible to splash transport, at least in areas where the silty air-fall layer had been stripped. By these changes, the relative effi-

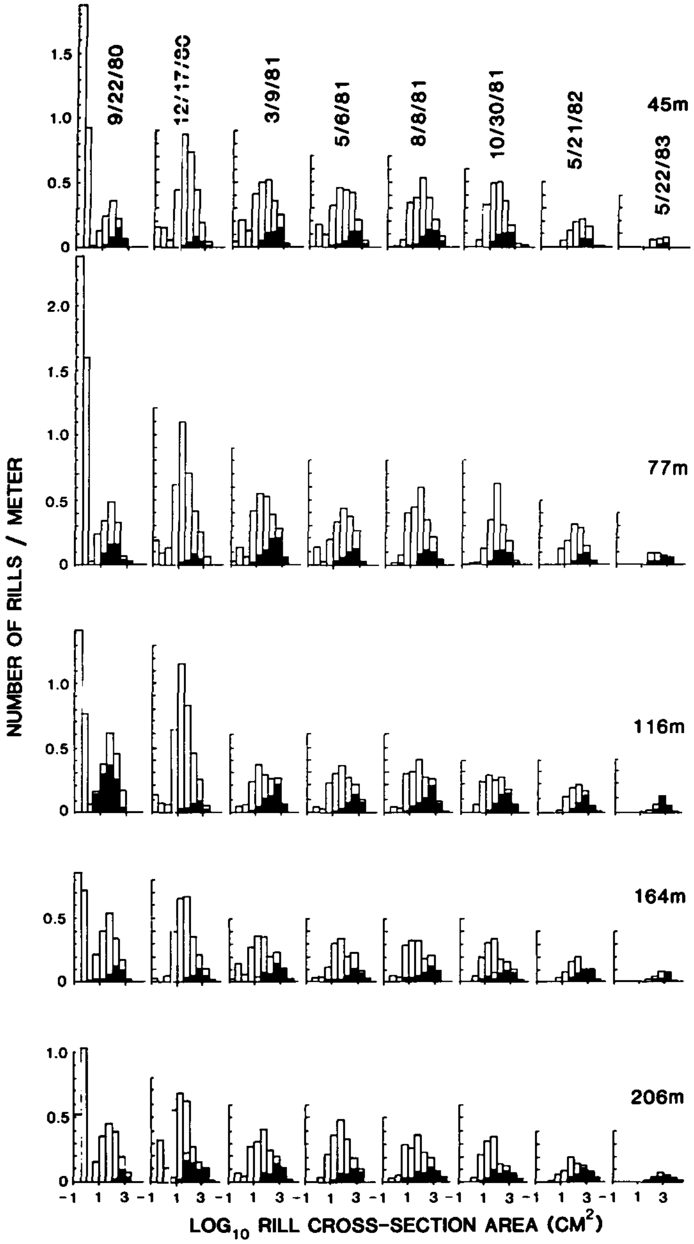

Figure 6. Frequency distribution of rill cross-sectional areas (cm^2) for the Shultz 2 site. Frequency distributions in half-log$_{10}$ increments are shown for each of five transects measured on eight occasions. Distance from the divide is indicated on the right-hand side for each transect. The rills are divided into those incised into tephra only (unshaded) and rills incised into the buried soil (shaded). Histograms represent 15,336 sets of width and depth measurements.

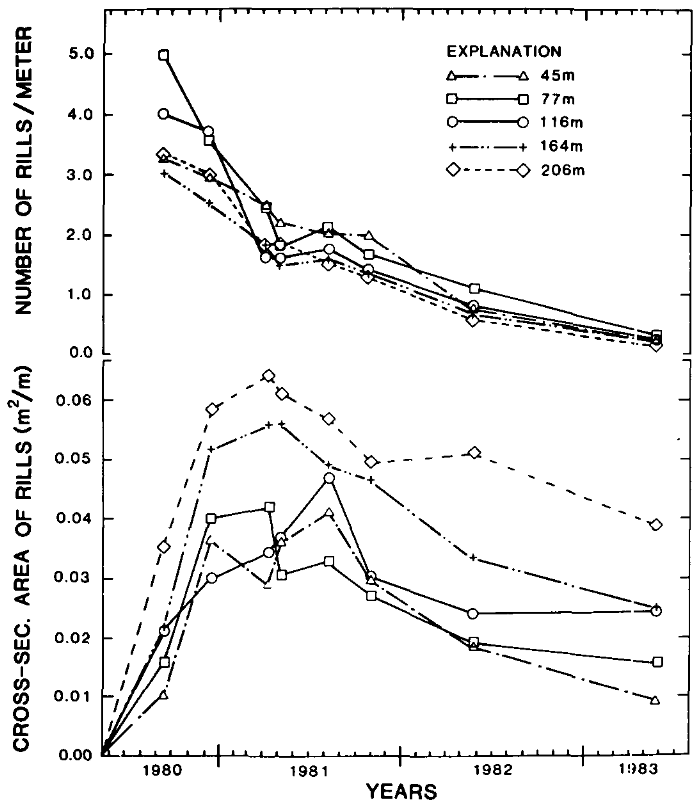

Figure 7. Variation with distance downslope and time of the number of rills per metre of transect and the sum of cross-sectional areas of rills on the Shultz 2 site.

cacy of sheetwash and rainsplash would be reversed, and sediment would be splashed from the divides into rills, where runoff would be either absent or too feeble to transport all the influx of sediment. The smaller rills thus would be filled and the surface irregularities smoothed (Dunne, 1980). This process of diffusing sediment away from protuberances and filling depressions in the surface would have been augmented by the effects of freeze and thaw.

The filling and complete obliteration of small rills are reflected in Figure 5c by the decline with time in the number of stakes at rilled localities; the partial filling of larger rills is represented by the net increase in surface elevation shown at some rilled sites. The process of filling would also be favored in rills that had incised permeable layers, because the rills would lose water through their beds. A simple calculation, however, shows that important amounts of water would be absorbed in this way only in rills that had incised to the forest floor or the gravelly layer of the tephra.

The result of these interactions between rillwash, sheetwash, and rainsplash would be that the drainage density of rills would decrease to a stable value after an initial increase and that short, intervening divides would assume a convex form, both rills and divides being in balance with the rate at which sediment is splashed from the divides into the rills. The balance would be modulated by slow transport of small, pumiceous gravel that would accumulate on the surface of some slopes and in rill beds (Fig. 8). The geomorphic evolution of a more stable form thus would be responsible for the radical decrease in erosion and drainage density before revegetation began (Fig. 9). We discuss elsewhere (B. D. Collins and T. Dunne, unpub.

data) revegetation and its significance for land management.

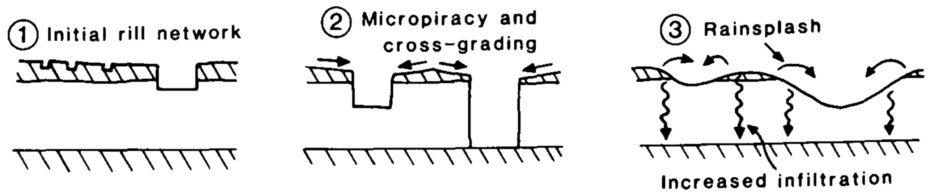

Figure 8. Schematic evolution of a stable rill network. 1. Early storms with a low intensity and small drop size falling onto an impermeable surface gave rise to a dense network of small, subparallel rills with a rectangular cross section. 2. Later, larger storms brought about a decreased drainage density due to micropiracy and cross-grading. Incision of larger rills was limited by the less erodible and more permeable buried soil. At the same time, infiltration capacity on divides increased due to the stripping and alteration of the silty air-fall layer. 3. As a result of these changes, rainsplash became dominant over rillwash and sheetwash, filling or obliterating rills and giving rise to a smoothed surface with convex divides between rills.

ESTIMATION OF BASIN-WIDE EROSION

Erosion in Cleared Forests

An analysis of the spatial patterns of erosion was made by Collins and others (1983) and is only summarized here to emphasize the factors governing basin-wide erosion. Erosion averaged over the scale of a hillslope was found to be affected by over-all hillslope gradient, tephra thickness, tephra texture, and the surface cover of vegetation or organic debris. Of these, surface cover had the greatest effect, and so the sites were first grouped according to three cover types: clear-cut or blast-cleared forest, downed forest, and singed forest. Average erosion rate was correlated with gradient, thickness, and texture for each cover type. A linear regression model was used, because it fit the data adequately. As the erosion rate changed with time, measurements for each posteruption year were analyzed separately.

During the first posteruption year, rills accounted for >70% of the total erosion in clear-cut or blast-cleared forests, where the tephra surface was smooth and bare immediately after the eruption. The relative importance of rill erosion and the degree to which it excavated colluvium increased with hillslope gradient. The less erodible colluvium reduced the rate of incision of the largest rills over most of the study area. Erosion of sandy air-fall tephra was slower than erosion of silty tephra with the same thickness and gradient. On the single site where both rill erosion and sheet erosion were measured (Castle 1), rills removed a larger proportion of the total sediment eroded than was the case on any of the sites on silty tephra. These rills, however, did not incise through the 15- to 30-cm-thick sandy layer to colluvium. During the second and third posteruption years, both rill and sheetwash erosion increased with the hillslope gradient, although more slowly than during the first posteruption year, because a stable geomorphic surface formed more rapidly on steeper slopes on which initial erosion was more intense.

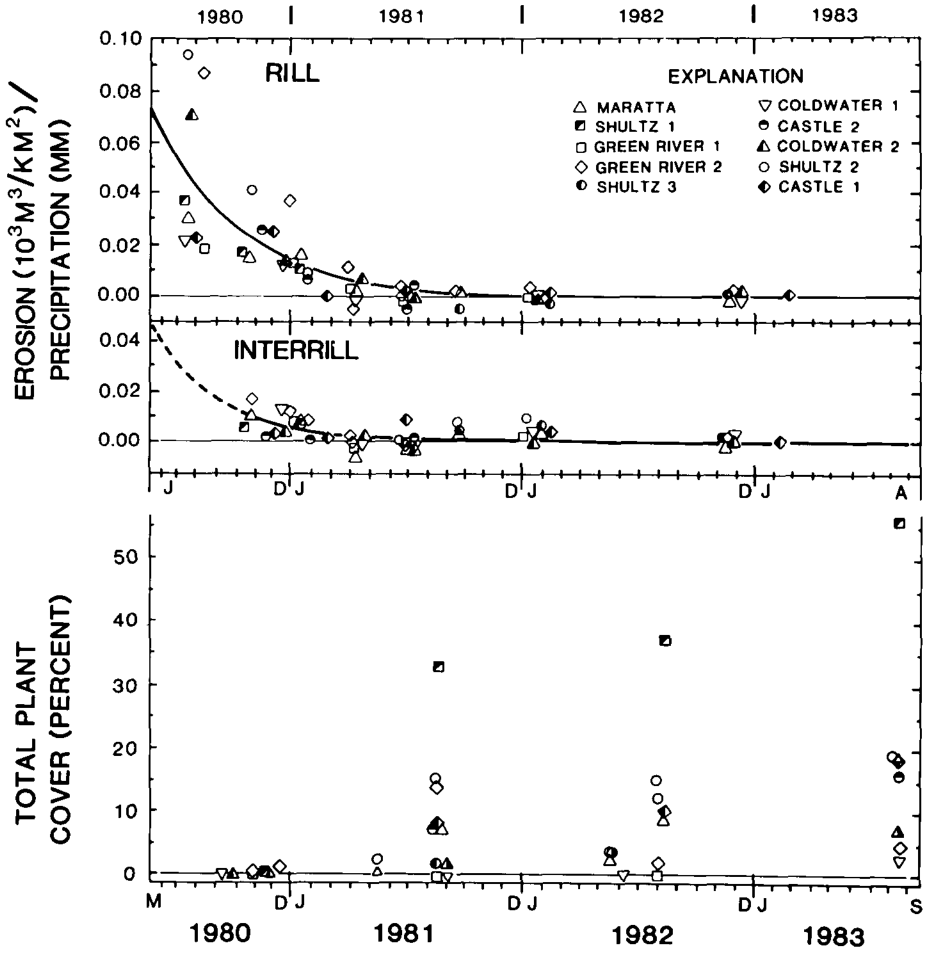

Figure 9. Erosion per unit of precipitation at all sites versus time since the eruption, for rills ($y = 0.073e^{-0.0072x}$, $R^2 = 0.57$) and interrill surfaces ($y = 0.046e^{-0.0091x}$, $R^2 = 0.32$), where x is days after the May 18 eruption and y is erosion in millimetres normalized by millimetres of precipitation. The timing of the decline in erosion is compared to that of changes in total cover provided by native and seeded plants.

Erosion in Downed and Singed Forests

Sheetwash on four hillslopes in blown-down forests ranged from 66% to 92% of that on cleared hillslopes with the same gradient (Collins and others, 1983). Rills were measured in

July 1981 on an additional six hillslopes under the fallen tree cover. On the 10 hillslopes, rill erosion under the organic debris ranged from 10% to 30% of its value on cleared slopes with similar gradients and silty tephra, and from 20% to 50% of its value on cleared, sandy tephra.

In downed forests, three factors combined to reduce erosion. Trees were toppled during the initial phase of the blast before the tephra was deposited, and so many trees were partially or completely embedded in the tephra. The mixture of trees and tephra would have reduced the local gradients and lengths of slopes above the downed trees, relative to slopes in the cleared forest. The measured relative areas of local slopes, however, were the same in the downed and cleared forests, because trees were in contact with tephra on only ~10% of the surface, and so the influence of embedded trees would not have been great. Trees downed by the eruption also intercepted rain, but the effects of reduced raindrop energy was small because of the height of most fallen trunks and the small area they covered (Collins, 1984). Most important were the effects of the bark, branches, splintered trunks, and pulled-up tree roots embedded in and mixed with the tephra. This organic debris provided cover against raindrop impact and also created avenues through which both unchanneled and channeled overland flow could infiltrate through the tephra to the underlying soil, thereby decreasing the total amount of surface runoff.

On hillslopes with singed trees, whether standing or fallen, less erosion occurred than elsewhere. Tephra layers were generally thin in the zone of singed trees, and so the topography of the forest floor commonly disrupted the surface, creating some of the same effect as did the downed forests. The dominant factor reducing erosion, however, was probably a protective mat of conifer needles several centimetres thick that fell after the eruption. Measurements of erosion stakes in a singed forest in the Hoffstadt Creek basin (Table 1) indicated annual erosion during 1980–1981 of 200 m^3/km^2, a rate from 1 to 2 orders of magnitude less than those measured in cleared or downed forests closer to the volcano.

Average Erosion in the Toutle River Basin

The computation of average erosion over the Toutle River basin during the first posteruption year was discussed in detail by Collins and others (1983). From regression analyses summarized above, erosion rates were assigned to all hillslopes according to their cover type, tephra thickness, tephra texture, and over-all gradient. The amount of sediment eroded was calculated by sampling the distribution of these controlling factors, using a grid of 498 equally spaced points. The total sediment yield was then calculated from erosion rates for each point. These calculations indicate that by May 1981, 12 × 10^6 t, or 11%, of the tephra on hillslopes in 1980 had been removed by water erosion. The grain-size composition of the eroded sediment was computed by using our textural analyses of the tephra and by assuming that sheetwash eroded only the surface air-fall layer and that rills eroded through the entire tephra layer.

Lehre and others (1983) surveyed the cross sections of stream channels draining our study sites throughout the year following the May 1980 eruption, demonstrating that sediment eroded from hillslopes was not stored on footslopes or in stream channels. By May 1981, 4 major lakes impounded by the debris avalanche on May 18 trapped 3.7 × 10^6 t, or 30%, of all sediment eroded from hillslopes. The remaining 8.4 × 10^6 t of sediment entered the Toutle River system; 36% was clay and silt, 58% was sand, and 6% was gravel.

Neglecting the effects of land management, the analysis indicated that 1.9 × 10^6 t of sediment were eroded during the second year, or only 16% of the first-year total. Results discussed by B. D. Collins and T. Dunne (unpub. data) indicate that erosion of scarified and salvage-logged hillslopes reduced the basin sediment yield by only 6% (0.1 × 10^6 t). During the third year, 0.9 × 10^6 t were eroded; land management was responsible for a 7% (0.06 × 10^6 t) reduction. Trapping of sediment in lakes reduced influx to the Toutle River to 1.1 × 10^6 and 0.5 × 10^6 t/yr in the second and third years, respectively, of which 30% was silt and clay, 65% was sand, and 5% was gravel.

Figure 10 shows the declining erosion of tephra and colluvium (circles and solid line) and of tephra alone (triangles and dashed line) for the Toutle River catchment. A power function fitted to the three points for tephra erosion, when projected into the future, predicts an average erosion rate of 0.5 mm/yr (700 $t/km^2/yr$) for 1984–1985 and of 0.1 mm/yr (140 $t/km^2/yr$) for 1989–1990. Integrating under the curve for tephra alone yields erosion of 39 × 10^3 t/km^2 (28 mm) in 1980–1982, 42 × 10^3 t/km^2 (30 mm) between 1980 and 1990, and only 700 t/km^2 (0.5 mm) from 1990 to 2000.

The curves do not include effects of land management. Scarification and logging practices used on parts of the basin would reduce erosion 0.2 mm (280 t/km^2) between 1980 and 1990. The curves are also extrapolated from rates that were measured prior to widespread recovery of plants, and it is likely that further growth in coming years will reduce the over-all erosion rate more than Figure 10 predicts (B. D. Collins and T. Dunne, unpub. data).

Neglecting effects of management and plant recovery, Figure 10 predicts that only ~1.5% of the tephra on hillslopes will be eroded by sheetwash and rill erosion before soil creep and other forms of mass wasting once again dominate sediment production.

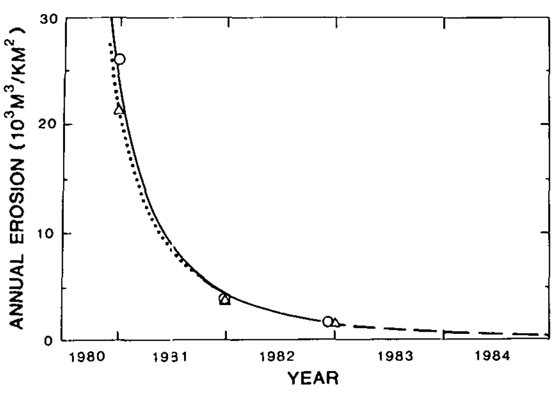

Figure 10. Measured and projected annual erosion of tephra and colluvium (circles and solid curve) and of tephra alone (triangles and dotted curve) for the Toutle River basin during the period 1980–1983 and the projected rate of erosion for 1980–1984. The regression equation for tephra and colluvium is $y = 8.0 \times {}^{-1.7}$ and for tephra alone is $y = 7.5 \times {}^{-1.6}$.

PATTERNS OF EROSION AND REVEGETATION AT OTHER VOLCANOES

Most other studies of recent tephra erosion suggest that, as at Mount St. Helens, the erosion rate peaked soon after the tephra erupted and then declined rapidly. The studies imply or specify that this decline was not caused by a recovery of vegetation, but instead by increased infiltration and decreased erodibility of the tephra layer, by exposure of more permeable and less erodible substrates, and, as our work further shows, by the development of a stable rill network. At Irazú, Waldron (1967) described an intensive rehabilitation program that included contour trenching and grass seeding while eruptions of tephra were in progress in February 1963 and ended in early 1965. The duration of runoff relative to precipitation decreased by the end of the first rainy season, when nearly one-third to one-half of the ash layer had been stripped by erosion. Although Waldron implied that the change resulted from the rehabilitation measures, the amount of erosion that he reported would have resulted in a widespread exposure of buried soil and in diminished runoff. Gullies on Fernandina Island remained unchanged after Hendrix's (1981) first visit two years after the eruption and before revegetation. Exposure of the buried soil at Parícutin was important in slowing the erosion rate (Segerstrom, 1966, p. 98–99), but erosion had also decreased on sparsely vegetated slopes where the tephra layer had not been stripped (Segerstrom, 1966, p. 98–99), implying that infiltration and erodibility of the tephra layer had also changed. The most rapid erosion of Volcan volcano in New Guinea occurred within the first four years after its creation in 1937, and revegetation did not occur until 1944, several years after the erosion rate had slowed (Ollier and Brown, 1971). Ollier and Brown's study did not include field measurements or observations until 31 yr after the eruption, but the importance of a stable rill network may be inferred, because the rills did not expose a permeable substrate.

Decreased erosion rates without exposure of buried substrates are demonstrated by two studies of Japanese volcanoes. At Mount Usu, the infiltration rate decreased after the ash eruptions, because the surface of the ash layer formed a "mortar-like crust due to . . . desiccation just after the big eruptions of 1977" (Kadomura and others, 1983, p. 132). The infiltration rate increased and the erosion rate decreased subsequently, because the surface "became porous as the result of surface erosion and leaching (Kadomura and others, 1978, p. 121). At Mount Sakurajima, similarly, there was a "big difference [in] the coefficient of permeability of the volcanic ash fall deposit between the time when the volcano is active and not active" (Shimokawa and Taniguchi, 1983, p. 168). Kadomura's are the only data that are sufficiently detailed to document that the erosion rate decreased more rapidly in rills than on divides, presumably because a stable rill network developed, and possibly because channels were lowered to a resistant layer with a high infiltration capacity.

ACKNOWLEDGMENTS

We had valuable discussions with F. J. Swanson, A. K. Lehre, W. E. Dietrich, B. Hallet, and L. H. Fairchild. A. K. Lehre helped to set out the first arrays of stakes and to develop a field methodology. Financial support was provided by the Pacific Northwest Forest and Range Experiment Station, Cooperative Agreement PNW-80-179; Washington State Department of Fisheries; the Weyerhaeuser Company; Sigma Xi; and the Geological Society of America.

REFERENCES CITED

Collins, B. D., 1984, Erosion of tephra from the 1980 eruption of Mount St. Helens [M.S. thesis]: Seattle, Washington, University of Washington, 181 p.

Collins, B. D., Dunne, T., and Lehre, A. K., 1983, Erosion of tephra-covered hillslopes north of Mount St. Helens: May 1980–May 1981: Zeitschrift für Geomorphologie Supplement, v. 46, p. 103–121.

Davies, D. K., Vessell, R. K., Miles, R. C., Foley, M. G., and Bovis, S. B., 1978, Fluvial transport and downstream sediment modifications in an active volcanic region, *in* Miall, A. D., ed., Fluvial sedimentology: Calgary, Alberta, Canadian Society of Petroleum Geologists, p. 61–84.

Dunne, T., 1980, Formation and controls of channel networks: Progress in Physical Geography, v. 4, p. 211–239.

Dunne, T., and Aubry, B. F., 1986, Evaluation of Horton's theory of sheetwash and rill erosion on the basis of field experiments, *in* Abrahams, A. D., ed., Hillslope processes: London, George Allen and Unwin, ch. 2.

Eggler, W. A., 1963, Plant life of Paricutin Volcano, Mexico, eight years after activity ceased: American Midland Naturalist, v. 69, p. 38–68.

Fiksdal, A., 1981, Infiltration rates of undisturbed and disturbed Mount St. Helens ash [abs.]: Association of Engineering Geologists Annual Meeting, 24th, Portland, Oregon, Abstracts, p. 34–35.

Franklin, J. F., and Dyrness, C. T., 1973, Natural vegetation of Oregon and Washington: U.S. Department of Agriculture Forest Service General Technical Report PNW-8, 417 p.

Griggs, R. F., 1918, The recovery of vegetation at Kodiak: Ohio Journal of Science, v. 19, p. 173–209.

——— 1919, The beginnings of revegetation in Katmai Valley: Ohio Journal of Science, v. 19, p. 318–342.

Hendrix, L. B., 1981, Post-eruption succession on Islas Fernandina, Galapagos: Madrono, v. 28, no. 4, p. 242–254.

Herkelrath, W. N., and Leavesley, G. H., 1981, Physical properties of volcanic ash from Mount St. Helens—Relationship to rain infiltration experiments: Agronomy Abstracts, p. 141.

Hoblitt, R. P., Miller, C. D., and Vallance, J. W., 1981, Origin and stratigraphy of the deposit produced by the May 18 directed blast, *in* Lipman, P. W., and Mullineaux, D. R., eds., The 1980 eruptions of Mount St. Helens, Washington: U.S. Geological Survey Professional Paper 1250, p. 401–409.

Horton, R. E., 1945, Erosional development of streams and their drainage basins: Bulletin of the Geological Society of America, v. 56, p. 275–370.

Kadomura, H., Yamamoto, H., Suzuki, R., and Suzuki, K., 1978, Some aspects of erosional form and processes on tephra-covered hillslopes produced by the 1977 eruption of Usu Volcano, Hokkaido, Japan: August 1977–early August 1978, *in* Usu eruption and its impacts on environment: Hokkaido, Japan, Hokkaido University, p. 121–139.

Kadomura, H., Imagawa, T., and Yamamoto, H., 1983, Eruption-induced rapid erosion and mass movements on Usu Volcano, Hokkaido: Zeitschrift für Geomorphologie Supplement, v. 46, p. 123–142.

Kieffer, S. W., 1981, Fluid dynamics of the May 18 blast at Mount St. Helens, *in* Lipman, P. W., and Mullineaux, D. R., eds., The 1980 eruptions of Mount St. Helens, Washington: U.S. Geological Survey Professional Paper 1250, p. 379–400.

Larson, K. R., and Sidle, R. C., 1980, Erosion and sedimentation data catalog of the Pacific Northwest: Portland, Oregon, U.S. Department of Agriculture Forest Service, 64 p.

Lehre, A. K., Collins, B. D., and Dunne, T., 1983, Post-eruption sediment budget for the North Fork Toutle River drainage, June 1980–May 1981: Zeitschrift für Geomorphologie Supplement, v. 46, p. 143–163.

Lipman, P. W., and Mullineaux, D. R., eds., 1981, The 1980 eruptions of Mount St. Helens, Washington: U.S. Geological Survey Professional Paper 1250, 844 p.

Miller, J. F., Frederick, R. H., and Tracey, R. J., 1973, Precipitation-frequency atlas of the western United States, Atlas 2, Volume 9: Silver Spring, Maryland, National Oceanic and Atmospheric Administration.

Moore, J. G., and Sisson, T. W., 1981, Deposits and effects of the May 18 pyroclastic surge, *in* Lipman, P. W., and Mullineaux, D. R., eds., The 1980 eruptions of Mount St. Helens, Washington: U.S. Geological Survey Professional Paper 1250, p. 421–438.

Mosley, P., 1973, Rainsplash and the convexity of badland divides: Zeitschrift für Geomorphologie Supplement, v. 18, p. 10–25.

Ollier, C. D., and Brown, M.J.F., 1971, Erosion of a young volcano in New Guinea: Zeitschrift für Geomorphologie, v. 15, p. 12–28.

Reid, L. M., Dunne, T., and Cederholm, J., 1981, Application of sediment budget studies to the evaluation of logging road impact: Journal of Hydrology (NZ), v. 20, p. 49–62.

Richards, A. F., 1965, Geology of the Islas Revillagigedo—3, Effects of erosion on Islas San Benedicto 1952–61 following the birth of Volcan Barcena: Bulletin Volcanologique, v. 28, p. 381–403.

Sarna-Wojcicki, A. M., Shipley, S., Waitt, R. B., Jr., Dzurisin, D., and Wood, S. H., 1981, Areal distribution, thickness, mass, volume, and grain size of air-fall ash from the six major eruptions of 1980, *in* Lipman, P. W., and Mullineaux, D. R., eds., The 1980 eruptions of Mount St. Helens, Washington: U.S. Geological Survey Professional Paper 1250, p. 577–600.

Segerstrom, K., 1950, Erosion studies at Parícutin, state of Michoacan, Mexico: U.S. Geological Survey Bulletin 965-A, 164 p.

——— 1960, Erosion and related phenomena at Parícutin in 1957: U.S. Geological Survey Bulletin 1104-A, p. 1–18.

——— 1961, Deceleration of erosion at Parícutin, Mexico: U.S. Geological Survey Professional Paper 424-D, p. D225–D227.

——— 1966, Parícutin, 1965—Aftermath of eruption: U.S. Geological Survey Bulletin 550C, p. 93–101.

Shimokawa, E., and Taniguchi, Y., 1983, Debris yield from hillside slope of active volcano, *in* Symposium on Erosion Control in Volcanic Areas, Seattle, July 1982: Ibaraki, Japan, Public Works Research Institute, p. 155–181.

Swanson, F. J., Fredriksen, R. L., and McCorison, F. M., 1982, Material transfer in a western Oregon forest ecosystem, *in* Edmonds, R. L., ed., Analysis of coniferous forest ecosystems in the western United States: Stroudsburg, Pennsylvania, Hutchinson and Ross, Inc., p. 233–266.

U.S. Department of Agriculture Soil Conservation Service and U.S. Weather Bureau, 1965, Mean annual precipitation, 1930–1957, State of Washington.

Voight, B., Glicken, H., Janda, R. J., and Douglass, P. M., 1981, Catastrophic rockslide avalanche of May 18, *in* Lipman, P. W., and Mullineaux, D. R., eds., The 1980 eruptions of Mount St. Helens, Washington: U.S. Geological Survey Professional Paper 1250, p. 347–377.

Waitt, R. B., Jr., 1981, Devastating pyroclastic density flow and attendant air fall of May 18—Stratigraphy and sedimentology of deposits, *in* Lipman, P. W., and Mullineaux, D. R., eds., The 1980 eruptions of Mount St. Helens, Washington: U.S. Geological Survey Professional paper 1250, p. 439–458.

Waldron, H. H., 1967, Debris flow and erosion control problems caused by the ash eruptions of Irazú Volcano, Costa Rica: U.S. Geological Survey Bulletin 1241-I, p. I1–I37.

Manuscript Received by the Society May 23, 1985
Revised Manuscript Received February 19, 1986
Manuscript Accepted February 20, 1986