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INITIAL HYDROLOGIC AND GEOMORPHIC RESPONSE FOLLOWING A WILDFIRE IN THE COLORADO FRONT RANGE

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ABSTRACT

A wildfire in May 1996 burned 4690 hectares in two watersheds forested by ponderosa pine and Douglas fir in a steep, mountainous landscape with a summer, convective thunderstorm precipitation regime. The wildfire lowered the erosion threshold in the watersheds, and consequently amplified the subsequent erosional response to shorter time interval episodic rainfall and created both erosional and depositional features in a complex pattern throughout the watersheds.

The initial response during the first four years was an increase in runoff and erosion rates followed by decreases toward pre-fire rates. The maximum unit-area peak discharge was 24 $m^3 s^{-1} km^{-2}$ for a rainstorm in 1996 with a rain intensity of 90 mm h^{-1}. Recovery to pre-fire conditions seems to have occurred by 2000 because for a maximum 30-min rainfall intensity of 50 mm h^{-1}, the unit-area peak discharge in 1997 was 6.6 $m^3 s^{-1} km^{-2}$, while in 2000 a similar intensity produced only 0.11 $m^3 s^{-1} km^{-2}$. Rill erosion accounted for 6 per cent, interrill erosion for 14 per cent, and drainage erosion for 80 per cent of the initial erosion in 1996. This represents about a 200-fold increase in erosion rates on hillslopes which had a recovery or relaxation time of about three years. About 67 per cent of the initially eroded sediment is still stored in the watersheds after four years with an estimated residence time greater than 300 years. This residence time is much greater than the fire recurrence interval so erosional and depositional features may become legacies from the wildfire and may affect landscape evolution by acting as a new set of initial conditions for subsequent wildfire and flood sequences. Published in 2001 by John Wiley & Sons, Ltd.

KEY WORDS: erosion; deposition; landscape evolution; sediment transport; wildfire

INTRODUCTION

Wildfire is a natural disturbance process that affects the interactions between biotic and abiotic components of an ecosystem and is frequently affected by human activities. Current ecological models treat wildfire as one component of the disturbance regime of many terrestrial ecosystems (Rogers, 1996). As a geomorphic agent wildfire can lower the intrinsic threshold of erosion in a watershed (Schumm, 1973), increase runoff and erosion (Rowe *et al.*, 1954; Krammes, 1960; Doehring, 1968; Anderson, 1974; Florsheim *et al.*, 1991) and cause rock weathering (Blackwelder, 1927). The magnitude of the disturbance depends upon the sensitivity of the system to erosion (Brunsden, 1980), the precipitation regime, the severity and areal extent of the burn, and the frequency of wildfires (Swanson, 1981). Wildfires change the infiltration properties of soils on a hillslope and reduce the amount of interception materials (canopy, litter, duff and organic debris) which protect the soil from raindrop impact and slow runoff. Infiltration generally decreases because chemical (DeBano, 1969; DeBano *et al.*, 1977; Giovannini *et al.*, 1988) and physical (Doehring, 1968; Wells, 1981; Giovannini and Lucchesi, 1983; Giovannini *et al.*, 1988) properties of the soil may have changed, making the soil more water-repellent. The magnitude of the hydrological and sedimentological response during the period after a wildfire or any natural disturbance can be quantified as the change in process rates (water discharge, hillslope

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and channel erosion, sediment transport, and deposition) from pre-fire conditions (Swanson, 1981). During the initial phase, process rates increase with time and reach a maximum and during the recovery phase, rates decrease; the length of time of these two phases constitutes the relaxation time. Relaxation times vary and may take only one to three years for sediment fluxes, up to thirty years for water discharges, and even longer for the regrowth of trees (Rowe *et al.*, 1949, 1954; Doehring, 1968; Scott and Williams, 1978; Swanson, 1981; Laird and Harvey, 1986; Potyondy and Hardy, 1994).

Geomorphic effects of and responses to wildfire depend upon the precipitation climatology and may be viewed at different temporal scales. Wildfires are common in the western United States where the climate is characterized, in general, by periods of very dry weather and periods of episodic, intense rainfall. At short time intervals, the hydrologic response to intense rainfall can be life-threatening and may cause economically damaging floods coupled with sediment impacts on recreation, aquatic biota, and water-supply systems. At long time intervals, the importance of wildfire as a geomorphic agent depends not only upon process rates integrated over long time intervals, but also upon the ratio of the relaxation time to the wildfire recurrence time, defined by Brunsden and Thornes (1979) as the transient form ratio. The mean wildfire recurrence interval for ponderosa pine-dominated ecosystems ranges from 20 to 50 years along the Colorado Front Range (M. Gonzalez, 2000, pers. comm.; Kaufmann *et al.*, 2000; Veblen *et al.*, 2000). However, relaxation intervals for this area are unknown and may depend upon the precipitation climatology. For example, several authors have hypothesized that channels in semi-arid climates are shaped by episodic rainfall (Thornes, 1976; Wolman and Gerson, 1978; Kochel, 1988; McCord, 1996) rather than more frequent, moderate rainfall that shapes more humid systems (Wolman and Miller, 1960). Thus, in semi-arid landscapes, channel forms may persist for a much longer time than in humid landscapes where the recurrence interval of the mean annual flood is 1 to 1.58 years (Wolman and Leopold, 1957; Leopold *et al.*, 1964; Dury, 1973). Persistence of erosional features can be measured by the refilling time while persistence of stored sediment can be measured by the residence time defined as the storage volume divided by the annual transport rate (Dietrich and Dunne, 1978; Swanson 1981).

At the scale of an individual fire, spatial inhomogeneities exist in the distribution of hillslope aspects and gradients, vegetation and burn severity. When these inhomogeneities are overlain by a different and inhomogeneous spatial pattern of intense convective rainfall, the result is a complex response (Schumm and Lichty, 1965; Schumm, 1973) of sediment transport and corresponding erosional and depositional features that can vary depending upon the spatial scale of each subwatershed (Trimble, 1995). Rainfall in arid, semi-arid and mountainous systems is often characterized by convective storms. These storms produce high-intensity episodic rainfall which has triggered some of the largest historical floods (Costa, 1987). Because wildfires lower the erosion threshold, even infrequent small-intensity rainfall when coupled with the spatial heterogeneity of wildfire effects may represent an important short-time-interval, geomorphic agent in arid and semi-arid mountainous systems. Thus, episodic wildfires indirectly create perturbations in the sediment budget that are manifested as time-varying transport rates and as changes in sediment-storage volumes that produce a complex pattern of erosional and depositional forms across the landscape.

This study reports on the hydrologic and geomorphic response after a wildfire in a semi-arid, mountainous system where the rainfall regime is characterized by short-duration, high-intensity summer rainfall. The short-term, transient response was measured during the first four years after the wildfire. The process rates are compared to available pre-fire rates for the Colorado Front Range and to rates in the chaparral mountainous system of southern California with a mediterranean semi-arid climate to assess the geomorphic importance of wildfire. These rates can then be utilized to model long-term landscape evolution in steep mountainous areas that are prone to wildfire and subject to intense rainfall.

BACKGROUND

Site description

Buffalo and Spring Creek watersheds are located within the Pikes Peak granite batholith (Figure 1), have shallow soils, and are sparsely forested by ponderosa pine (*Pinus ponderosa*) on south-facing slopes and have a higher density of Douglas fir (*Pseudotsuga menziesii*) on north-facing slopes. The Spring Creek watershed

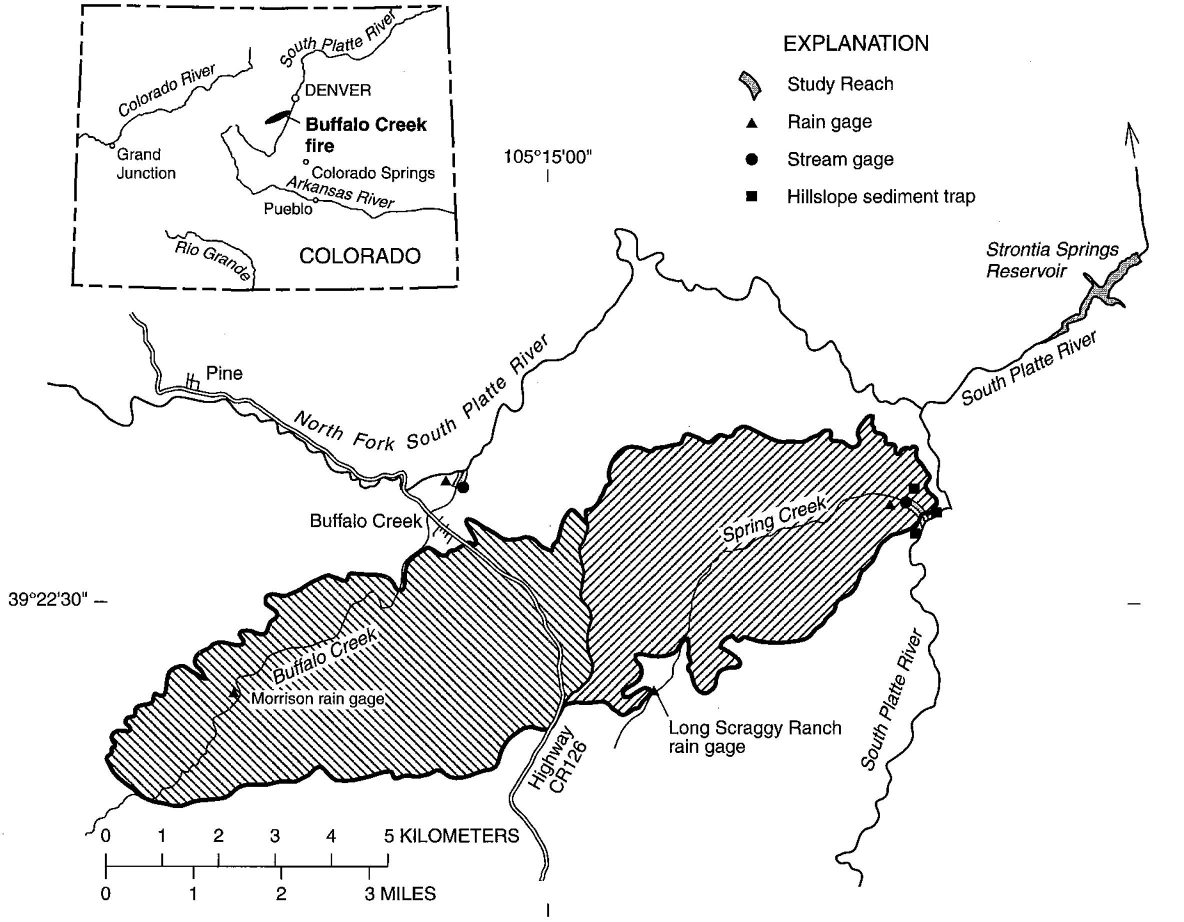

Figure 1. Location of the portions of two watersheds burned by the Buffalo Creek Fire (two crosshatch patterns), study reaches, and US Geological Survey rain and stream gauges (06706800, Buffalo Creek at mouth of Buffalo Creek, Colorado; 06701970, Spring Creek above mouth near South Platte, Colorado; 392133105184401, Buffalo Creek rain gauge at Morrison Creek, Colorado; and 392144105132401, Spring Creek rain gauge at Long Scraggy Ranch, Colorado). All hillslope sediment traps are within the Spring Creek watershed. Only 21 per cent of the Buffalo Creek watershed burned while 79 per cent of the Spring Creek watershed burned

is steeper but smaller (20–30 per cent slopes and 2680 hectares) than the Buffalo Creek watershed (10–20 per cent slopes and 12 240 hectares). The channel density (about 7·0 km^{-1}) is similar for both watersheds and higher than older Piedmont watersheds (4·3 km^{-1}; Schumm, 1956) but lower than watersheds in the granitic mountains of southern California (9·7 km^{-1}; Schumm, 1956). Soils are generally coarse (about 7 per cent silt and clay, 35 per cent sand, 58 per cent gravel) with a median diameter of 2·6 to 2·9 mm (Martin and Moody, 2001). On south-facing hillslopes, soils are Typic Ustorthents and 3–4 m thick; and on north-facing hillslopes, they are Typic Cryorthents and 6–30 m thick (Blair, 1976; Moore, 1992; Welter, 1995). These soils have a typical erodibility factor, K, of 0·49 m^{-1}, a high runoff potential when thoroughly wet (primarily because of the very shallow depth to bedrock), and are considered to be highly erodible if the soil cover is disturbed (Moore, 1992).

The Buffalo Creek Fire (18–24 May 1996) was a fast-moving wildfire that began near the town of Buffalo Creek in the Colorado Front Range southwest of Denver, Colorado (Agnew *et al.*, 1997). The crown fire was wind driven and burned 4690 hectares in the Buffalo and Spring Creek watersheds in a cigar-shaped pattern oriented downwind. More than 60 per cent of the burned area was designated as a high-intensity burn defined as a complete burning of the forest duff, heating of the soil surface, and combustion of all fine fuels

Table I. Characteristics of Buffalo and Spring Creeks watersheds

Characteristics	Buffalo Creek	Spring Creek
Watershed level	6	6
Watershed area (ha)	12 240	2680
Burned area (ha)	2570	2120
Elevation range (m)	2010–3180	1880–2360
Relief ratio in the burned area	0·020	0·046
Main channel length in burned area (km)	7·3	5·9
Channel lengths in burned area (km)	180[a]	150
Bifurcation ratio	3·9[b]	4·1
Average valley width near mouth (m)	35	27
Range in channel width near mouth (m)	3–13	1–26
Main channel slope (%)	1–2	3–4
Channel density (km^{-1})	7·1[b]	6·9
Distance of mouth from Strontia Springs Reservoir (km)	18	4·8
Baseflow: June, July, August 1997–98 (m^3 s^{-1})	0·7	0·07

[a] Channel density × burned area
[b] Average of three subwatersheds

in the canopy (Bruggink *et al.*, 1998). Only 21 per cent of the lower part of the Buffalo Creek watershed was burned compared to 79 per cent of the Spring Creek watershed (Figure 1; T. Clark, 1996, pers. comm.); thus, fewer of the steep channels were affected by the wildfire in the Buffalo Creek than in the Spring Creek watershed (Table I). Several rain storms in June and July 1996 caused flash floods in both watersheds and transported sediment and fire-related debris into Strontia Springs Reservoir, a water-supply reservoir for the cities of Denver and Aurora (Figure 1).

Precipitation regime

The burned area is in a semi-arid climate with a precipitation regime characterized by summer convective storms and winter snow storms. The mean annual precipitation is about 440 mm and about half of this is from approximately 2100 mm of snow (maps 16, 19 and 21, Hansen *et al.*, 1978). The area is located within a severe thunderstorm zone (Henz, 1974), along the foothills of the Colorado Front Range. During the summer (defined in this paper as June, July, August and September; 122 days) when storms are most frequent, average summer rainfall at two nearby weather stations exceeds 2·5 mm d^{-1} for 16·6 to 22·1 days each summer (Kassler, 1950–1995, and Cheesman, 1951–1995). The burned area is in a region that has rapid changes in rainfall over short distances. The estimates for the maximum 30-min rainfall intensity for the 2-, 10- and 100-year recurrence intervals were 30, 50 and 90 mm h^{-1}, after interpolating between isopluvials on charts published by Miller *et al.* (1973).

METHODS

The hydrologic response to the wildfire was determined by measuring rainfall and runoff in the watershed while the geomorphic response was determined by estimating or measuring sediment erosion rates on hillslopes (interrill and rill), changes in sediment storage, and sediment transport rates out of the watersheds and into a downstream reservoir. More details about these methods are reported by Moody and Martin (2001).

Rainfall–runoff

One rain gauge was installed in the upper portion of each burned watershed and a rain gauge and stream gauge were installed near the mouth of both Buffalo and Spring Creeks (Figure 1). These were maintained by the US Geological Survey from April to September and provided digitized values at 5-min intervals (US Geological Survey, 1997–1999). These data were used to determine duration, total accumulation and intensity of rainfall and peak discharge of runoff. Rainfall intensity was measured as I_{30}, the maximum amount of rain during a 30-min period and expressed as millimetres per hour.

Interrill erosion

We measured interrill erosion rates caused by natural precipitation events for four years (1997–2000). During 1997, two study areas were selected in a high-severity burned area within the Spring Creek watershed – one north aspect and one south aspect (Figure 1). After 1997, two additional areas were selected in an unburned area – one north aspect and one south aspect. All the study areas were on about 50 per cent slopes in the Spring Creek watershed with a median slope of 20–30 per cent. In each area, we installed four modified versions of 1-m wide, Gerlach sediment trap (Gerlach, 1967; Fitzhugh, 1992; Moody and Martin, 2001) to collect water and sediment. Sediment traps were located near the channel to measure the actual delivery of sediment to the channel. Visual observations throughout the watershed indicate that hillslopes were mostly convex with little storage at the base of the hillslope.

Four hillslope sediment traps permit a small sample-size estimate of the 95 per cent confidence limits for mean sediment flux. We express the transport of the soil down interrill areas and down rills as a flux of sediment mass across a 1-m wide contour per day (kg m^{-1} d^{-1}) and we normalize these fluxes by the amount of rainfall during each collection period. Additionally, in order to compare our data with other published data we also computed sediment yields as the mass per unit burned area per year. However, converting sediment fluxes to sediment yields can be misleading because the extent of the upslope contributing area is unknown even though hillslope plots are often bounded. We express the yield as kg ha^{-1} a^{-1} which represents a relatively small extrapolation from the actual size of our hillslope plots. Additional sediment traps consisting of tarps (1·7 m by 2·3 m) were anchored to the ground at the base of south-facing and north-facing hillslopes to collect the flux of sediment to the main channel of Spring Creek as a result of the freeze–thaw and snowmelt processes.

Rill erosion

Rill erosion during two major floods in 1996 and 1997 was estimated from aerial photographs and field measurements. The number and spatial distribution of rills on hillslopes were counted and mapped on aerial photographs (1 : 3000 scale) of two subwatersheds in the Spring Creek watershed. One subwatershed, W960, is a south-facing, third-order watershed with an area of 7·0 ha and an estimated channel density of 21 km^{-1} after the wildfire; while W1165 is a north-facing, fourth-order watershed with an area of 3·7 ha and an estimated channel density of 48 km^{-1}. Additional field measurements of rill length and cross-sectional area were made in W960, W1165 and in other subwatersheds. The eroded volumes for these two subwatersheds were calculated as the product of the mean cross-sectional area, mean rill length, and the number of rills that actually delivered sediment to the channels. The yield (volume per unit area) was then extrapolated to estimate the rill erosion volume for the Spring and Buffalo Creek watersheds.

We measured rill erosion rates caused by natural precipitation events for three years (June 1998–November 2000). Three sediment traps were installed on the south aspect of a high-severity burned area near the interrill sediment traps. Each sediment trap was located on a different rill at various distances from the head of the rill and collected water and sediment.

Sediment storage

Erosion in drainages was measured in the two Spring Creek subwatersheds (W960 and W1165) following two major floods in 1996 and 1997 and the results were extrapolated to estimate the initial erosion in the Spring and Buffalo Creek watersheds. Drainages may be unchannelized with no inflection point in a cross-sectional profile or they may be channelized with an inflection point. Our estimates of drainage erosion include pre-fire channels and drainages channelized by post-fire erosion. Erosion (volume of stored sediment lost per unit channel length or the cross-sectional area) was determined every 5 m along all drainages. The pre-flood surface was estimated by extrapolating the post-flood surface across the channel or by using tree roots left exposed after the floods along the sides of the channels, which in some cases were unbroken and spanned the entire channel.

Changes in storage (erosion and deposition) of sediment in the main channel of Spring Creek were measured for five years by aerial photogrammetry and ground survey methods. Photogrammetry was used to determine the initial erosion and deposition in 1996. Later, a series of closely spaced channel cross-sections in the study reach near the mouth of each watershed (Figure 1) was resurveyed between June 1997 and October 2000. Valley widths were 25–35 m and cross-sections were initially spaced 10 m apart to measure the volume within each study reach. Each study reach started at the mouth and extended upstream to the stream gauge. The study reach in Buffalo Creek was 480 m long and in Spring Creek it was 1490 m long. We found that changes in volume at several neighbouring cross-sections were similar during 1997, thus in 1998, 1999 and 2000 the interval between cross-sections was increased to approximately 30 m. Initially, the absolute location and elevations of each cross-section were measured with an electronic surveying instrument (Nikon 720 DTM) and then they were remeasured with an automatic level, metric tape and surveying rod.

Channel sediment transport

A sediment rating-curve was established for the mouth of each watershed by collecting bedload and suspended-load samples and by measuring the water discharge at selected times during the year. Four replicate samples of bedload were collected during steady flow conditions using a modified Helly-Smith bedload sampler (Emmett, 1980; Hubbell *et al.*, 1986) referred to as the US BLH-84 (Druffel *et al.*, 1976; USGS, 1990; Ryan and Porth, 1999; Moody and Martin, 2001). Suspended load was collected between each bedload replicate using a 0.450 l, pint-jar fitted with a cap and a 3 mm diameter isokinetic nozzle (Edwards and Glysson, 1988; Meade and Stevens, 1990). One discharge measurement was made between replicates two and three using a Price AA current meter or surface floats when the water depth was too shallow for the current meter.

Indirect measurements were used to determine sediment volume and discharge during flash floods when it is too dangerous to sample sediment or to measure water discharge directly. Sediment volumes transported by these flash floods were deposited in the study reach (which expands downstream from 8 m at the upper end

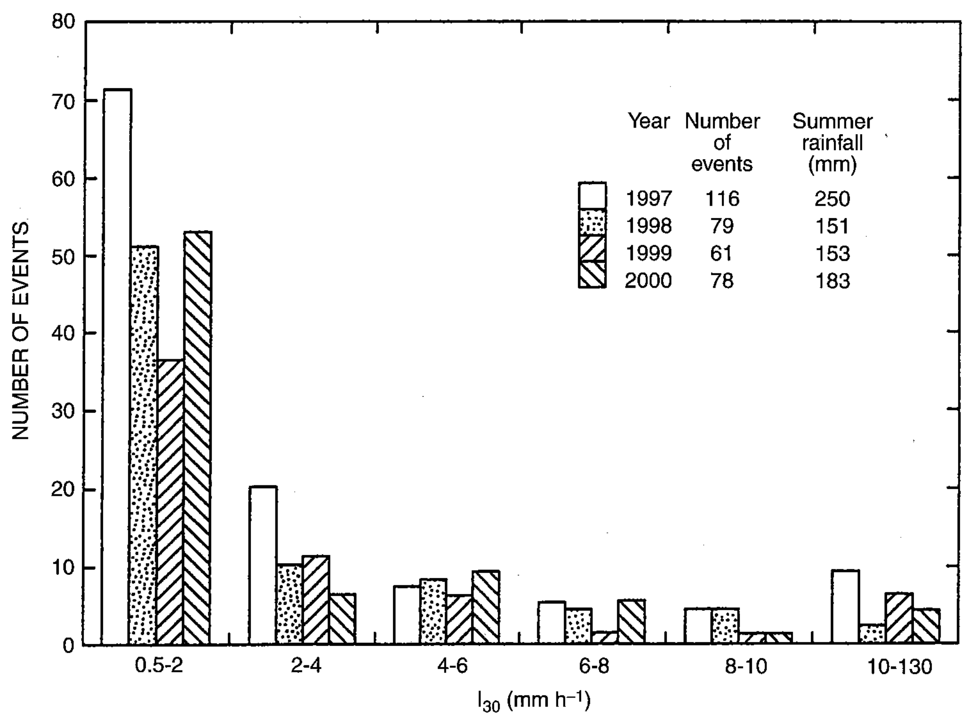

Figure 2. Histograms of the distribution of rainstorms by intensity class during June, July, August and September for 1997, 1998, 1999 and 2000 for the rain gauge at Spring Creek above mouth near South Platte, Colorado

to 60 m at the mouth) and during one extreme flood (31 August 1997) some sediment from the Spring Creek watershed was deposited as a temporary dam across the South Platte River. Sediment volumes were calculated from the channel surveys before and after each flash flood (Moody and Martin, 2001). These volumes were certainly a minimum estimate as some sediment was transported down the South Platte River; however, field observations indicate this was probably much smaller than the volume of sediment deposited in the study reach and in the temporary dam. Peak discharges were determined by high-water marks and slope-area method (Dalrymple and Benson, 1976). For flood events without high-water marks, a simple analytical model was developed to predict the peak discharge and duration based on Nash's linear reservoir model (Nash, 1958). Buffalo Creek could not be modelled by this method because most of the volume was transported directly into the North Fork of the South Platte River and not stored within the study reach. Total annual output for each watershed was based on the sediment rating-curve, the discharge record between flash floods, and estimates of the volume of sediment moved during each flash flood.

Sediment transport into the Strontia Springs Reservoir was estimated by measuring the difference in the stored volume divided by the time between bathymetric surveys. Each survey was a single longitudinal profile of the bottom along the reservoir axis recorded on a continuous strip-chart (Lowrance Model X16). These profiles were in the upper end of the reservoir where it was narrow (15–100 m). Here, a single profile was sufficient to represent the average bottom elevation and avoided the problems of sound scattering from the steep canyon walls. The volume of coarse sediment (sand and gravel) that accumulated in the upper end of Strontia Springs Reservoir between bathymetric surveys was calculated from a digital contour map of the reservoir and the longitudinal bathymetric profiles. Any conversion between sediment volume and sediment mass in this paper is based on an average bulk density (1700 kg m^{-3}) of collected bedload material.

RESULTS

Rainfall–runoff

The number and intensity of rainstorms in the two watersheds varied and the number of flash floods decreased during the four-year study period. During the summer of 1997, the four gauges had an average of 25 days with more than 2·5 mm of rain, but in 1998, 1999 and 2000 there were 19, 16 and 18 days, which are similar to the long-term average of 19 days for the Kassler and Cheesman weather stations. In general, more rainstorms occurred in 1997 in each intensity class (Figure 2) than in 1998, 1999 or 2000, and values of I_{30} greater than about 10 mm h^{-1} caused flash floods (Table II) which we have defined to have peak discharges about ten times the average summer baseflow (Table I).

Unit-area peak discharges (peak discharge divided by drainage area) in Buffalo and Spring Creeks decreased after the first significant precipitation event on 12 July 1996. In Spring Creek this storm produced runoff of 24 m^3 $s^{-1}km^{-2}$ of burned watershed or 80 per cent (72 mm h^{-1}) of the average estimated I_{30} (90 mm h^{-1}) during that storm. This unit-area peak discharge was about one-half the unit discharge for the largest rainfall–runoff floods for Colorado (Jarrett, 1990). For Spring Creek, unit-area peak discharge is related to I_{30} by a power law, which changes slope (or exponent) at about 5 to 10 mm h^{-1} (Figure 3). For a given rainfall intensity the unit-area peak discharge appears to decrease slightly from 1997 through 2000.

Interrill erosion

Estimates of the pre-fire erosion rates were made by measuring the summer sediment flux on neighbouring unburned hillslopes in 1998 and 1999. This flux was $0{\cdot}14 \pm 0{\cdot}02$ kg m^{-1} (Martin and Moody, 2001) and was similar to sediment fluxes (0·0–1·0 kg m^{-1}) measured in other unburned areas of the Colorado Front Range (Bovis, 1974; Morris, 1983; Morris and Moses, 1987; Welter, 1995).

Estimates of interrill erosion during the first summer after the wildfire (1996) were based on indirect evidence. Aerial and oblique photographs showed that the grey ash was removed and the hillslope was rilled after the rainstorms during June and July of 1996. Small pedestals of sediment capped and protected by stones or resistant material (can lids, horseshoes, etc.) stood 5 to 10 mm above the ground surface and were assumed to indicate the depth of erosion in interrill areas in 1996 throughout the burned watersheds. This estimate is similar to values for other burned watersheds (Table III). If we assume an erosion depth of 5 mm and a zone

Table II. Change in flood characteristics in the watersheds burned by the Buffalo Creek Fire: 1996–2000

Date	Buffalo Creek Watershed			Spring Creek Watershed		
	I^{a}_{30} (mm h^{-1})	Peak discharge (m^3 s^{-1})	Unit-area discharge m^3 s^{-1} km^{-2})	I^{b}_{30} (mm h^{-1})	Peak discharge (m^3 s^{-1})	Unit-area discharge (m^3 s^{-1} km^{-2})
1996						
12 June	na	na	na	na	20	0·94
12 July	80[c]	450[e]	18	~90[c,d]	510[e]	24
23 August	~30	40[d]	1·6	na	30	1·4
14 September	10–18[d]	5	0·2	na	7	0·33
1997						
6 June	19·2	13	0·51	14·0	0·0057	0·00027
28 July	15·2	13	0·51	12·2	1·1	0·052
29 July	15·2	30·5[f]	1·2	19·1	5·0[f]	0·24
31 July	29·6	8·3	0·32	32·4	3·6	0·17
2 August	8·1	8·2	0·32	2·5	0·014	0·00066
9 August	12·2	9·9	0·39	10·2	0·57	0·027
26 August	11·5	0·7	0·027	19·6	6·6	0·31
31 August	7·9	5·3	0·21	51·9	140[f]	6·6
1998						
8 July	5·0	No increase above baseflow		12·4	0·020	0·00094
9 July	3·2	No increase above baseflow		25·6	48[f]	2·3
31 July	30·5	Gauge damaged		44·8	82[f]	3·9
31 August	5·0	0·11	0·0043	10·4	0·0085	0·00040
1999						
8 July	2·5	No increase above baseflow		10·6	0·014	0·00066
11 July	9·6	0·20	0·0078	15·0	0·062	0·0029
17 July	13·8	No increase above baseflow		20·8	0·040	0·0019
28 July	7·4	No increase above baseflow		25·4	0·14	0·0066
29 July	15·5	5·1	0·20	18·2	6·4	0·30
4 August	6·8	0·080	0·0031	15·2	0·91	0·043
17 August	0·5	No increase above baseflow		12·0	0·15	0·0071
2000						
16 July	20·0	No increase above baseflow		49·2	2·4	0·11
17 July	36·6	No increase above baseflow		19·2	0·065	0·0031
20 August	2·2	0·028	0·0011	10·6	0·031	0·0015

Includes floods in either watershed when the peak discharge was greater than ten times the baseflow for June, July and August 1997 and 1998 or when the maximum 30-min intensity, I_{30}, was greater than 10 mm h^{-1}.
[a] Average rainfall intensity of the Morrison and Buffalo Creek gauges.
[b] Average rainfall intensity of the Long Scraggy Ranch and Spring Creek gauges.
[c] This is an average of the maximum 1-h intensities of 110 mm h^{-1} at Long Scraggy Ranch and 75 mm h^{-1} near the Spring Creek gauge (Henz, 1998; Jarrett, 2001).
[d] R. Jarrett, pers. comm., 1996.
[e] Yates *et al.* (2000)
[f] Indirect discharge measurement.

of no erosion starting at the hillslope ridge of 5 m (based on visual estimates), then the interrill erosion for W960 is 280 m^3 and for W1165 it is 90 m^3 or an average of 32 m^3 ha^{-1}. Based on yield, the response in the 1996 was a 240-fold increase for the south-facing watershed and a 150-fold increase for the north-facing watershed (Table III). Extrapolating this yield to Spring and Buffalo Creek watersheds gives an estimate of the sediment yield to the channels in the 1996 of 150 000 m^3.

Measurements of interrill erosion rates, during the first year after the wildfire (1997), indicated more sediment was eroded from north- than from south-facing severely burned hillslopes. The average sediment

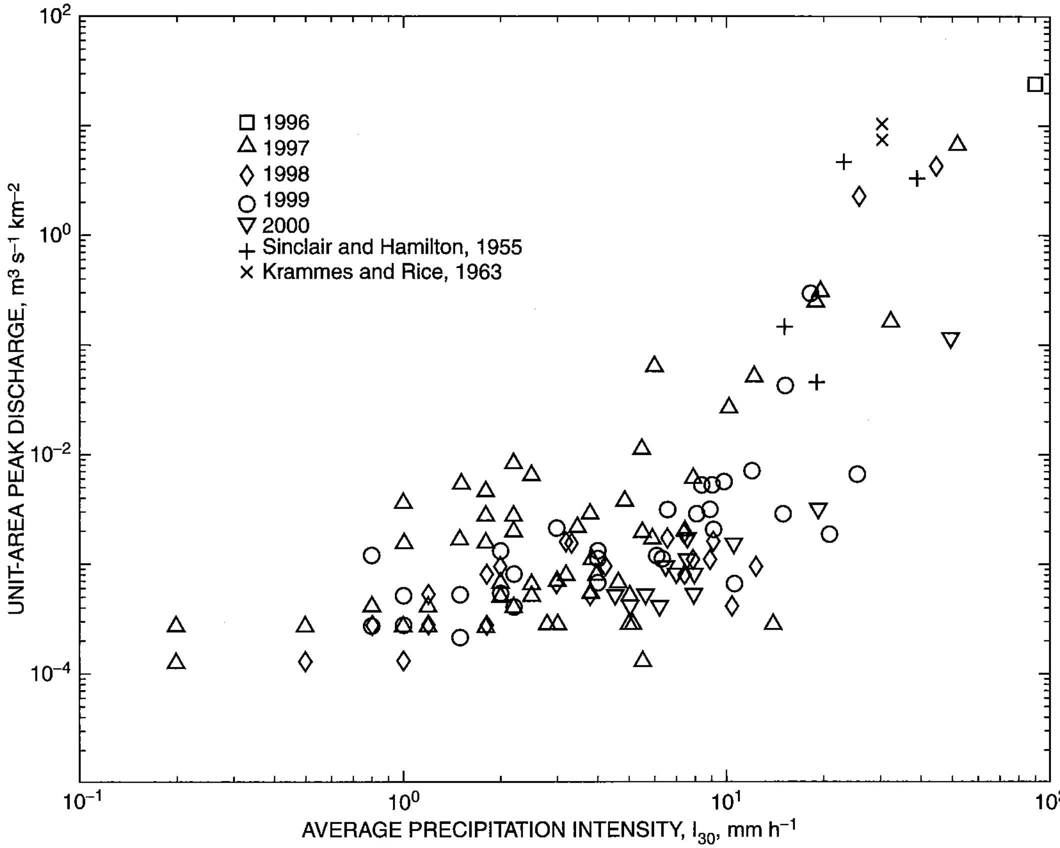

Figure 3. Spring Creek discharge response to rainfall intensity. The I_{30} values are the average for rain gauges at Long Scraggy Ranch and Spring Creek

flux during the summer was at least 0·048 kg m^{-1} d^{-1} from north-facing and 0·0070 kg m^{-1} d^{-1} from south-facing hillslopes. These values are minimal estimates of the sediment flux because in 1997, the study began in late July, missing the sediment transport by rainfall in June and early July (Figure 4A) and the rainstorm on 31 August 1997 overwhelmed the sediment traps and only part of the eroded sediment was collected on the north-facing hillslope. The sediment fluxes for the summer of 1997, 5·9 and 0·85 kg m^{-1}, are similar to fluxes (2·9–4·0 kg m^{-1}) reported by Morris and Moses (1987) within the first year after another wildfire in the Colorado Front Range. One estimate of the average sediment yield to the stream channels in the burned watersheds is to assume all the sediment collected by the sediment traps would have been delivered to the channel. Then the yield is twice the product of the channel length in the burned areas (Table I) and the average summer sediment flux (3·4 kg m^{-1}) or 1300 m^3.

Average interrill erosion rates (north- and south-facing) decreased during the second, third and fourth summers after the wildfire (1998, 1999 and 2000). This decrease was not a result of less rainfall because the severely burned north-facing slopes produced significantly more sediment per millimetre of rainfall in 1997 than in 1998, 1999 or 2000 (Figure 4B). During the second summer, the average erosion rate was 0·22 kg m^{-1} or about twice the pre-fire erosion rates, and during the third and fourth summers after the wildfire, the average was 0·10 and 0·066 kg m^{-1}, similar to pre-fire erosion rates. Estimates of the average sediment yield to the stream channels by interrill erosion are 85 m^3, 40 m^3 and 26 m^3 for 1998, 1999 and 2000 respectively. The period of accelerated hillslope erosion was very short and occurred primarily in the first and second year after the wildfire, similar to what has been observed by others in the San Gabriel Mountains (Doehring, 1968; Rice,

Table III. Equivalent depths of erosion on hillslopes and sediment yields in selected unburned and burned locations

Location	Equivalent depth of erosion after fire[a] (mm)	Yield ($kg\ ha^{-1} y^{-1}$)		
		Unburned	Burned	Response
San Gabriel Mountains, California	0.5–28[b]	880 ± 870[c]	45 600[d] 146 000[e]	50-fold 170-fold
San Gabriel Mountains, California	0. 5–28[b]	880 ± 870[c]	73 000[f] 890 000[f]	80-fold 1000-fold
Bandelier National Monument, New Mexico[g]	1–6	Unknown	130 000	Unknown
Frasier Experiment Forest, Colorado[h]	0·002	39 ± 16	na	na
Spring Creek, subwatershed W960, Colorado	5	280 ± 49	68 000	240-fold
Spring Creek, subwatershed W1165, Colorado	5	280 ± 49	41 000	150-fold
Australia[i]	0.1–2.8	Unknown	2500 48 000	Unknown
South Africa[j]	0.06–3.5	Unknown	1000 60 000	Unknown

[a]A bulk density of 1700 kg m^{-3} was assumed.
[b]Troxell and Peterson (1937); Rowe *et al.* (1951); Sartz (1953); Krammes (1960); Copeland and Croft (1962); Doehring (1968); Striffler and Mogren (1971); Atkinson (1984)
[c]Anderson *et al.* (1959)
[d]Krammes (1960): north-facing hillslopes
[e]Krammes (1960): south-facing hillslopes
[f]Kotok *et al.* (1935); Rowe *et al.* (1951); Krammes and Osborn (1969)
[g]Wells (1978)
[h]Leaf (1966): subalpine forests
[i]Blong *et al.* (1982); Atkinson (1984)
[j]Scott and van Wyk (1992); Scott (1993)

1974; Wells, 1986). Sediment fluxes decreased to near pre-fire rates after three years and, thus, the relaxation time for hillslope erosion rates is about three years.

Rill erosion

Rill erosion during the first summer after the wildfire (1996), was estimated for north- and south-facing hillslopes. Mean rill length was estimated as the average length of overland flow (Horton, 1945) minus the length of the zone of no erosion starting at the hillslope ridge (about 5 m). Some characteristics of rills include: (1) the number of rills in the two subwatersheds was essentially the same (319 in W960 and 370 in W1165) with an average rill spacing of about 10 m (some hillslopes in the watershed had no rills); (2) the average rill top-width at the base of the hillslope was 0·36 m (includes rills in other subwatersheds) or about 3·6 per cent of the hillslope; (3) the average rill length was about 20 m in W960 and 5 m in W1165; and (4) the average rill cross-sectional area was 0·02 m^2 ($n = 681$) for rill in several north- and south-facing watersheds. The total volume of rill erosion was 100 m^3 in the south-facing watershed (W960) and 40 m^3 in the north-facing watershed (W1165). The average sediment yield from rills in the 1996 was 13 $m^3\ ha^{-1}$ and when extrapolated to the Spring and Buffalo Creek watersheds it was 61 000 m^3.

No rill erosion measurements were made during 1997 and measured erosion rates for 1998, 1999 and 2000 are based on the three rill traps on a south-facing slope. From June 1998 to November 2000, rills widened slightly as the winter freeze–thaw process eroded sediment from the sides and deposited it on the bottoms where opportunistic plants like yellow evening-star (*Mentzelia speciosa*; Laurie Huckaby, 1999, pers. comm.) sprouted and helped trap the sediment. Rill erosion rates increased dramatically when rain intensity

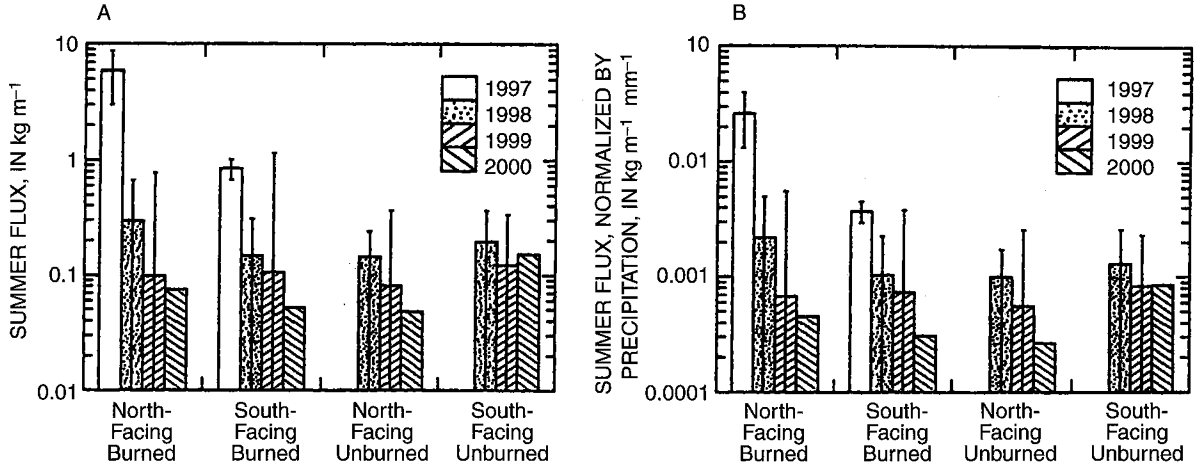

Figure 4. Sediment flux from hillslope traps in the Spring Creek watershed: (A) summer flux; (B) summer flux normalized by the amount of rainfall during the the summer (June, July, August and September). The top of each solid bar is equal to the mean of the four traps and the horizontal bars represent the 95 per cent confidence limits. Most of the sediment moves during the summer and, thus, these values approximate the annual flux. No traps were deployed in 1997 in the unburned area

exceeded about 30 mm h^{-1}. For example, the time-averaged sediment flux during the summer of 1998 was 0·0053 kg m^{-1} d^{-1} when rain intensities were less than 29 mm h^{-1} while during the summer of 1999, the average flux was 0·18 kg m^{-1} d^{-1} when the rain intensity was 35 mm h^{-1} during one storm (17 July 1999). The daily flux was 22 kg m^{-1} d^{-1} for this single storm that accounted for more than 90 per cent of the rill erosion in 1999. Our estimate of the average sediment yields to the stream channels during the summer is based on these time-averaged fluxes, channel length in the burned areas, and the rill density (3·6 per cent). Estimated yield to the stream channels by rill erosion was 310 m^3 in 1997 (where we assumed the large rainstorm on 31 August 1997 produced rill erosion of the same order of magnitude as the rainstorm on 17 July 1999), and measurements of average yield were 10 m^3, 310 m^3 and 10 m^3 in 1998, 1999 and 2000 respectively.

Sediment storage

Erosion dominated in the low-order watersheds with steep gradients, and deposition dominated in the higher-order watersheds with lower gradients. During the 1996 floods in the Buffalo Creek watershed, eroded sediment was deposited on existing alluvial fans at the mouths of tributaries, filled the adjacent main channel, and covered the floodplain near the mouth of the tributary with sediment ($D_{50} = 1{\cdot}9$ mm, $\sigma = 2{\cdot}4$ mm). In other places the floodplains were left exposed. In Spring Creek, the 1996 floods removed any pre-existing flood plain, incised some old alluvial fans, filled the entire main channel with eroded sand and gravel ($D_{50} = 4{\cdot}8$ mm, $\sigma = 3{\cdot}3$ mm), and deposited some sediment on existing fans or created new fans. Since 1996, steady flow has slowly evacuated sediment from the Buffalo Creek channel and transported it into the North Fork of the South Platte River, but sediment on the floodplains has been, for the most part, untouched and thus stored in the system. In Spring Creek, extensive post-fire and flood deposits of sediment have created an active braided channel that fills the entire valley floor. During long periods, steady flow in Spring Creek has created a channel with a floodplain at some locations; however, during short periods, flash floods have eroded the floodplain throughout the valley and deposited additional sediment from upstream creating a new braided surface (Moody, 2001). We estimate the floods of 1996 deposited an average depth of 0·5 m of sediment in the main channel of Spring Creek (Figure 5), and between 1997 and 1999 an additional 1·0 m of aggradation has decreased the gradient of the channel near the mouth.

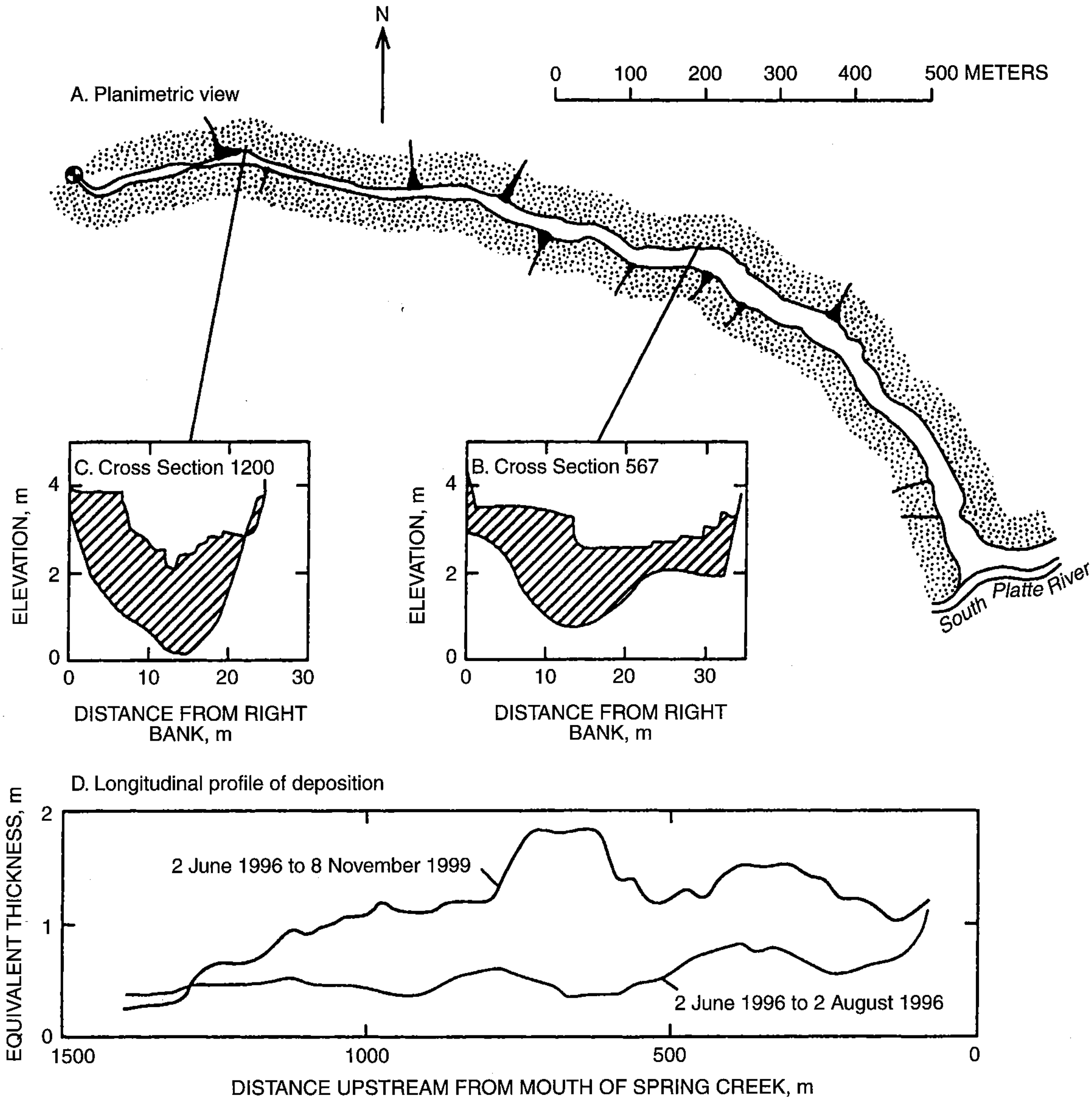

Figure 5. Location and thickness of wildfire-related sediment in the study area of Spring Creek. (A) Planimetric view with alluvial fans indicated by black triangles at the end of some tributaries. (B and C) Two of 147 cross-sections showing channel fill and overbank floodplain deposits. (D) Longitudinal profile of the equivalent thickness deposited during the first summer after the wildfire (1996) and also the net aggradation between 1996 and 1999. The equivalent thickness is equal to cross-sectional area of the deposit divided by the average channel width of the study reach (27 m) and smoothed with a running seven-point average

Erosion

Erosion of unchannelized and channelized drainages after the wildfire (1996) was greater than deposition in the two subwatersheds (W960 and W1165). The south-facing watershed (W960) had a net erosion of 1800 m^3 and the north-facing watershed (W1165) had net erosion of 470 m^3 of sediment. Sediment erosion, however, was not spread evenly among the channels within the watershed. While some of the first-order channels (side slopes were perpendicular to elevation contours) often resembled rills (side slopes were approximately parallel to elevation contours) in size, these first-order channels were created by water discharged from a series of converging rills occupying a hollow (Welter, 1995) at the head of the first-order channel. Only about

10 per cent of the eroded sediment came from first-order and 20 per cent came from second-order channels. The majority (70 per cent) of the eroded sediment came from third- and fourth-order channels, similar to observations made in the Snowy Mountains of Australia (Brown, 1972). The average equivalent sediment yield from channels in these two subwatersheds was 190 m^3 ha^{-1} which extrapolates to 890 000 m^3 for the Spring and Buffalo Creek watersheds.

Sediment transport out of the watersheds

Sediment transport rates increased after the wildfire. Pre-fire, total transport rates (bedload and suspended load) were available for Buffalo Creek (Figure 6) but not for Spring Creek. The sediment response in Buffalo Creek was a 20-fold increase at 1·0 m^3 s^{-1} and a 7·2-fold increase at 2·0 m^3 s^{-1}, similar to the 5·7-fold increase reported by Anderson (1974) in coastal Oregon forests. Without a pre-fire sediment rating-curve for Spring Creek, the absolute response is unknown. After the wild-fire, the total sediment transport rate in Spring Creek was about five to ten times the transport rate in Buffalo Creek. Before the wildfire, the bedload to suspended-load ratio for Buffalo Creek averaged 10 ± 8 (Williams and Rosgen, 1989). After the wildfire, the ratio was 6 ± 4 for Buffalo Creek and 14 ± 24 for Spring Creek indicating more fine material was available for transport.

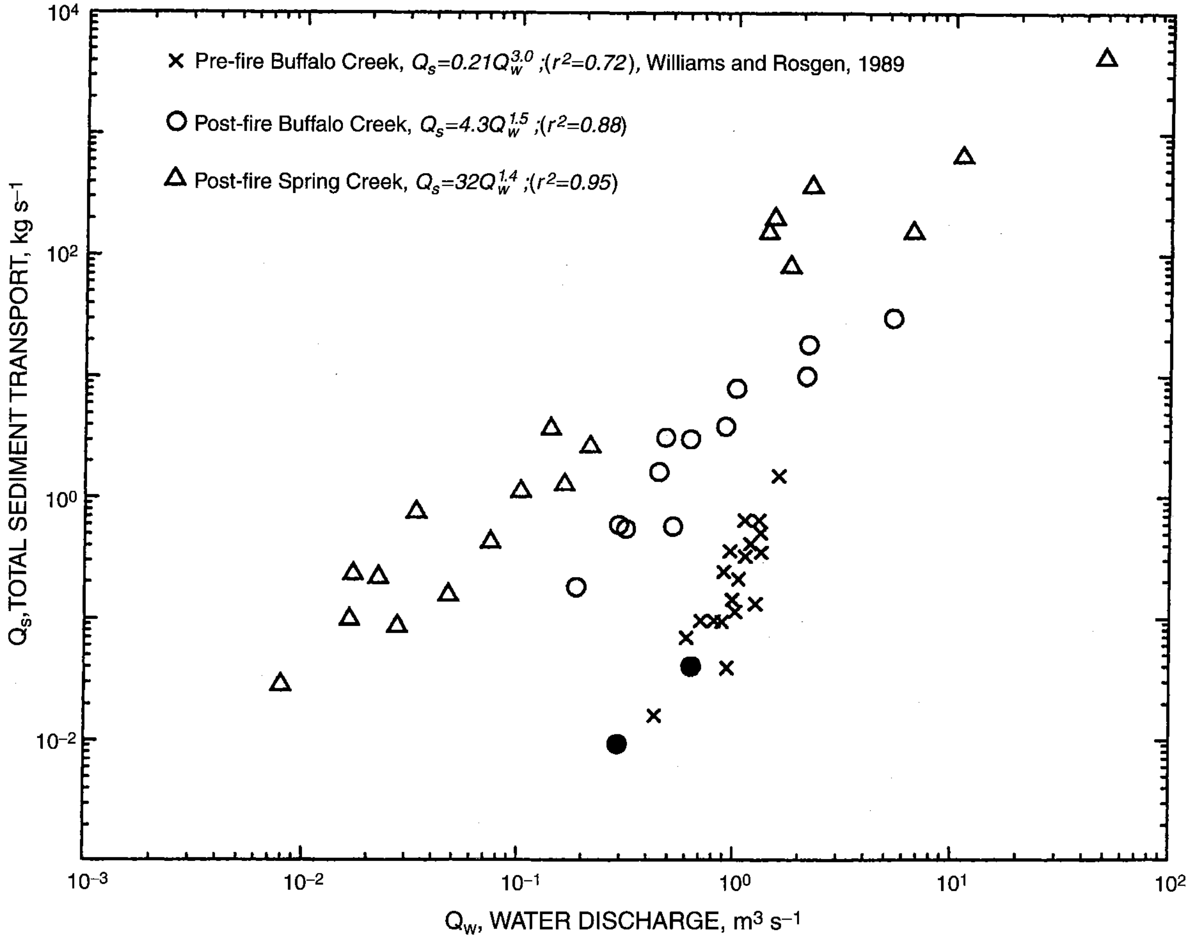

Figure 6. Sediment transport versus water discharge for Buffalo Creek and Spring Creek. The solid circles are measurements made in July 1998 and June 2000 when almost no sand or gravel was in Buffalo Creek and the streambed was cobbles and boulders similar to pre-fire conditions

Sediment transport was both an episodic and steady process; however, before the wildfire, sediment transport was probably supply-limited and after the wildfire it was transport-limited (Stallard, 1985; Moody, 2001). In Spring Creek, flash floods conditions accounted for 67 per cent of the sediment transported from the watershed and steady-flow conditions accounted for 33 per cent while in Buffalo Creek flash floods conditions accounted for 15 per cent and steady-flow conditions accounted for 85 per cent. The combined output for Spring Creek and Buffalo Creek is listed in Table IV. The outputs from the channels have shown a definite decrease and we estimate that the relaxation time is probably longer than four years, similar to the response in steep mountainous terrain in Arizona (about four years; Pase and Ingebo, 1965) and in Washington state (more than seven years; Helvey, 1980).

Sediment transport into the reservoir

Strontia Springs Reservoir trapped most of the coarse- and fine-grained sediment from the burned watersheds. The reservoir is relatively small (length 2700 m; storage volume $9{\cdot}5 \times 10^6$ m^3) with an 85 per cent trapping efficiency (Borland, 1978) that retains the coarsest fraction (sands and gravels) but passes some of the fine fraction depending upon the size of the flood. The initial floods in 1996 were so large that they transported some of the bedload and suspended-load sediment from the burned watersheds into the reservoir in a few hours or days. Part of the suspended load (silt and clay) was trapped in the reservoir but some passed through the reservoir during the 1996 flash flood and was trapped behind the Marston Diversion and Chatfield Dams farther downstream. The bedload, however, settled out and created a delta with an approximately 10-m slip face (Figure 7A, September 1996) in the upper end of Strontia Springs Reservoir. Deposition onto and movement of this delta depended upon reservoir operations and the occurrence of flash floods in the burned areas upstream (Figure 7).

The initial post-fire sediment deposits in Strontia Springs Reservoir that occurred during the summer of 1996 consisted of coarse-grained sediment in the delta and fine-grained sediment deposited further downstream.

Table IV. Coarse (sand and gravel) sediment transport by water year into Strontia Springs Reservoir based on change in sediment volume in the reservoir

Deposition period	Days	Volume (m^3)	Sediment transport rate: Volume (m^3 d^{-1})	Mass (kg s^{-1})	Unit width[a] (kg m^{-1} s^{-1})	Comments
1996						
18 May–13 September 1996	2	31 000	16 000	310	16	Initial input probably occurred in about 2 days.
13 September–2 October 1996	19	21 000	1100	22	1·1	Reservoir level was lowered during deposition period.
1997						
2 October 1996–27 June 1997	268	12 000	45	0·89	0·045	Winter
27 June–13 August 1997	47	21 000	450	8·9	0·45	–
13 August–12 September 1997	30	3100	100	2·0	0·10	Large flash flood occurred on 31 August.
1998						
12 September 1997–22 May 1998	252	41 000	160	3·1	0·16	Winter
22 May–15 July 1998	54	50 000	930	18	0·90	Water level was lowered during the spring.
15 July–3 August 1998	19	30 000	1600	31	1·6	Large flash flood occurred on 31 July.
3 August–23 October 1998	81	15 000	190	3·7	0·18	–
1999						
23 October 1998–4 June 1999	224	26 000	120	2·4	0·12	Winter

[a] Width of the South Platte River at the entrance to Strontia Springs Reservoir is about 20 m.

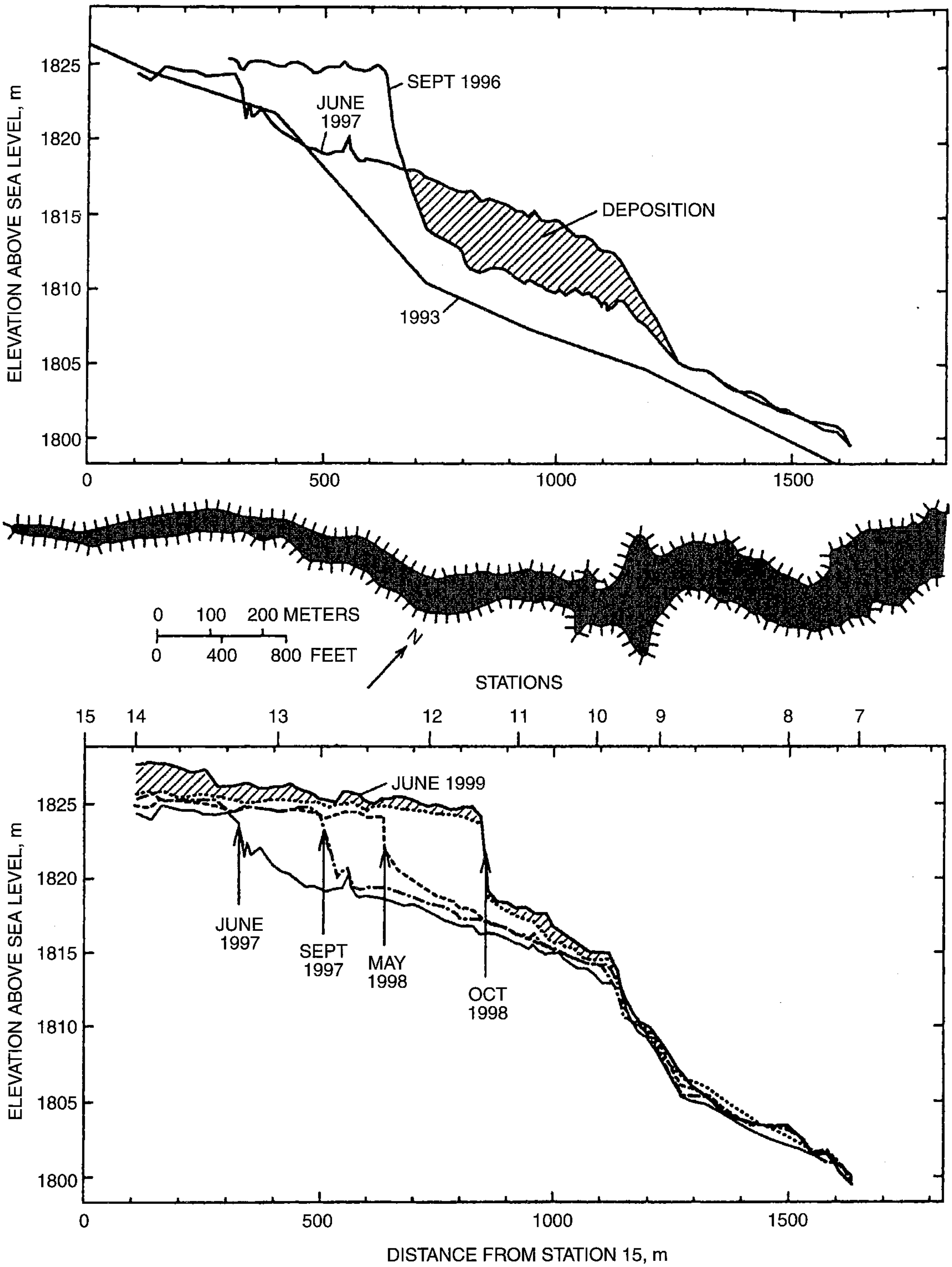

Figure 7. Location and thickness of the wildfire-related sediment in the upper end of Strontia Springs Reservoir. The numbers on the planimetric view in the middle of the figure correspond to the Denver Water Department's survey stations. (A) The pre-fire 1993 surface is shown for comparison with the post-fire September 1996 and June 1997 surfaces. (B) The location of the delta for five surveys in 1997, 1998 and 1999

Measurements of the delta indicated that 52 000 m^3 of fire-related, coarse-grained sediment was deposited on top of existing sediment. Based on a few sediment cores collected from the lower end of the reservoir during the winter of 1996–97, we estimated 0·5 m of fine-grained sediment (about 100 000 m^3) was deposited in the reservoir. Field measurements indicated an additional 2500 m^3 of mostly fine-grained sediment (12 per cent clay, 66 per cent silt, 21 per cent sand and 1 per cent gravel) was deposited downstream behind the Marston Diversion Dam. The total deposition during the summer of 1996 was therefore 154 000 m^3.

Sediment transport into the reservoir after 1996 reflected the complex response of reservoir operations and the transport and storage of sediment upstream. The pre-fire, bedload transport rate of the South Platte River (0·86 kg s^{-1}) was measured by Borland (1978) before the Strontia Springs Dam was built. Bedload transport rates into the reservoir after the wildfire ranged from 0·89 kg s^{-1} to 310 kg s^{-1} (Table V). This range of transport rates probably depended upon (1) sediment storage in the channel reach upstream from the reservoir, (2) operation of Strontia Springs Reservoir, and (3) reservoir operations upstream of the storage reach. For example, the transport rate after the flash flood on 31 August 1997 seems relatively low (2 kg s^{-1}). Sediment from this flood was probably stored in the channel reach upstream from the Strontia Springs Reservoir because discharge from upstream reservoirs was decreased at the end of summer, but transported (18 kg s^{-1}) during the following spring when the downstream reservoir was lowered and water was released from upstream

Table V. Sediment budget for watersheds burned by the Buffalo Creek Fire by water year

Process	Watershed			Reservoir Input (m^3 a^{-1})
	Input (m^3 a^{-1})	Storage (m^3)	Output (m^3 a^{-1})	
1996				
Hillslope interrill erosion (14 per cent)	150 000	150 000		
Hillslope rill erosion (6 per cent)	61 000	211 000		
Channel erosion (80 per cent)	890 000	1 101 000		
Transport out of watersheds		947 000	[a]154 000	
Transport into reservoir		947 000		154 000
1997				
Hillslope interrill erosion	1,300	948 000		
Hillslope rill erosion	310	949 000		
Transport out of watersheds		874 000	75 000	
Transport into reservoir		874 000		36 000
1998				
Hillslope interrill erosion	85	874 000		
Hillslope rill erosion	10	874 000		
Transport out of watersheds		790 000	84 000	
Transport into reservoir		790 000		136 000
1999				
Hillslope interrill erosion	40	790 000		
Hillslope rill erosion	310	790 000		
Transport out of watersheds		737 000	53 000	
Transport into reservoir		737 000		26 000
2000				
Hillslope interrill erosion	26	737 000		
Hillslope rill erosion	10	737 000		
Transport out of watersheds		734 000	3,100	
Transport into reservoir				not measured
Total	1 103 000	734 000 67 per cent	369 000 33 per cent	352 000 32 per cent

[a] Output is based on deposition in Strontia Springs Reservoir.

reservoirs. Thus, the response varied from no increase during winter months to about a 360-fold increase that lagged the response of the burned watersheds by up to six months.

DISCUSSION

Hydrologic response

The rainfall–runoff relation for Spring Creek watershed has an apparent change in slope at about 10 mm h^{-1} (Figure 3) which may be caused by relative storm size, threshold intensity, or both. It is possible that some of the discharge measurements made at the mouth of Spring Creek may represent the effect of rain storms smaller in size than the Spring Creek watershed and thus may have affected only a few subwatersheds with smaller drainage areas. The unit-area peak discharge calculated using the drainage area of the Spring Creek watershed would, therefore, be less than the actual unit-area peak discharge. This may have the greatest effect for low-intensity storms, if low intensities correspond to smaller-sized rain storms; unfortunately, rainfall statistics of this nature do not exist for summer convective rainstorms (N. Doesken, pers. comm., 2000). Another possible explanation is that rainfall intensities greater than 10 mm h^{-1} may exceed the average infiltration rate of the watershed or a threshold intensity such that runoff is dominated by sheet flow which produces flash floods. A similar threshold intensity was reported by Mackay and Cornish (1982) for watersheds on the Bega Batholith in New South Wales. In the Spring Creek watershed, most rainstorms in 1999 and 2000 with intensities greater than 10 mm h^{-1} produced unit-area peak discharges which were generally an order of magnitude less than those in 1997. This suggests that the threshold of critical intensity may be increasing and might explain the decrease in flood magnitude in 1999 and 2000 (Table II). This decrease in rainfall–runoff response after the wildfire is especially clear between 1997 and 2000. In 1997, a rain intensity of about 19 mm h^{-1} produced 0·31 m^3 s^{-1} km^{-2} while in 2000 a similar intensity produced only 0·0031 m^3 s^{-1} km^{-2}. Also in 1997, an intensity of about 50 mm h^{-1} produced 6·6 m^3 s^{-1} km^{-1} but in 2000 a similar intensity produced only 0·11 m^3 s^{-1} km^{-2}. This indicates a possible relaxation time for runoff of about three to four years, similar to other studies that measure relaxation times of two to seven years for peak discharges (Brown, 1972; Rowe *et al.*, 1954).

Geomorphic response

Sediment budget. The sediment budget for these burned watersheds changed with time during the first few years after the wildfire and may have reached a steady-state condition in about four years. Erosional inputs to the sediment budget were hillslope (interrill and rill) and channel erosion, which produced an estimated 1 101 000 m^3 of sediment during the first summer after the wildfire (Table V). Most of the eroded sediment came from low-order watersheds and was stored as floodplain deposits, alluvial fans and channel fill in higher-order watersheds. The initial transport rates out of the watersheds were relatively high (154 000 m^3 a^{-1}) and the predicted residence time for sediment stored in the watersheds was only seven years. After the first summer (1996) the inputs from the hillslopes were essentially negligible contributions to the budget (Table IV). The residence time by the end of the 2000 water year had increased to 240 years because the transport rate decreased to 3100 m^3 a^{-1}. If we assume in the future (1) transport rates will decrease to pre-fire rates typical of Buffalo Creek (about 0·1 kg s^{-1}, Figure 6) and (2) active transport does not occur during the winter months (*c.* 240 days), then the combined output from Spring and Buffalo Creeks would be about 2500 m^3 a^{-1}. The equivalent residence time is about 300 years. In this sediment budget for the initial transient period after a wildfire, erosion and transport rates decrease relatively quickly with time and corresponding residence times of stored material increase with time to the pre-fire or steady-state condition.

Depositional features. Fire-related deposits on alluvial fans at the mouths of tributaries and as channel deposits in the Buffalo and Spring Creek watersheds do not have the obvious fire signature of charcoal. The initial floods in 1996 and 1997 may have been of such magnitude that this evidence was swept downstream out of the watersheds (Gonzalez and Hunt, 1999). However, these fire-related deposits do seem to have significant amounts of ash or fine, burned particulate matter because when these sediments are reworked by streams, the water becomes blacker, indicating that possibly some fire-related material in the deposits is being dissolved or suspended by the water. These deposits still persist in some places along the main channels in

Spring Creek (Figure 5) and Buffalo Creek and probably exist in many other locations from other wildfires as well but are not easily identified as fire-related sediments (Wohl and Pearthree, 1991; Meyer *et al.*, 1995; Gonzalez and Hunt, 1999).

Sediment deposits easily recognized as related to fire (containing charcoal and partially burned debris) were deposited in Strontia Springs Reservoir. The stratified character of this wildfire-related sedimentation is similar to very fine sedimentation in lakes (Swain, 1978) or in the ocean (Mensing *et al.*, 1999). Much of the large, burned organic debris floated on the surface of the reservoir and was not incorporated into the sediment because it was removed by the Denver Water Department. If the dam had not trapped the sediment and burned woody debris, it probably would have been transported through Waterton Canyon and potentially deposited on an alluvial fan at the mouth of the canyon with characteristics similar to the alluvial fans extending from canyon mouths along the San Gabriel Mountains (Doehring, 1968) and to alluvial fans described by Meyer and Wells (1997) after the Yellowstone fire. Charcoal layers within upward-finding alluvial sediments were exposed by the incision of a smaller alluvial fan within the Buffalo Creek fire by the 1996 floods. These layers have radiocarbon dates indicating perhaps three previous fire–flood sequences spanning 2900 years BP (Elliott and Parker, 1999). Fire-related depositional features in valley-fill deposits on canyon floors in western Colorado had radiocarbon dates indicating three fire-related events corresponding to about 1900, 5600 and 10 300 years BP (Scott *et al.*, 1999, 2000). Additional evidence for the persistence of fire-related depositional features was found by Meyer *et al.* (1995) in Yellowstone National Park where fire-related alluvial fans contributed to the development of fluvial terraces during the last 6000 years, and also by Wohl and Pearthree (1991) in the Huachuca Mountains of southeastern Arizona. The steady-state residence time of stored sediment resulting from the Buffalo Creek fire is about 300 years, which is greater than the recurrence interval for wildfires (20 to 50 years) in the Colorado Front Range, so that these depositional features may also become persistent landscape features in this area.

Erosional features. Our data suggest north-facing hillslopes erode faster than south-facing hillslopes and may, over geological time, have an asymmetrical effect on landscape evolution (Melton, 1960; Dohrenwend, 1978; Branson and Shown, 1989; Wende, 1995). Differences in the thickness of the granite residuum on north- (6–30 m) and south-facing (3–4 m) slopes described by Blair (1976) in this area support asymmetrical erosion. In this semi-arid mountainous environment, the morphology of the rills changed little in four years in contrast to rills studied by Schumm (1956) in a humid environment where the rills formed and disappeared annually. Rill erosion appears to also be asymmetrical based on some field observations and measurements, aerial photographs, and some preliminary mapping. It may be possible in a semi-arid environment that some rills resulting from wildfires eventually evolve into gullies and then into major landscape drainages, thus increasing the channel density of south-facing relative to north-facing hillslopes.

Unchannelized drainages were incised by runoff in many low-order watersheds. The eroded sediment may represent colluvium that refilled channels after the last extreme erosional event, similar to the situation observed by Rice (1974) in the San Gabriel Mountains. Studies in nearby areas of the Colorado Front Range indicate the basal age of colluvium in similar unchannelized drainages or hollows is from 1550 to 13 500 BP (Welter, 1995). We have already observed this refilling by several processes: (1) dry ravel has begun to fill all channels; (2) relatively low rainfall has deposited small alluvial fans in lower-order channels but has not exported sediment out of the subwatershed; and (3) the freeze–thaw process on hillslopes during the winter months has produced a flux of sediment into channels on warm sunny days (south-facing flux was about $0{\cdot}9$ kg m^{-1} d^{-1}; north-facing flux was about $0{\cdot}09$ kg m^{-1} d^{-1}). This refilling time may be much greater than the wildfire recurrence interval and, thus, these erosional features may persist in some state of refilling when the next fire–flood sequence occurs.

CONCLUSIONS

A rainfall–runoff relation was developed for watersheds burned by a wildfire in a region where the wildfire recurrence interval is 20 to 50 years. This region of the Colorado Front Range is dominated by episodic, short-duration, high-intensity rainfall and the maximum unit-area peak discharge from burned watersheds was 24 m^3 s^{-1} km^{-2} for an I_{30} rain intensity of about 90 mm h^{-1}. This maximum discharge probably occurs only if rainstorms are about the same size as the burned watershed. By 2000, the runoff response to similar rain intensities had decreased significantly and suggests a minimum relaxation time of three to four years.

Erosional and depositional landforms were formed after this wildfire in steep, semi-arid mountainous terrain. Measurements indicated rill erosion accounted for 6 per cent, interrill erosion for 14 per cent, and drainage erosion for 80 per cent of the initial erosion which occurred during the first two summers after the wildfire. The erosional response on hillslopes ranged from 150- to 240-fold increase in yields and a relaxation time of three to four years. This erosion produced rills and incised drainages in low-order watersheds and sediment deposits in higher-order watersheds. Sediment transport rates in channels seemed to have a relaxation time (about four to five years) slightly greater than erosion rates on the hillslopes.

Relaxation times for this study appear to be short relative to wildfire recurrence intervals. This agrees with Swanson's (1981) model that explicitly assumes the relaxation or recovery time is much less than the fire recurrence interval and, thus, suggests fire-related geomorphic evidence may disappear before the next fire disturbance. However, the relatively short relaxation times result in relatively long residence times (*c.* 300 years) and refilling times (possible 1000 to 10 000 years) of depositional and erosional features. These geomorphic features may, therefore, persist and become initial conditions affecting subsequent sediment transport in response to the next wildfire disturbance. In this manner, wildfire in Colorado and perhaps in other areas of the western United States may provide an important link between short time-interval and small spatial-scale processes and longer time intervals and larger spatial scales of landscape evolution.

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