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Alpine debris flows triggered by a 28 July 1999 thunderstorm in the central Front Range, Colorado

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Abstract

On 28 July 1999, about 480 alpine debris flows were triggered by an afternoon thunderstorm along the Continental Divide in Clear Creek and Summit counties in the central Front Range of Colorado. The thunderstorm produced about 43 mm of rain in 4 h, 35 mm of which fell in the first 2 h. Several debris flows triggered by the storm impacted Interstate Highway 70, U.S. Highway 6, and the Arapahoe Basin ski area. We mapped the debris flows from color aerial photography and inspected many of them in the field. Three processes initiated debris flows. The first process initiated 11% of the debris flows and involved the mobilization of shallow landslides in thick, often well vegetated, colluvium. The second process, which was responsible for 79% of the flows, was the transport of material eroded from steep unvegetated hillslopes via a system of coalescing rills. The third, which has been termed the "firehose effect," initiated 10% of the debris flows and occurred where overland flow became concentrated in steep bedrock channels and scoured debris from talus deposits and the heads of debris fans. These three processes initiated high on steep hillsides (>30°) in catchments with small contributing areas (<8000 m^2), however, shallow landslides occurred on slopes that were significantly less steep than either overland flow process. Based on field observations and examination of soils mapping of the northern part of the study area, we identified a relation between the degree of soil development and the process type that generated debris flows. In general, areas with greater soil development were less likely to generate runoff and therefore less likely to generate debris flows by the firehose effect or by rilling. The character of the surficial cover and the spatially variable hydrologic response to intense rainfall, rather than a threshold of contributing area and topographic slope, appears to control the initiation process in the high alpine of the Front Range. Because debris flows initiated by rilling and the firehose effect tend to increase in volume as they travel downslope, these debris flows are potentially more hazardous than those initiated by shallow landslides, which tend to deposit material along their paths.

Keywords: Colorado; Front Range; Alpine environment; Debris flow; Rainfall; Topography; Alpine soils; North American Monsoon

1. Background and introduction

Sediment transport rates above timberline in the Front Range of Colorado are relatively low compared to other mountain environments and are dominated by rockfall, talus, and debris flow processes (Caine, 1986; Menounos, 1996). Debris flows, rapid gravity-driven mass flows of grains and intergrannular fluid (Varnes, 1978; Iverson and Vallance, 2001), are an important sediment transport process and natural hazard in high alpine environments (e.g. Caine, 1976; Costa and Jarrett, 1981; Rapp and Nyberg, 1981; Rickenmann and Zimmermann, 1993; Becht et al., 2005). In contrast to more temperate settings

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where debris flows are typically generated from landslide failure of the soil mantle (Campbell, 1975; Iverson et al., 1997), debris flows in the alpine are frequently initiated by the erosion and entrainment of hillslope and channel material by overland flow (e.g. Fryxell and Horberg, 1943; Curry, 1966; Berti et al., 1999; Hürlimann et al., 2003). Sediment entrainment by overland flow has been shown to be the dominant mechanism for debris flow generation in steep terrain burned by wildfire (Wohl and Pearthree, 1991; Meyer and Wells, 1997; Cannon et al., 2001a), and in poorly vegetated arid (David-Novak et al., 2004) and semi-arid landscapes (Griffiths et al., 2004).

Theoretical understanding of sediment transport under steady rainfall conditions (Dietrich et al., 1992; Montgomery and Foufoula-Georgiou, 1993) and empirical evidence from soil-mantled landscapes in temperate humid climates (e.g. Montgomery and Dietrich, 1994a; Vandaele et al., 1996) indicate that topographic form controls the spatial location of channel initiation. These findings have been used to partition the landscape into dominant channel initiation (saturation excess overland flow or shallow landslide) processes using a threshold relation between topographic slope angle and contributing area (Dietrich et al., 1992; Montgomery and Dietrich, 1994b). Similar topographic thresholds have been used in areas burned by wildfire to identify the initiation of debris flow transport (Cannon et al., 2001b). However, because debris flow generation by runoff processes is in part dependent on the availability of adequate material for transport (Bovis and Jakob, 1999), the topographic thresholds tend to vary with spatial variation in rock type and weathering intensity (Cannon et al., 2003).

The purpose of this case study is to document the debris flow activity associated with a 28 July 1999 thunderstorm in a 240-km^2 area roughly centered on the Continental Divide (Fig. 1). Overland flow and landslide processes triggered about 480 debris flows that were concentrated on the flanks of Mount Parnassus, on the western flanks of Lenawee Mountain and Grizzly Peak near the Arapahoe Basin Ski area, and the northern flank of Torreys Peak (Fig. 1). Field examination of debris flows and comparison of initiation locations with digital topography and soils mapping of the northern part of the area indicates that most runoff debris flows were initiated on steep slopes where bedrock is shallow or exposed at the ground surface. Landslide debris flows were initiated on significantly less steep slopes in thick colluvium often with well-developed soils and vegetal cover. A detailed map showing the debris flow source areas and deposits was published previously and is available online (http://pubs.usgs.gov/of/2003/ofr-03-050/, Godt and Coe, 2003).

The debris flows did not cause any fatalities or injuries, but Interstate Highway 70 (I-70) and U.S. Highway 6 (U.S. 6) were blocked for about 25 h disrupting traffic on the major east–west routes through the mountains. Impacts to property were largely confined to the Arapahoe Basin ski area where about US$200,000 in damage to facilities was reported (Henceroth, 2000; Coe et al., 2002). We begin this paper with a description of the physiographic and geologic setting of the study area and follow with an analysis of the triggering rainfall. We then discuss our observations of debris flow initiation processes and their correlation with surficial cover type and present grain-size analyses of colluvium in debris flow source areas and of debris flow deposits. We conclude with results from an analysis of the topography of debris flow source areas and evaluate these source areas in the context of previously proposed topographic thresholds for channel initiation. Our hope is that the baseline information presented in this paper can be used in future regional modeling efforts to assess debris flow hazard.

1.1. Physiographic and geologic setting

The study area is located about 50 km west of Denver, CO in the central Front Range (Fig. 1) and is underlain predominantly by Precambrian biotitic gneiss and quartz monzonite with scattered Tertiary intrusions (Lovering, 1935; Bryant et al., 1981), which have undergone varying degrees of hydrothermal alteration and mineralization (Tweto and Sims, 1963). Much of the area is in the alpine zone, defined here as the area above timberline (~3500 m elevation). Vegetative cover is primarily coniferous forest below timberline. Above timberline, hillside surfaces either are covered with alpine tundra or are unvegetated bedrock and colluvium. No glaciers or permanent snowfields are located in the study area; however, the highest north-facing slopes may hold snow well into the late summer. Elevations in the study area range from about 2900 m in the valley of Clear Creek to nearly 4350 m at the summit of Grays Peak (Fig. 1). The terrain is characterized by glaciated valleys and unglaciated interfluves (Barsch and Caine, 1984; Madole et al., 1998). The unglaciated interfluves are generally either bedrock arêtes or low-relief ridges capped by a bouldery diamicton that may be remnants of Tertiary-age alluvium and debris flow deposits (Madole, 1982). The area was glaciated several times during the Pleistocene, but only three ages of glacial deposits can be distinguished. Ice of the most recent glaciation (Pinedale age) descended to an elevation of about 2500 m in the Clear Creek valley. Pinedale ice is estimated to have disappeared between 14,000 and 12,000 ^{14}C years BP

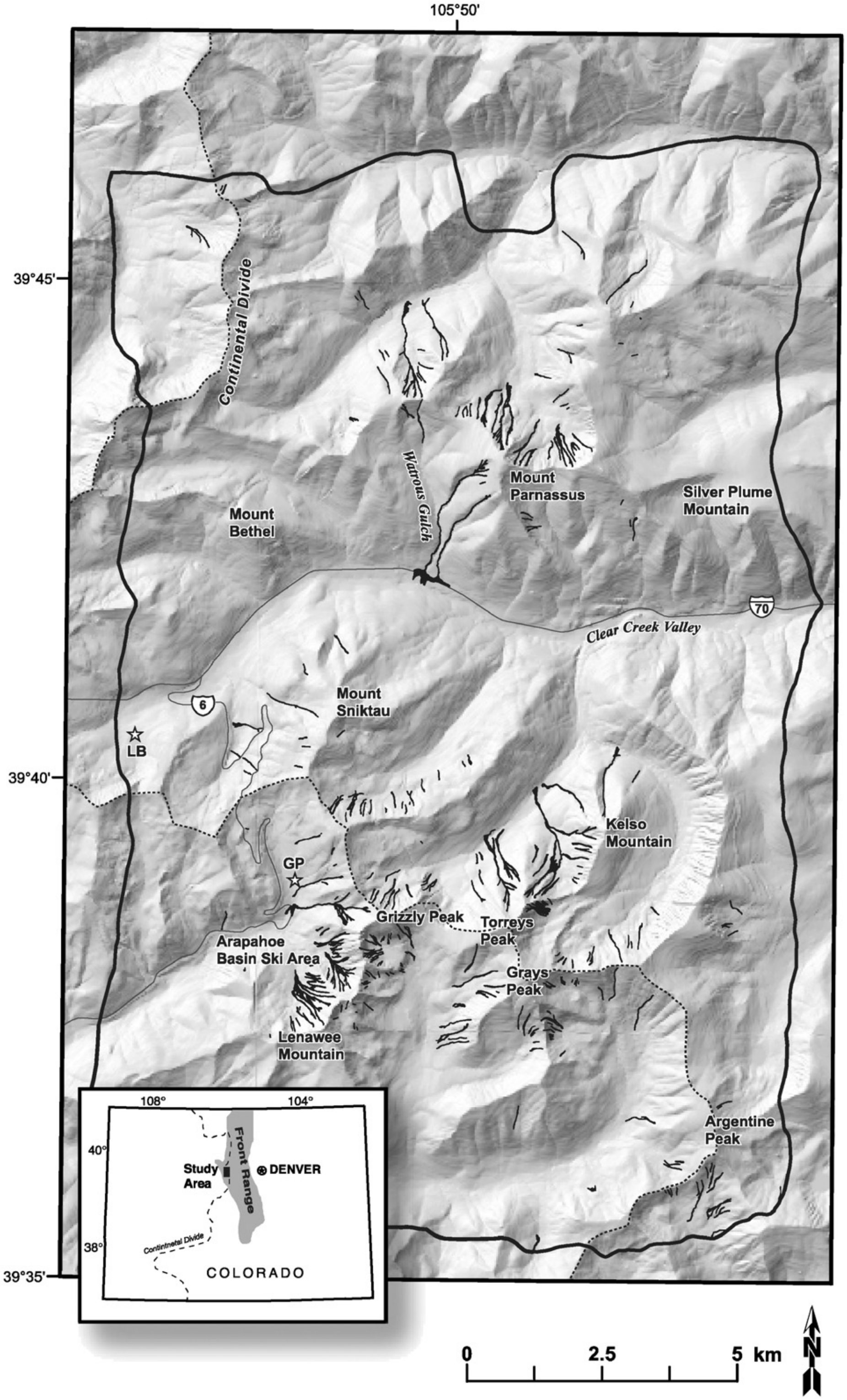

Fig. 1. Map showing the study area in the central Front Range of Colorado, USA and the distribution of debris flows. The Continental Divide is shown as a dashed line. Limit of aerial photography is shown as a heavy black line. Interstate Highway 70 (I-70) follows the Clear Creek Valley. SNOTEL stations are shown as stars and labeled LB for Loveland Basin and GP for Grizzly Peak.

(Madole et al., 1998). Most drainages in the study area contain late Pleistocene glacial deposits (Madole et al., 1998). Large Quaternary landslide deposits have been identified on the highest summits of Mount Parnassus and Mount Sniktau (Bryant et al., 1981).

At the heads of glaciated valleys are cirques — bowl-shaped recesses that were eroded by glacial activity (Madole et al., 1998). Steep ephemeral channels drain the head and sidewalls of the cirques. Fractured bedrock in the cirque head and sidewalls provides abundant material for the development of large talus deposits at the foot of steep slopes. The talus deposits are comprised of coarse angular rock fragments and bouldery debris derived primarily from rockfall (Caine, 1986; Madole et al., 1998). Debris flows and fluvial processes modify the talus deposits and transport material downslope thereby forming debris fans (White, 1981; Caine, 1986; Madole et al., 1998). The debris fans are composed of angular clasts and finer-grained material and often support some vegetation, unlike the bare talus deposits. Downslope from the debris fans are relatively flat-floored stream valleys filled by glacial and alluvial deposits. The flat

Table 1
Names and brief description of the generalized soil units in the northern part of the study area (from Retzer, 1962) correlated with dominant debris flow initiation process

Soil name	Brief soil description	Debris flow initiation process(es)
Vasquez loam — V	Soil formed in depressions and basins above timberline, usually at lower elevations than Ptarmigan soils. Contains fine sandy loam to rock materials that have fallen or washed into basins from adjacent Ptarmigan soils, alpine land, and rock areas. Soil is moist to wet most of the year; vegetation is commonly a dense growth of willows, sedges, and grasses. Less than 10% to >50% of surface is bare of vegetation.	None, but commonly an accumulation zone for debris from debris flows that initiate on steeper slopes at higher elevations
Ptarmigan loam — P	Soil formed on steeply sloping, well-drained areas above treeline, surface soil is 13 to 18 cm thick, dark gray or black with a high organic content and thickly matted with roots. Subsoil is a gravelly loam that is yellowish brown to dark-yellowish brown and loose, with relatively few roots compared to the surface soil. The portion of silt and clay is highest in the surface soil and decreases with depth. Overall, the soil is moderately to highly permeable. Less than 10% to >50% of surface is bare of vegetation.	Shallow landsliding, some rilling
Alpine rimland — A	Steep slopes and exposed, elongated areas parallel to ridgetops with loose, sandy, gravelly, and rocky materials, little to no organic soil, none to very little vegetation, commonly catchment areas for windblown snow from ridgetops in cirque basins, surface erosion is common.	Rilling, some shallow landsliding
Rock talus — Rb	Accumulations of loose rock ranging from immense single rocks to coarse gravel that take the form of fans, aprons, or rock glaciers; runoff from heavy rainfall or snowmelt is rare; seeps and springs common near base of accumulation; vegetation can grow up through rocks when accumulation is thin.	Source of debris mobilized by firehose from higher elevations, some shallow landslides and rilling
Rock outcrop — Ra	Great masses of exposed rock that form mountain peaks and walls of cirques (includes some individual rocks of various sizes); melting snow and summer rain quickly runoff outcrops; common source areas for snow avalanches.	Catchment area for firehose, some rilling

Soil units have been generalized from the 13 Retzer (1962) units as follows. V = Vasquez loam, Vasquez loam slightly bare, moderately bare, largely bare. P = Ptarmigan loam, Ptarmigan loam slightly bare, moderately bare, largely bare, outcrop. A = Alpine rimland and alpine wind-eroded land. Rb = rock slides. Ra = rock outcrop.

valley floors effectively limit the export of coarse-grained material from the high alpine hillslope system to the main stem drainages (Barsch and Caine, 1984; Caine, 1986).

Surficial deposits and soils in the area are derived from, or formed on, the predominantly metamorphic bedrock or glacial deposits and range from exposed

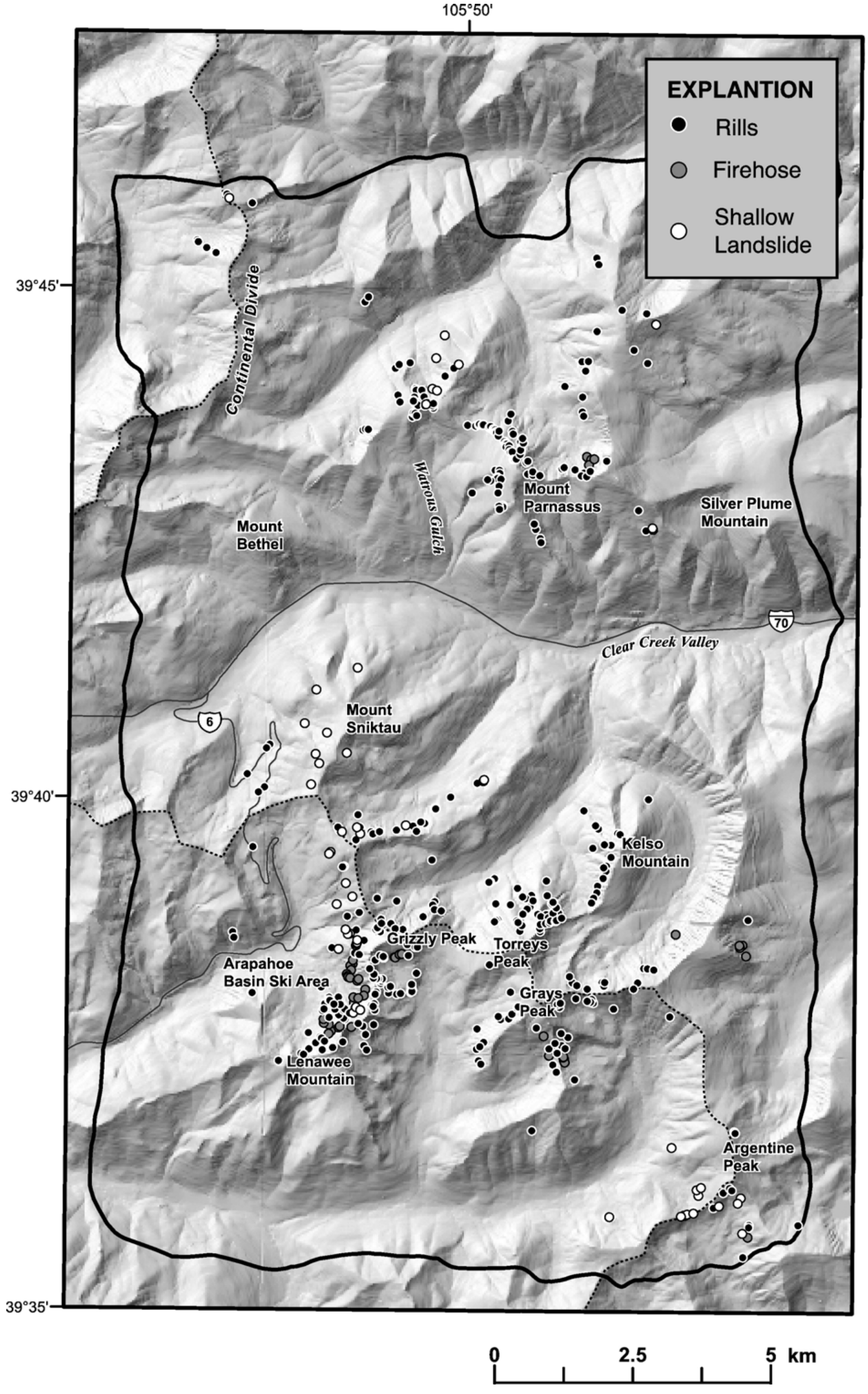

Fig. 2. Map showing debris flow initiation points and initiation process type.

bedrock to well-developed, organic-rich soils. Retzer (1962) described the northern third of the study area as steep, windswept rocky grassland and mapped 13 soil units within the area. Surficial materials were found to vary primarily because of the microclimate effects of aspect and exposure on vegetation growth and soil development rather than from differences in parent rock (Retzer, 1962). Nine of the 13 map units are classes of two organic soils: Ptarmigan and Vasquez. Ptarmigan and Vasquez soils are also known as Cryepts, cold soils with A/Bw/C profiles (Birkland et al., 2003). The classes were defined based on the percentage of soil surface that lacked vegetation. Additional units include rock outcrop, alpine rimland, alpine wind-eroded land, and rock talus (Retzer, 1962). Alpine rimland and wind-eroded land both consist of nonvegetated areas of loose debris on steep slopes that are commonly adjacent to ridgelines. Table 1 lists 5 generalized soil units based on Retzer's (1962) mapping. Soil units were generalized by combining the vegetation classes of the two organic soils, and alpine rimland with alpine wind-eroded land.

2. Methods and approach

Debris flow source areas, flow paths, and deposits were mapped from 2 sets of large-scale color aerial photography taken on 26 October 1999 and 11 September 2000 (Godt and Coe, 2003). The scale of the photography allowed us to accurately identify debris flow features as small as about 0.5 m across. Debris flows were mapped if their features were fresh, that is, they appeared to have a lighter tone than their surroundings and were not vegetated. Most mapped debris flows were field checked to verify that they were, in fact, 1999 features. In many parts of the study area, we were able to verify the timing of mapped debris flows because they affected roads or ski area infrastructure. In several cases, eyewitness accounts were available. As part of the mapping and field checking process, debris flows were classified according to the process that contributed the majority of material to each debris flow (Fig. 2). We identified two processes (the firehose effect and rilling) in which most of the material appeared to be initially transported by running water, and a third in which the material was mobilized from shallow landslides. Each of these processes is described in detail later in the paper.

Samples for particle-size analysis were collected from debris flow deposits and from the colluvium in debris flow source areas and analyzed using sieve and hydrometer methods conforming to ASTM D422 standards (American Society for Testing and Materials, 2002). Results from the particle-size analysis are reported using the ASTM D422 scale and classification in which gravel-size particles are between 75 and 4.76 mm in diameter, sand-size particles are between 4.76 and 0.074 mm in diameter, and particles smaller than 0.074 mm are classified as mud (silt and clay sizes).

Slope angle, contributing area upslope of debris flow source areas, and other measures derived from elevation data, are often used to characterize the topographic from of hillslope locations that are susceptible to debris flow initiation (Ellen et al., 1993; Montgomery and Dietrich, 1994b; Cannon et al., 2001b; Bathurst et al., 2003). In this study, debris flow initiation points were mapped as the farthest upslope position where fresh erosion or incision was visible in the aerial photographs. In some cases, a single mapped debris flow has multiple initiation points, all of which may not be mapped. This is particularly true for debris flows initiated by rilling because individual rills are typically small (<1 m wide). Thus, the number of initiation points is partially a function of the scale of the mapping rather than an exact count of every individual failure location. In addition, as rills form they may grow retrogressively in an upslope direction. Since any evidence of retrogression is obliterated during the formation of the rill, we are forced to assume the rills initiated at the highest location where erosion is visible.

Slope angle and upslope contributing area for the debris flow initiation points were compiled mainly from USGS 10-m digital elevation models (DEMs), except for a small area (~ 30 km^2) in the southeastern part of the study area where only 30-m data are available. We resampled these 30-m data to 10-m spacing to conform to data for the rest of the study area. Slope angles were computed using the SLOPE commands in the GRID module of ArcInfo[1]. Upslope contributing areas were computed using the D-infinity method of Tarboton (1997). The horizontal travel distance for each debris flow was calculated in ArcInfo by manually digitizing a curvilinear line from the debris flow initiation point along the debris flow path to the farthest extent of the mapped deposit.

3. Rainfall

A vigorous North American Monsoon (Adams and Comrie, 1997) during the summer of 1999 generated anomalously high rainfall that caused widespread flooding, landslides, and debris flows in Colorado (Soule,

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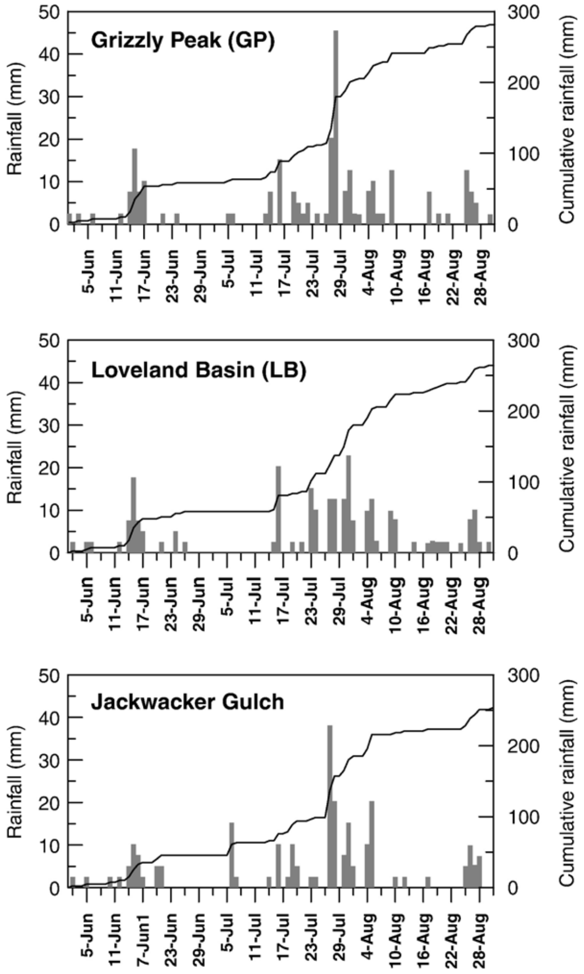

Fig. 3. Daily and cumulative rainfall for the summer season of 1999 recorded at two SNOTEL stations in the study area (Grizzly Peak and Loveland Basin) and the Jackwacker Gulch station just south of the study area. Debris flows occurred on 28 July 1999. See Fig. 1 for station locations.

1999; Avery et al., 2002; Coe et al., 2002; Coe and Godt, 2003; Godt and Coe, 2003; Godt and Savage, 2003). The North American Monsoon generally begins to affect the weather in the Front Range in early July with a shift of lower and middle tropospheric flow to a more northerly direction, thus causing tropical moisture to flow north from the eastern Pacific Ocean and the Gulf of California (Adams and Comrie, 1997). Two SNOTEL (SNOwpack TELemetry, U.S. Dept. of Agriculture) meteorological stations (Crook, 1977) are located within the study area (Grizzly Peak and Loveland Basin) and a third (Jackwacker Gulch) is located about 2.5 km south of the southern boundary of the study area (Fig. 1). Rainfall amounts are collected hourly at Grizzly Peak (3386 m), and this station therefore provides the best ground-based record of the storm. The Loveland Basin (3217 m) and Jackwacker Gulch (3341 m) stations only record daily rainfall.

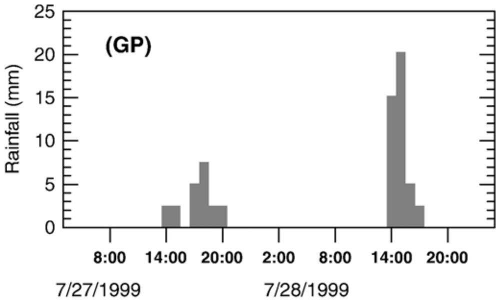

Fig. 4. Hourly rainfall for the Grizzly Peak SNOTEL station. Debris flows occurred on 28 July 1999. Times shown are Mountain Standard Time (MST).

3.1. Summer season, 1999

Prior to July 28, the summer of 1999 was marked by a moderate amount of rainfall within the study area (Fig. 3). A short, wet period in the middle of June was followed by a 3-week dry period of little or no precipitation. Beginning about July 13, rainfall was recorded on most days at the three stations, thus marking a change to the monsoon weather pattern. Precipitation totals for the period of July 13 to July 27 were 66 mm, 74 mm, and 71 mm at Loveland Basin, Jackwacker Gulch, and Grizzly Peak stations, respectively. On July 27, the day before the debris flows occurred, 13 mm of rainfall was recorded at Loveland Basin, 38 mm at Jackwacker Gulch, and 20 mm at Grizzly Peak. The 38 mm of rain at Jackwacker Gulch

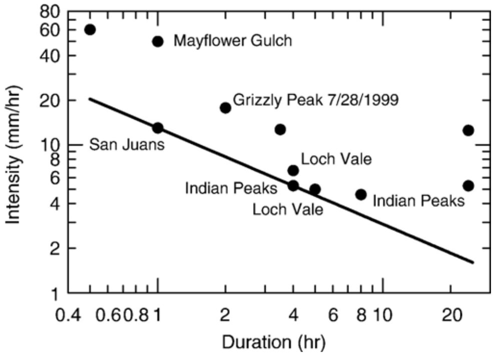

Fig. 5. Proposed rainfall intensity–duration threshold for the alpine zone of Colorado modified from Menounos (1996). Loch Vale, Mayflower Gulch, and Indian Peaks represent rainfall that triggered debris flows in central Colorado. Unlabeled points and San Juans are from Caine (1980). The threshold is defined by the equation $I=13.0D^{-0.65}$, where I is rainfall intensity in mm/h and D is rainfall duration in hours.

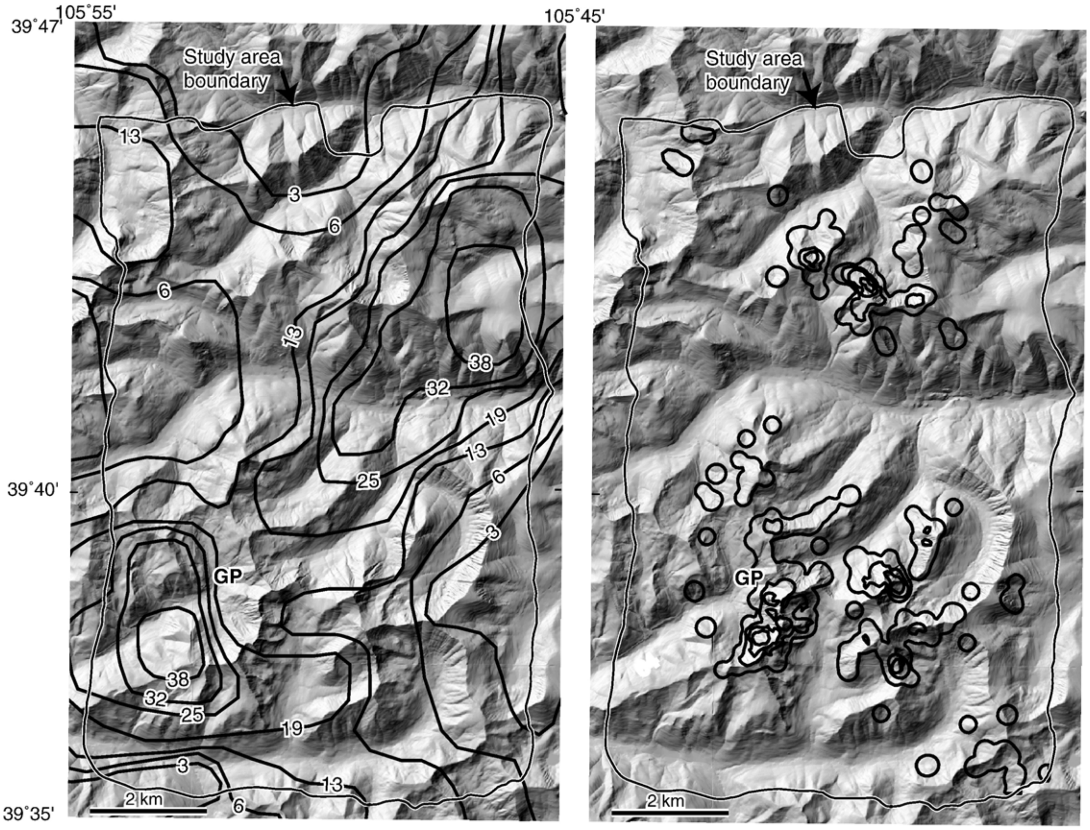

Fig. 6. Shaded relief map showing contours of rainfall (in mm) between 14:00 and 15:00 MST on 28 July 1999 estimated from NEXRAD data (left) and debris flow concentration contours (right) in the study area. Debris flow contours of 1, 5, 10, 15, 20, and 25 debris flows/0.25 km^2 were created using debris flow initiation points and an automated moving count circle method (Savage et al., 2001). GP is the location of the Grizzly Peak SNOTEL station.

may indicate that debris flows in the southernmost part of the study area were triggered on July 27 rather than July 28, but this cannot be confirmed because eyewitness accounts are not available. Melting snow was not a factor in triggering the debris flows. The SNOTEL stations indicate that there was no snow on the ground at any of the stations by June 21. Many of the slopes that produced debris flows, such as those in the Arapahoe Basin ski area, were completely free of snow and only a few very small patches of snow on the highest slopes were observed in the field in the days immediately following the occurrence of the debris flows.

3.2. 28 July 1999

Data from 28 July 1999 show that 43 mm of rain fell at the Grizzly Peak station in the 4-h period beginning at 13:00 Mountain Standard Time (MST) (Fig. 4). The first 2 h of the storm were the most intense with 15 mm recorded by 14:00 and another 20 mm by 15:00 MST. However, only 13 mm was recorded at the Loveland Basin station near the western edge of the study area, and 20 mm at the Jackwacker Gulch station south of the study area on this same day. These observations indicate that the storm covered much of the study area, but that total amounts, and presumably rainfall intensity, varied spatially with a maximum near Grizzly Peak. Work by Payton and Bredecke (1985) suggests that rainfall of the duration and intensity recorded at the Grizzly Peak SNOTEL station has a return period of about 100 years compared to other stations in the Front Range at similar elevations. This rainfall intensity lies well above a debris flow threshold proposed (Fig. 5) for the Front Range (Menounos, 1996).

Estimates of the spatial extent of intense rainfall are available from the NEXt Generation RADar (NEXRAD) station KFTG located near Denver, CO (see David-Novak et al., 2004 for a discussion of radar rainfall estimates). These data from the afternoon of July 28 show that the storm developed near the Continental Divide in Summit

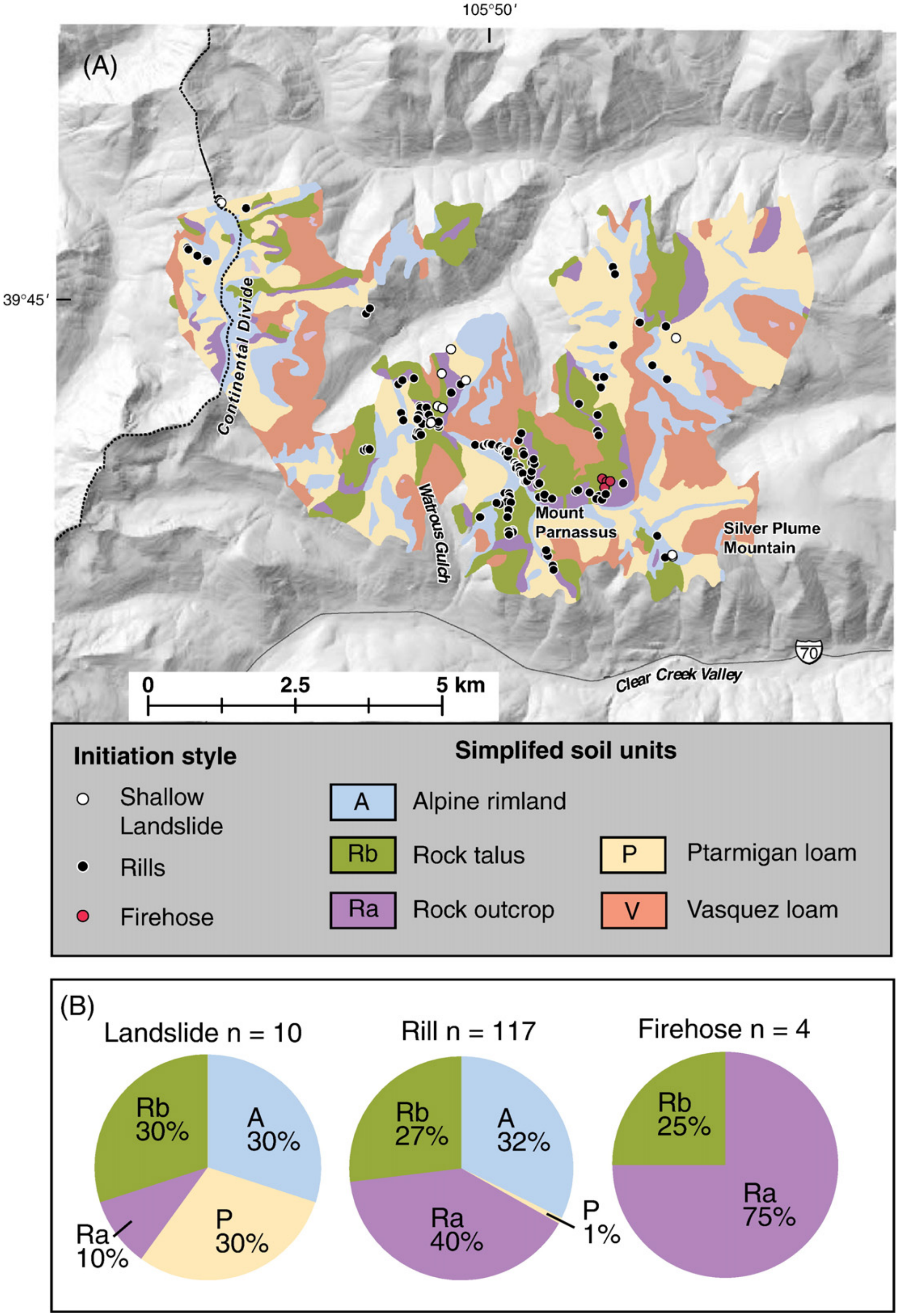

Fig. 7. Generalized soils map (A) from Retzer (1962) and debris flow initiation locations in the northern half of the study area. Pie charts (B) show the percentage of the initiation locations by process type and generalized soil unit.

and Clear Creek Counties and moved to the northeast (Wieczorek et al., 2003). The heaviest hourly rainfall estimated from NEXRAD was about 38 mm between 14:00 and 15:00 MST (Fig. 6). Storm totals estimated from NEXRAD show a similar spatial pattern to the peak hourly rainfall. Over the Grizzly Peak SNOTEL station, NEXRAD estimated about 19 mm of rainfall between 14:00 and 15:00 MST (Fig. 6), whereas the rain gage recorded 20 mm (Fig. 4). NEXRAD rainfall estimates (Fig. 6) compared with a contour map of debris flow initiation locations shows that most of the debris flows were initiated where rainfall intensities exceeded 13 mm/h, the threshold Caine (1976) identified for the San Juan Mountains in southwest Colorado; however, the heaviest hourly rainfall and highest concentrations of debris flows were not spatially coincident. This observation indicates that other variables, such as topography and character of hillside materials, in addition to adequate intensity and duration of rainfall, influenced the location and abundance of debris flow activity. The influence of these other variables on debris flow source areas is explored in a later part of this paper.

4. Results and discussion

Field observations of source areas showed that debris flows were initiated by three processes: i) initiation by shallow landsliding, ii) initiation by coalescing rills, and iii) initiation by the “firehose effect.” Comparison of a generalized soils maps of the northern half of the study area by Retzer (1962) with debris flow initiation locations (Fig. 7A) and field observations indicate that the initiation process was function of the abundance and character of hillside materials. In general, as the soils become better developed the amount of runoff generated during the storm decreased, therefore the number of debris flows initiated by runoff processes decreased. Landslide debris flows initiated in Ptarmigan soils (i.e., well-developed soil) and in thick colluvium, rilling debris flows initiated in loose colluvium where bedrock is close to the surface, and fire hose debris flows were commonly initiated on bare rock outcrop (i.e., no soil development).

4.1. Initiation by shallow landsliding

About 11% of the debris flows were mobilized from shallow, discrete landslides on steep, mostly-vegetated hillslopes. The mechanism by which rainfall infiltration initiates shallow landslides is well known (Campbell, 1975; Reid et al., 1988; Iverson et al., 1997; Iverson, 2000). Debris flows generated from shallow landslides are generally associated with colluvial hollows on steep, soil-mantled hillslopes in temperate environments such as the Oregon Coast Range and the San Francisco Bay area in North America. To our knowledge, no accounts of this type of debris flow in the alpine environments of the United States are in the literature.

Shallow landslides were most abundant in Ptarmigan soils and in areas mantled with thick colluvium (Table 1; Fig. 7B). All the shallow landslides that we visited in the field started on hillslopes covered with alpine tundra vegetation and had scars that were roughly crescent-shaped with steep headscarps commonly 0.5 to 2 m high. The head scarps formed in a 10–20 cm thick, organic (A horizon) soil that has abundant small roots underlain by sandy colluvium containing angular bedrock clasts. The landslides were translational failures about a meter thick and the failure surface was typically within the colluvium and not at the interface between colluvium and bedrock as often described for shallow landsliding in temperate climates (e.g. Campbell, 1975;

Fig. 8. Photograph of debris flow that mobilized from a shallow landslide on the west flank of Grizzly Peak. Relief from the landslide to the stream valley at the bottom of the photograph is about 500 m. Photograph taken on 29 July 1999.

Ellen et al., 1988). Shallow landslides mobilized as debris flows and traveled downslope, often without eroding the hillslope (Fig. 8). This is in contrast to the firehose and rilling debris flows, in which most of the transported material was derived from the hillslope, channels, talus deposits, and debris fans. The debris flows generated by landslides left poorly sorted, matrix-supported levees along their lateral margins and lobate deposits at the downslope end.

4.2. Initiation by rilling

About 79% of the debris flows initiated as single rills or as a system of coalescing rills on unvegetated hillslopes underlain by noncohesive materials. Rills on hillslopes in the study area were as much as 1 m in width and depth. Often downslope from the system of rills, single channels or gullies were eroded into the surface. Gullies were generally deeper and wider than the rills and as much as 4 m in width and depth. Rills in other settings are typically smaller and have widths and depths of a few tens of centimeters (Selby, 1993, p. 232). Rill formation is generally attributed to the concentration of overland flow in microchannels. The concentration of overland flow increases the flow depth and thus the component of shear stress acting parallel to the slope mobilizing loose sediment (Horton, 1945). Slight variations in hillslope topography play an important role in localizing rill formation and in the subsequent creation of a series of alternating benches and steps, or plunge pools, in the rill channel (Rickenmann, 1997; Gimenez et al., 2004). The formation of rill channels takes place by erosion and deepening of plunge pools by turbulent flow (Cannon et al., 2003) upslope progression of the headwalls of plunge pools, (Johnson and Rodine, 1984; Bull, 1997; Collison, 2001) and bank failures. Experimental results (Oostwoud Wijdenes and Ergenzinger, 1998) showed that erosion and sediment transport by rainfall and overland flow on hillslopes occurred over a continuum of water–solid concentrations ranging from clear-water flow, to concentrated, and hyperconcentrated bed load and suspended flow, to debris flow. Debris flows temporarily dammed rill channels, creating new rills or plunge pools, and quickly changed to hyperconcentrated flows when mixed with additional water from overland flow. Debris flow initiation from rilling is common on steep hillslopes where vegetation has been recently burned by wildfire (Johnson and Rodine, 1984; Wohl and Pearthree, 1991; Meyer and Wells, 1997; Cannon et al., 2001a, 2001b).

In the study area, rilling was most abundant in areas mapped as rock outcrop and talus and alpine rimland (Fig. 7B). The hillslopes that generated rill debris flows were typically topographically concave (swales) and mantled by abundant loose angular clasts in a sandy matrix. The colluvium mantling the swales is of variable depth and bedrock is often exposed (Fig. 9). The patchy colluvial cover yields a spatially variable hydrologic response to rainfall. The loose, coarse-grained colluvium is highly permeable. The saturated hydraulic conductivity ranges up to about 1.5×10^{-3} m/s based on two field tests at depths of <1 m using a Guelph permeameter (Fig. 9). The bedrock outcrops in the swales are typically fractured but in comparison to the colluvium,

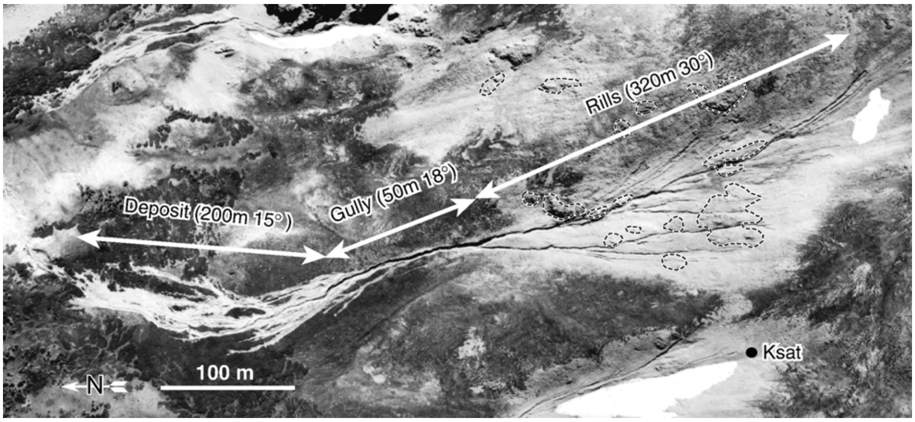

Fig. 9. Vertical aerial photograph of debris flows on the north side of Mount Parnassus showing the rill network, incised gully, and deposit. The 320-m-long rill zone is an unvegetated swale and is an area of late lying snow with an average slope of 30° typical of Alpine rimland described in Table 1. Areas where mostly bedrock is exposed at the ground surface are outlined with dashed lines. The 50-m-long gully is incised more than 2 m in depth with small lateral levees overlying the pre-debris flow surface. The 200-m deposition zone is a composite of crosscutting gullies, levees, and fan deposits. The saturated conductivity of the colluvium was measured at the location labeled “Ksat” in the lower right-hand part of the photograph.

are relatively impermeable. Thus during the 1999 thunderstorm, rill formation was largely dependent on the local hydrologic response of the various hillslope materials. Rills were probably formed by a combination of overland flow from relatively impermeable materials and lateral flow through highly permeable materials. Downslope of the swales, the rills coalesced and deepened forming gullies several meters in width and depth that incised the middle parts of the hillslopes (Fig. 9). The erosion of channel material during the July 1999 rainstorm was greatest in these gullies. At debris flow sites initiated by rilling within the study area, the evidence of debris flow (e.g. matrix-supported levees and debris lobes) highest on the hillslope was often observed along the flanks of gullies. Downslope from the gullies were debris fans and ephemeral or perennial stream channels. The heads of these debris fans, which in some cases are formed in part by rockfall deposits, were commonly incised by the gullies. Debris flows from 1999 primarily deposited material on the middle and lower parts of the previously existing debris fans. Fig. 9 is an aerial photograph showing a debris flow initiated by rilling in Alpine rimland as mapped by Retzer (1962) on the north side of Mount Parnassus. Unpublished field mapping indicates that deposits were formed by a sequence of flows or flow pulses that left a series of crosscutting boulder rich levees and lobes of debris. We also identified erosional and depositional features in many of the deposits that appeared to be dominantly fluvial in origin.

4.3. Initiation by the “firehose effect”

About 10% of the debris flows were initiated by what Johnson and Rodine (1984) have termed the “firehose effect”. The firehose effect is caused by the mobilization of material by a concentrated flow of water, just as if the material had been washed away by a “firehose” (Fryxell and Horberg, 1943; Curry, 1966; White, 1981; Johnson and Rodine, 1984; Coe et al., 1997; Griffiths et al., 2004). Eyewitness accounts in both the Colorado Rocky Mountains (Curry, 1966) and the Dolomites of Italy (Berti et al., 1999) describe the firehose process as the concentration of surface flow in steep rocky headwaters channels that impact loose debris. In Italy, debris flows were generated by the progressive entrainment of surface water with loose debris from the channel bed (Berti et al., 1999; Berti and Simoni, 2005). Material was transported as a debris flow, the deposit of which came to rest in the channel, blocking it. The debris dam was breached creating another debris flow that scoured the channel and entrained debris as it flowed.

In the study area, firehose debris flows were typically formed on steep bedrock slopes (rock outcrop of Retzer, 1962, Table 1, Fig. 7B) above talus deposits. Overland flow of water resulting from intense rainfall was concentrated in steep, predominately bedrock-lined channels (Fig. 10). When the flow reached the base of the bedrock slopes, it mobilized the bouldery talus material and eroded deep gullies into the heads of the

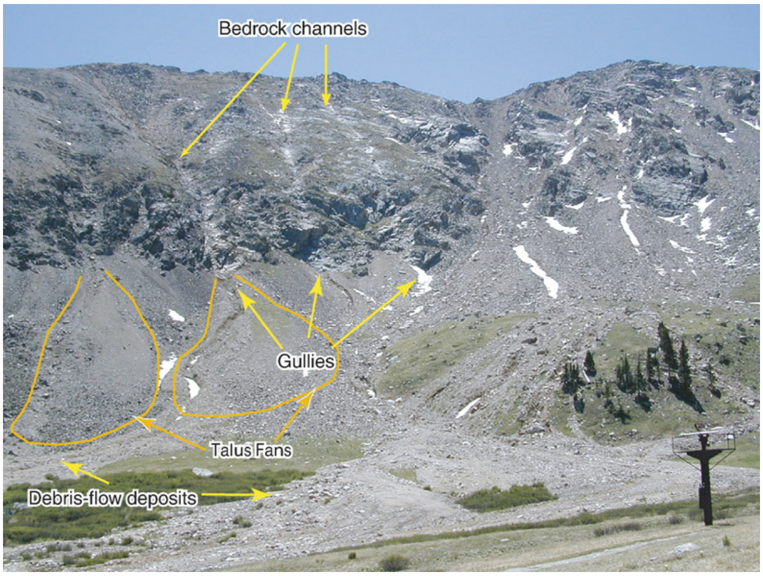

Fig. 10. Northwest flank of Lenawee Mountain showing the scoured bedrock channels and deeply incised gullies at fan heads. Much of the debris deposited in the foreground was scoured from the fans by streams of water running off the bedrock channels thereby applying a ‘firehose effect’. Relief from foreground to the top of the ridge is about 400 m. Photograph taken on June 13, 2002.

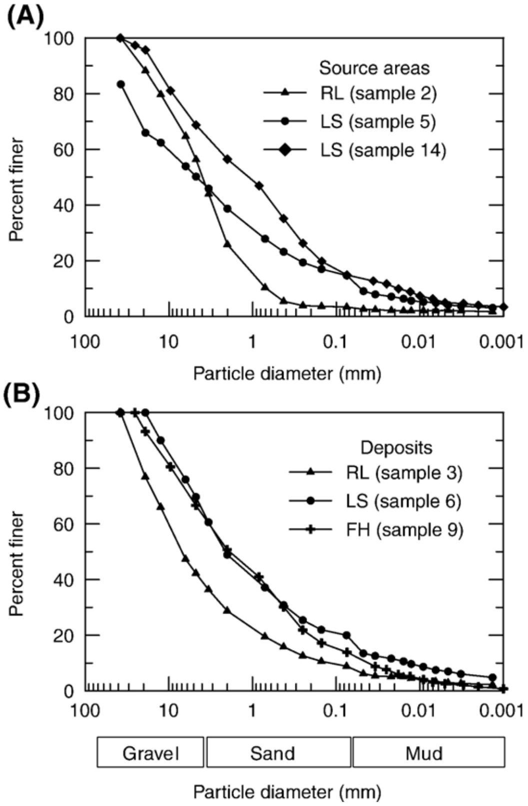

Fig. 11. Particle-size distributions for samples of matrix collected from (A) colluvium in debris flow source areas and (B) matrix of debris-flow deposits. Samples 2 and 3 were collected from the source and deposit of the same debris flow. RL=rill, LS=landslide, and FH=firehose initiation styles.

talus deposits and debris fans. The material was deposited lower on the fans and downslope in stream valleys.

4.4. Particle-size distribution of initiation locations and deposits

The particle-size distribution of colluvium in debris flow source areas and of debris flow deposits was determined by visually estimating the relative percentage of clasts and matrix in the field and from photographs and by mechanical sieving and hydrometer tests of matrix samples in the laboratory. For comparison, five samples were collected from rilling and landslide debris flow source areas and nine samples were collected from resulting debris flow deposits. Large angular clasts, defined here as particles larger than about 75 mm in long dimension, are abundant in both the hillside colluvium and in debris flow deposits. Large clasts make up between 40% and 60% of the colluvium in debris flow source areas and between 40% and 70% of debris flow deposits. Results from laboratory tests (Fig. 11A and B) show that the matrix (all materials smaller than 75 mm) of both the colluvium and debris flow deposits are dominated by sand- and gravel-sized particles and that the percentage of fine-grained material (mud) ranges from 2% to 12%.

Compared to the colluvium in initiation locations, debris flow deposits are generally richer in mud (silt and clay size particles), which was entrained from the channel as the debris flow traveled downslope. At Watrous gulch, one debris flow location where samples were collected from both the source area (sample 2) and the deposit (sample 3), the deposit contains a slightly higher percentage of mud, a lower percentage of sand and gravel, and a higher percentage of clasts compared to the source area (Fig. 11A and B). Based on field observations, the clasts in this deposit are larger and more rounded than in the colluvium of the source area.

The entrainment of fines either released by wildfire or from fine-grained deposits in channel banks and pools has been identified as important in rheological explanations of debris flow deposit morpohology (e.g. Johnson and Rodine, 1984; Meyer and Wells, 1997; Berti et al., 1999). Rheological explanations of debris flow deposition typically state that the fluid yield strength of debris slurry is increased suddenly as widespread pressure diffusion freezes the levees and lobes that characterize debris flow deposits. Recent experimental work has shown that deposit morphology results not from the fluid yield strength of the debris slurry, but rather from the preferential sorting of large particles and friction concentration along flow margins where high pore-fluid pressures are absent (Major, 2000; Iverson, 2003). Slurries with as little as 2% mud exhibit debris flow behavior and leave deposits with characteristic debris flow morphology, and additional mud helps sustain high pore-fluid pressures in the interior of debris flow slurries promoting travel (Iverson, 1997; Iverson, 2003).

4.5. Topographic characterization of debris flow initiation locations

Examination of topographic variables derived from 10-m digital elevation data shows that debris flows were initiated on steep slopes in small drainage basins. Slopes angles at debris flow initiation locations ranged between 9° and 62° (Fig. 12A) and are normally distributed about a mean of about 37°. Contributing areas were typically <8000 m^2 (Fig. 12B) and are approximately log-normally distributed. Comparisons between the distributions of slope angles and upslope contributing areas of debris flow initiation locations with the distributions

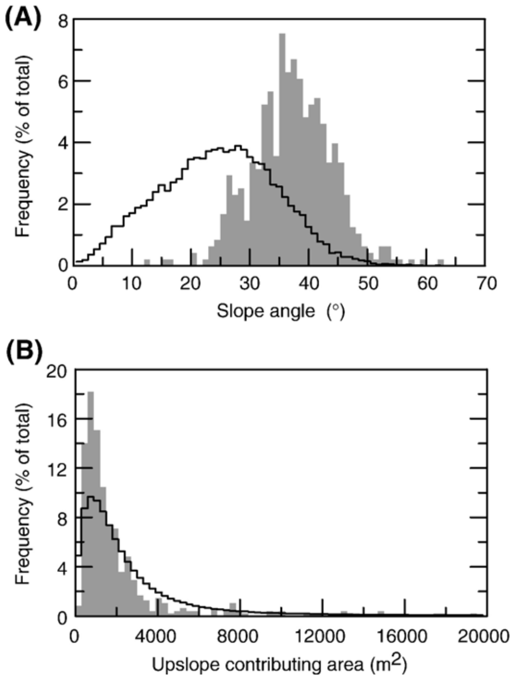

Fig. 12. Frequency distributions of (A) slope angles, and (B) upslope contributing areas for debris flow initiation locations (bars) and for the entire study area (black solid lines). Frequency is expressed as the percentage of total for each data set.

of the same variables in the study area as a whole show that slope angles >32° and contributing areas $<3000\ m^2$ were preferentially susceptible to debris flow initiation (Fig. 12). Landslides initiated debris flows on shallower slopes that those initiated by rilling and firehose processes (Fig. 13A). Rills were initiated over the broadest range of slope angles. Statistical tests (ANOVA) show that the mean slope angles vary significantly ($p<0.0001$) among the initiation styles (Fig. 13A). Contributing areas above rills varied over the largest range compared to firehose or landslide initiation locations, however, statistical tests (ANOVA) show that the mean contributing area is not statistically different ($p>0.25$) among the initiation styles (Fig. 13B).

Distance traveled by the debris flows, defined as the distance between the mapped initiation location and the distal end of the deposit, varies significantly ($p<0.01$) among the initiation styles based on statistical tests (ANOVA) with the firehose flows traveling the least distance. Because multiple rills often contributed to a debris flow deposit, in general these debris flows traveled the furthest (Fig. 13C). Additionally, field observations indicate that debris flows from landslides typically lost material as they traveled downslope, whereas debris flows from runoff processes (firehose and rilling) increased in volume as they traveled downslope. These observations suggest that debris flows initiated by runoff processes are potentially more hazardous than landslide debris flows.

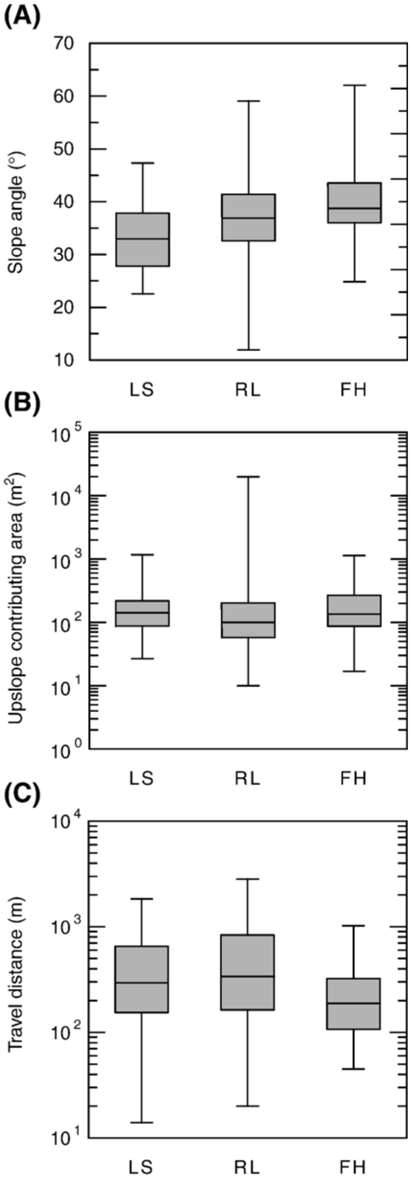

Fig. 13. Box and whisker plots showing the median (horizontal line), upper and lower quartiles (gray boxes), and range of slope angles (A), upslope contributing areas (B), and travel distances (C) for shallow landslides (LS, $n=54$), rills (RL, $n=378$), and firehose effect (FH, $n=46$). Mean slope angles (ANOVA $p<0.0001$) and travel distances (ANOVA $p<0.05$) vary significantly among the process types whereas mean contributing areas do not. See text for details of ANOVA statistical tests.

Following Dietrich et al. (1992) and Montgomery and Dietrich (1994a), we plotted upslope contributing area and the tangent of slope angle in log–log space for each debris flow initiation location for the three initiation styles with the expectation that debris flows initiated by erosion would lie to the upper left of those mobilized from landslides (Fig. 14A–C). Although the slope–area data for each process type overlap, in general, the debris flows initiated by overland flow processes (rills and firehose) plot towards the right side of the slope–area graph compared to the debris flows initiated by landsliding. This reflects the statistically significant variance in the mean slope angles of the initiation locations described above and indicates that the high alpine landscape of the Front Range cannot be distinctly divided into areas dominated by overland flow and landslide erosion processes based on topography alone, unlike what has been suggested by Dietrich et al. (1992) and Montgomery and Dietrich (1994a) for soil-mantled landscapes in humid, temperate climates. This is because overland flow that drives the rilling and firehose processes is not saturation excess overland flow, but rather overland flow that is generated by intense rainfall on surficial materials with relatively low infiltration capacities (Horton, 1933; Betson, 1964). Thus, channel heads are not spatially coincident with the initiation of erosion and entrainment of sediment by debris flows in landscapes with abundant exposed bedrock and variable surficial cover such as the alpine of the Front Range.

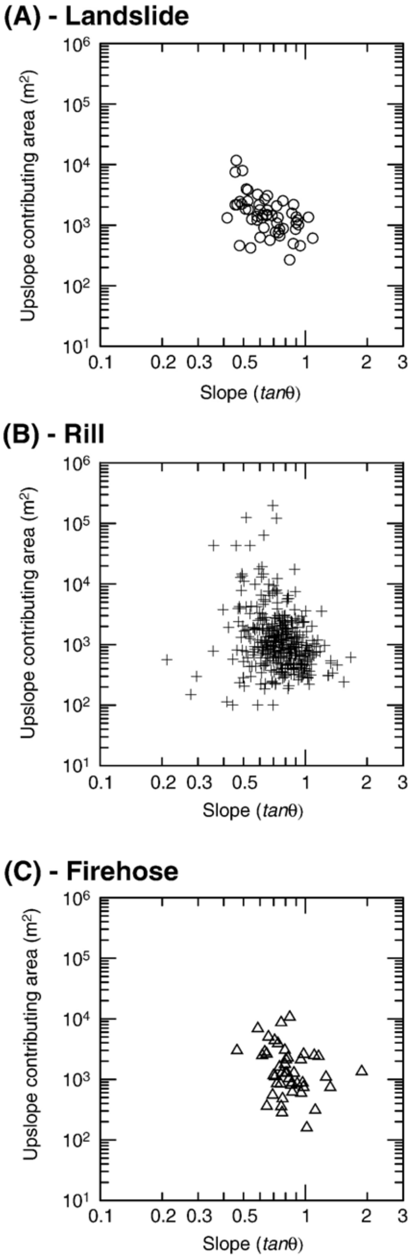

Fig. 14. Slope-contributing area relations for shallow landslides (A), rills (B), and firehose effect (C) initiation locations. Slopes (θ) are shown as the tangent of the angle in degrees calculated from a 10-m USGS DEM following Montgomery and Dietrich (1994a).

4.6. Implications for debris flow hazard mapping

Identification of the areas that might be inundated by debris flows (debris fans) and estimates of flow volume are required to quantify debris flow hazard. Debris flows initiated by overland flow processes entrained material from hillslopes and stream channels and increased in volume, whereas debris flows initiated from shallow landslides tended to deposit material as they traveled. Thus, rilling and firehose debris flows are potentially more hazardous than debris flows initiated from shallow landslides. Our results show that slope–area relationships developed to identify the location of channel heads in soil-mantled landscapes with humid climates (Dietrich et al., 1992; Montgomery and Dietrich, 1994a) cannot be reliably used to partition the alpine landscape into dominant debris flow initiation style because of the spatial variability of hillside materials and their transient response to high intensity rainfall. We suggest that surficial geologic maps depicting the character and abundance of hillside and channel materials, in addition to digital elevation data, are needed for quantitative regional debris flow hazard assessments in alpine landscapes. A crude, but effective surficial geologic map could consist of four basic units; bedrock, talus, bare colluvium, and well-vegetated colluvium.

5. Conclusions

Intense rainfall on 28 July 1999 associated with the North American Monsoon initiated widespread debris flows in the central Front Range of Colorado. Overland flow processes (rilling and firehose effect) initiated most of the 480 debris flows with shallow landsliding accounting for about 11% of the total number of debris flows. The character and spatial distribution of the surficial cover (soil development and vegetation) and the presence of bedrock channels high on hillslopes of adequate gradient, rather than topography alone, appear to control the initiation processes of debris flows in the high alpine of the central Front Range. Volumes of runoff-initiated debris flows increased as they traveled downslope, whereas volumes of infiltration-initiated debris flows decreased as they traveled downslope. Thus, runoff-initiated debris flows are potentially more hazardous than infiltration-initiated debris flows. Because debris flows in alpine environments in the Front Range appear to be initiated largely by erosive processes, debris flow hazard assessment of this area should not rely on mapping techniques developed for soil-mantled landscapes that are susceptible to shallow landsliding. Alternately, we advocate an approach that identifies both the potential source areas using maps of surficial cover and topography and the areas likely to be inundated by debris flows (debris fans).

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