

Wildfire as a hydrological and geomorphological agent

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Abstract

Wildfire can lead to considerable hydrological and geomorphological change, both directly by weathering bedrock surfaces and changing soil structure and properties, and indirectly through the effects of changes to the soil and vegetation on hydrological and geomorphological processes. This review summarizes current knowledge and identifies research gaps focusing particularly on the contribution of research from the Mediterranean Basin, Australia and South Africa over the last two decades or so to the state of knowledge mostly built on research carried out in the USA.

Wildfire-induced weathering rates have been reported to be high relative to other weathering processes in fire-prone terrain, possibly as much as one or two magnitudes higher than frost action, with important implications for cosmogenic-isotope dating of the length of rock exposure. Wildfire impacts on soil properties have been a major focus of interest over the last two decades. Fire usually reduces soil aggregate stability and can induce, enhance or destroy soil water repellency depending on the temperature reached and its duration. These changes have implications for infiltration, overland flow and rainsplash detachment. A large proportion of publications concerned with fire impacts have focused on post-fire soil erosion by water, particularly at small scales. These have shown elevated, sometimes extremely large post-fire losses before geomorphological stability is re-established. Soil losses per unit area are generally negatively related to measurement scale reflecting increased opportunities for sediment storage at larger scales. Over the last 20 years, there has been much improvement in the understanding of the forms, causes and timing of debris flow and landslide activity on burnt terrain. Advances in previously largely unreported processes (e.g. bio-transfer of sediment and wind erosion) have also been made.

Post-fire hydrological effects have generally also been studied at small rather than large scales, with soil water repellency effects on infiltration and overland flow being a particular focus. At catchment scales, post-fire accentuated peakflow has received more attention than changes in total flow, reflecting easier measurement and the greater hazard posed by the former. Post-fire changes to stream channels occur over both short and long terms with complex feedback mechanisms, though research to date has been limited.

Research gaps identified include the need to: (1) develop a fire severity index relevant to soil changes rather than to degree of biomass destruction; (2) isolate the hydrological and geomorphological impacts of fire-induced soil water repellency changes from other important post-fire changes (e.g. litter and vegetation destruction); (3) improve knowledge of the hydrological and geomorphological impacts of wildfire in a wider range of fire-prone terrain types; (4) solve important problems in the determination and analysis of hillslope and catchment sediment yields including poor knowledge about soil losses other than at small spatial and short temporal scales, the lack of a clear measure of the degradational significance of post-fire soil losses, and confusion arising

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from errors in and lack of scale context for many quoted post-fire soil erosion rates; and (5) increase the research effort into past and potential future hydrological and geomorphological changes resulting from wildfire.

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1. Introduction

Over short and long timescales, wildfire (i.e. uncontrolled or naturally occurring fire) can be an important, if not the major, cause of hydrological and geomorphological change in fire-prone landscapes. Rock surfaces may be subject to widespread weathering, though damage to or loss of vegetation and litter cover represent the obvious changes in the landscape and they can affect the interception, evapotranspiration and storage of rainfall and can also influence snow accumulation and melting behaviour in affected landscapes. The heating of soils tends to alter their physical and chemical characteristics, including water repellency behaviour and stability of aggregates. These changes often result in enhanced hydrological and geomorphological activity, with considerably more overland flow on slopes and increased discharge and peakflow, channel changes and substantially increased hillslope soil redistribution and catchment sediment yields. Where conditions are suitable, the fire-induced changes can cause increased wind erosion, increases in snow-avalanche activity, gravity flow of dry sediments, landslides and debris flows. The magnitude and duration of post-fire increased hydrological and geomorphological activity can vary enormously and depend on an often complex interplay of factors including site and fire characteristics, and also post-fire rainfall patterns. Many researchers have attempted to determine accurately and predict its hydrological and geomorphological effects, but this has been made difficult by the unpredictable nature of wildfire. Building a coherent knowledge base, however, has proved difficult as results of this work vary widely, associated with regional differences in climate, terrain and fire characteristics, but also due to differences in research approaches and scales of investigation. Furthermore, research has been inter-disciplinary in nature and also has hazard and conservation implications so that much of the literature is disseminated amongst a number of government agency reports and conference proceedings as well as peer-reviewed academic journals across a range of disciplines. There have been a number of reviews dealing with aspects of fire impacts on hydrology and geomorphology (e.g. Anderson et al., 1976; Tiedemann et al., 1979; Swanson, 1981; McNabb

and Swanson, 1990; Robichaud et al., 2000; Neary et al., 2003), but some are now outdated, most appear in non-peer reviewed literature, feature mainly post-fire conservation issues or deal more with prescribed fire rather than the impacts of wildfire. In addition, there is an emphasis on North American examples with much less attention paid to the literature from other parts of the world (notably Europe, and the Mediterranean Basin, southern Africa and Australia) where fire impacts have differed due to different types of natural environmental conditions and land use history.

This paper attempts to provide a critical review of the available literature concerning the hydrological and geomorphological impacts of wildfire. It considers impacts on hydrology, fluvial geomorphology, rock weathering, mass movement processes and soil erosion from the range of fire-affected regions that have been studied world-wide. Over the last 20 years or so, a growing body of literature has been concerned with wildfire impacts in the Mediterranean Basin in particular, but also other parts of the world, notably Australia and South Africa, which has given different perspectives on hydrological and geomorphological impacts. Examples in this review are drawn mainly from studies in forest and woodland and to a lesser extent from scrub and grassland areas, with an emphasis on the results of field monitoring and observations of processes under natural rather than simulated rainfall. The hydrological and geomorphological effects of prescribed fires, the modelling of wildfire impacts and remedial measures to alleviate these impacts are not considered.

2. Wildfire: global significance, causes, frequency and severity

2.1. Global significance and causes

Wildfire is an important disturbance factor in most vegetation zones throughout the world and is believed to have been more or less common since late Devonian times (Schmidt and Noack, 2000). In many ecosystems it is a natural, essential, and ecologically significant force, shaping the physical, chemical and biological attributes of the land surface. The main causes are lightning, volcanoes and human action, the latter

being now the main cause in, for example, the densely populated Mediterranean Basin (FAO, 2001). According to the FAO (2001), several hundred million hectares of forest and other vegetation types are estimated to burn annually throughout the world consuming several billion tonnes of dry biomass. For example, in AD 2000, c. 351 million ha burnt. In the Mediterranean Basin alone during the 1990s, approximately 50,000 wildfires affected annually c. 600,000 ha of forest and other wooded land. In response to global warming over the next few decades, fire extent is expected to increase (McCarthy et al., 2001).

2.2. Frequency and severity

Many hydrological and geomorphological impacts of wildfire are influenced by its frequency and severity. Frequency varies widely amongst vegetation types and climates. For example, it can vary from 6000 years in a temperate European forest to less than 50 years in a south-east Australian eucalypt forest (Wheelan, 1995). Return intervals are affected by climate and more recently by human action, which has caused additional ignitions and directly altered fuel loads and characteristics. For example, in Europe, long-term intensive land use has produced a patchwork of forest, shrub and cultivated land with species composition and fuel loads bearing little relationship to pre-settlement patterns. Consequently, current fire frequencies and behaviour are far from ‘natural’. Rural depopulation leading to increases in fuel load in abandoned areas (FAO, 2001) and the introduction of highly flammable tree species (especially pine and eucalypt) have increased fire frequency. Intensive land management in western North America has had a comparatively short history with consequently different impacts on wildfire behaviour. For example, in some regions European settlement initially reduced fuel load by forest thinning and cattle grazing during the last century, but fire suppression subsequently increased fuel loads to artificially high levels and led to less frequent but more severe wildfire, as documented for ponderosa pine forests in the southwest (Allen et al., 2002). In Australia, the distribution of tree species and amounts of fuel load were affected by, and reached relative equilibrium with, the Aboriginal population prior to European settlement (Dodson and Mooney, 2002). Today, fire suppression is common but less comprehensive than in North America. Replacement of native eucalypts in some areas with non-native, highly flammable coniferous species has introduced different fire behaviour and recovery patterns (Wheelan, 1995).

Fire severity depends on the interactions between burning, especially its duration (that is, the length of time burning occurs at a particular point) and intensity (i.e. the rate at which thermal energy is produced), and the characteristics of the biomass, soil, terrain and local climate. Burn severity is commonly, though not always, classified according to the degree of destruction of above-ground live and dead biomass, which is dependent on fuel type, amount and characteristics and also burn intensity and duration (see review by Neary et al., 1999 for more detail and Table 1 for an example of a fire intensity and a fire severity classification). Despite its acclaimed importance for understanding hydrological and geomorphological impacts (e.g. Swanson, 1981; Prosser, 1990; Inbar et al., 1998; De Luis et al., 2003), existing classifications of fire severity are not always reliable predictors of some of the critical changes to the soil with respect to these impacts.

In forests, fires are often classified into three main types: ground, surface and crown (or canopy). Ground fires affect the organic layer of decaying leaves and other plant parts. They can advance slowly, usually generating fires of moderate intensity. Surface fires also affect plants and shrubs and scorch the bases and crowns of trees. Crown fires burn the higher leaves and branches and may be dependent on surface fires or independent of them with flames spreading directly from

Table 1

Example of a fire intensity and an associated severity rating for eucalypt-dominated sclerophyll vegetation communities in south-eastern Australia based on Cheney (1981), Jasper (1999) and Shakesby et al. (2003). Fire intensity (related to the rate at which thermal energy is produced) and severity (related to duration of burning) are broadly related

Fire intensity ^a (kW m ⁻¹)	Max. flame height (m)	Severity rating	Post-fire vegetation characteristics
≤500	1.5	Low	Only ground fuel and shrubs <2 m high burnt
501–3000	5.0	Moderate	All ground fuel and shrub vegetation <4 m high consumed
3001–7000	10.0	High	All ground and shrub vegetation consumed and lower tree canopy <10 m high scorched
7001–70,000	10–30	Very high	All green vegetation including tree canopy up to 30 m, and woody vegetation <5 mm diameter consumed
70,001–100,000+	20–40	Extreme	All green and woody vegetation <10 mm diameter consumed

^a The fire intensity index, as defined by Byram (1959).

crown to crown (www.fs.fed.us/rm/pubs/rmrs_rn022_02). Although typically the most spectacular form of wildfire and indirectly affecting hydrological and geomorphological processes through reduced rainfall interception and evapotranspiration, crown fires may actually have little or no direct effect at ground level, where the nature of surface and ground fires may be more important.

3. Direct effects of fire on rock, vegetation and soil

Apart from the visible reduction of vegetation and litter and also the changes to faunal populations and their activities resulting from wildfire, burning can cause significant changes to the ground surface that either constitute a direct geomorphological change (e.g. fire-induced rock weathering) or cause hydrological and geomorphological processes to operate at changed rates during the post-fire period until environmental conditions similar to those found prior to burning become re-established. Amongst the changes that can influence hydrological and geomorphological processes are alterations to aggregate stability, porosity, organic matter and water repellency characteristics. These are considered in the following sub-sections. Such changes influence the hydrological and geomorphological processes during the post-fire period, and these are discussed in detail in Section 4.

3.1. Rock weathering

That fire can be an important rock weathering agent has been long recognized (e.g. Blackwelder, 1926;

Griggs, 1936; Emery, 1944; Journaux and Coutard, 1974; Ballais and Bosc, 1992), but much of what is known is based on descriptions rather than quantitative data (e.g. Ollier, 1983; Ollier and Ash, 1983; Birkeland, 1984; Bierman and Gillespie, 1991; Dragovich, 1993). This is perhaps not surprising given the inevitable unpredictability of fires capable of causing substantial weathering. The need nevertheless for quantitative data has come sharply into focus in recent years with the development of various rock exposure dating methods, which depend on assumptions about long-term rates of weathering and erosion. Consequently, even direct measurements made after individual fires have been published (Zimmerman et al., 1994; Dorn, 2003). Laboratory experiments have revealed insights into how different lithologies respond to fire and their susceptibility to weathering following exposure to fire (e.g. Goudie et al., 1992; Allison and Goudie, 1994; Allison and Bristow, 1999). It has been shown, for example, that fire effects depend to a large extent on rock physical properties and vary with lithology, boulder size and water content. Such experiments, however, have not so far been applied to questions of the style or rates of fire-induced rock weathering.

The most commonly reported weathering effect of fire is spalling, in which often lensoid-shaped rock flakes up to 3 cm in size become detached (Fig. 1). After fire, such spalling affected over 70–90% of exposed surfaces of predominantly granitic gneiss boulders on moraines in the Fremont Lake region, Wyoming, with an estimated average thickness of c. 2.5 mm of rock removed (Zimmerman et al., 1994).



Fig. 1. Weathering of Hawkesbury Sandstone bedrock, Nattai National Park, New South Wales, Australia, caused by wildfire in December 2001. Lighter-toned patches indicate where lensoid-shaped flakes have been removed. The contrast in tone between these patches and the remaining dark rock surfaces suggests that the flakes were removed after the fire. Note hand trowel for scale.

Adamson et al. (1983) estimated that c. 50% of the original surfaces of burnt sandstone surfaces in the Blue Mountains, New South Wales, Australia had lost 2–6 kg m⁻² after a wildfire. From remeasurements after a fire in 2000 in the Sierra Ancha Mountains, Arizona of boulders originally surveyed in 1989, Dorn (2003) considered that deriving an average figure for the weathering effect was misleading in view of post-fire measurements on two rock types (sandstone and diorite) revealing a strongly bimodal pattern of losses, with c. 20–60% of surfaces losing > 76 mm thickness of rock and c. 30–58% of surfaces showing no loss. Only a few measurements lay near the average figure located between these two thickness extremes. Ballais and Bosc (1994) also noted a difference in fire-induced spalling between different rock types: whereas limestone and sandstone were affected, gneiss and mica schist were apparently not.

In addition to spalling, there are reports of vertical fracturing (affecting an entire boulder usually <30 cm in diameter) and irregular linear and curvilinear fractures (Ollier, 1983; Ollier and Ash, 1983; Dragovich, 1993). Observations of unscorched fractured faces on spalled fragments (and on cracked surfaces) and correspondingly fire-blackened faces of boulders pock-marked by unburnt patches (e.g. Ollier and Ash, 1983) are strong evidence that rock breakdown takes place at a late stage during the fire (Fig. 1), possibly even during cooling or indeed up to days or even months after fire (Ballais and Bosc, 1994).

Over the long term, observations suggest that fire can lead to considerable weathering. Dragovich (1993) cited boulder shape in the Pilbara, Western Australia as evidence of this. Large boulders had both rounded and angular faces whereas small clasts were mostly angular, explained by fire weathering causing rounding of larger boulders and occasional vertical fracturing in small ones. The small angular clasts were viewed as representing rock fragments spalled during fire. She concluded that in semi-arid and arid environments where chemical weathering is slow and limited to a thin surface layer, mechanical breakdown of rocks by fire must make a major contribution to rock weathering, a view endorsed by Dorn (2003). Humphreys et al. (2003) estimated that if just 1% of rock surfaces in the sandstone tablelands of south-eastern Australia were affected every 20 years, this would represent a denudation rate of 6 m My⁻¹ (assuming a rock density of 2 g cm⁻³). This conservative estimate is similar to rates of landscape lowering estimated from knickpoint retreat, valley widening, cliff retreat and plateau lowering (Van der Beek et al., 2001) suggesting that it is a

significant erosion process in this humid temperate environment. In mountain locations in southern France prone to freeze–thaw weathering, Ballais and Bosc (1994) estimated that rock weathering by fire might be as much as 10–100 times more effective than frost action over the long term.

3.2. Removal of vegetation and litter cover and changes to microbial and faunal activity

Depending on its severity (Section 2.2), wildfire can remove some or all of the vegetation and litter cover, thereby altering key variables in the hydrological cycle. It temporarily reduces or stops transpiration, interception and surface storage capacity for rain (retention and detention) (Tiedemann et al., 1979; Loaiciga et al., 2001). As regards the latter, fire tends to destroy obstacles that can reduce water storage allowing erosive overland flow to occur more readily on the soil surface (see Section 4), although remaining litter may form so-called litter dams which tend to trap sediment on shallow-angled slopes (e.g. Mitchell and Humphreys, 1987; Díaz-Fierros et al., 1994). Estimates have been made of water storage capacity for each centimetre thickness of litter, examples being 0.5 mm of water depth for a pine forest in the USA (Neary et al., 2003) and 3 mm for commercial eucalypt stands in Portugal (Leighton-Boyce, 2002). Exposure of the bare soil surface leaves it susceptible to raindrop impact and entrainment by overland flow, but also imposes a different thermal regime on the soil associated with changes in soil moisture dynamics and increased solar heating.

Microbial populations and mycorrhizae activity in the top few centimetres of the soil are initially reduced following burning, but numbers of certain autotrophic microbes can quickly surpass pre-fire levels (Neary et al., 1999). Soil microbial and fungal activity can directly affect soil erodibility through the secretion of cohesive compounds and the production of stabilising fungal hyphae. By playing an important role in nutrient cycling, microbial activity is also relevant indirectly to hydrological and geomorphological processes through its impact on post-fire vegetation recovery rates (Klopatek, 1987). Soil-dwelling invertebrates are more mobile than micro-organisms and therefore have a greater potential to escape heating by burrowing deeper into the soil (Certini, 2005). Post-fire faunal activity is usually drastically reduced, but where populations are sustained they may affect post-fire hydrology and erosion by providing pathways for infiltrating water, by translocating easily eroded soil to the surface and by

causing its direct downslope transference or bio-transfer (Dragovich and Morris, 2002).

3.3. Changes to the structure and hydrological properties of soil

Many authors have stressed the importance of soil properties affecting post-fire soil erosion rates (e.g. Giovannini et al., 1988; Sevink et al., 1989; Giovannini and Lucchesi, 1991; Imeson et al., 1992; Kutiel and Inbar, 1993). When soil is heated to 270–400 °C and sufficient oxygen is available, organic matter, which may include fine root mats, is combusted. In turn, this reduces the stability of soil structure and aggregates. Heating to above 460 °C drives off hydroxyl (OH) groups from clays and thus irreversibly disrupts their structure (DeBano et al., 1998). Although such high soil temperatures are not necessarily reached during burning, most researchers maintain that burning results in a more friable, less cohesive and more erodible soil (DeBano et al., 1998; Scott et al., 1998; Neary et al., 1999). The changes to the soil, however, depend on soil type and temperatures reached in the soil during burning (Guerrero et al., 2001). In some soils, the heat generated during a fire produces a new aggregation of particles by recrystallization of Fe and Al oxides (Giovannini et al., 1990), and if soil temperature remains below the destruction threshold for water repellency (see below), the wettability of aggregate surfaces may be reduced, resulting in increased aggregate stability (Giovannini and Lucchesi, 1983; Giovannini, 1994; Mataix-Solera and Doerr, 2004). The temperature thresholds at which soil property changes reportedly occur have been derived almost exclusively from controlled laboratory experiments and prescribed fires. Whilst these may well be applicable to wildfire conditions, our knowledge base of actual soil temperatures reached during wildfires, as opposed to prescribed fires, is limited because unlike flame temperatures, soil temperatures reached during burning are rarely measured directly and must therefore be inferred from post-fire indicators (e.g. soil colour; Ulery and Graham, 1993; Wondafraash et al., 2005).

The most frequently cited hydrological change to soil concerns its wettability. Although often characteristic of soils in long unburnt terrain under a wide variety of vegetation types (Doerr et al., 2000), fire can induce water repellency in non-repellent soil, and either enhance or reduce pre-existing surface repellency, depending on the amount and type of litter consumed and on the temperature reached (DeBano and Krammes, 1966; DeBano et al., 1970; Doerr et al.,

1996, 2004). Depending on its severity and persistence, water repellency can reduce or entirely prevent soil wetting for periods ranging from seconds to months (see review by Doerr et al., 2000). During burning, hydrophobic organic substances in the litter and topsoil become volatilized, creating a pressure gradient in this heated layer, which causes some material to escape to the atmosphere and some travelling downward until condensation occurs in cooler parts on or below the soil surface (DeBano et al., 1976). Apart from redistributing and concentrating hydrophobic substances in the soil, heat generated by fire is also thought to improve the bonding of these substances to soil particles (Savage et al., 1972) and make them more water-repellent by pyrolysis (Giovannini, 1994) and by conformational changes in the structural arrangement of the hydrophobic compounds (Doerr et al., 2005). Based on laboratory studies, water repellency is known to be intensified at temperatures of 175–270 °C, but destroyed above 270–400 °C, independent of soil type, but depending on heating duration (e.g. DeBano et al., 1976; Doerr et al., 2004). Where oxygen deficiency prevents the combustion of hydrophobic compounds, the temperature at which destruction of water repellency occurs may rise to 500–600 °C (Bryant et al., 2005). Based on laboratory examinations of critical shear stress for the initiation of erosion of different forest soils, similarities in temperature thresholds (if heated under air) for changes in critical shear stress and for water repellency have been found by Moody et al. (in press), suggesting that these two soil properties are linked. They concluded that this link may arise from similarities in the various types of cementation processes exhibited by cohesive mixed-grain soils at different temperatures.

As regards the longevity of fire-induced changes to water repellency, comparatively little is known because long-term post-fire monitoring is rare and such observations as have been made vary widely. For example, in coniferous forests in the USA, fire-induced water repellency has reportedly persisted for as long as 6 years (Dyrness, 1976) or as little as a few months (DeBano et al., 1976). In comparing the results of different studies on water repellency, however, caution is warranted as, owing to differences in methodology and the often substantial spatial and temporal variability in repellency, uncertainties in establishing the degree, spatial distribution and longevity of repellency may be considerable. More details concerning fire effects on water repellency and associated uncertainties are given in DeBano (2000a), Doerr et al. (2000), and Doerr and Moody (2004).

Although some soils underlain by permafrost may thaw deeply during fire, others may be relatively little affected immediately after fire (Brown, 1965; Swanson, 1996). More important than heating during the fire is the removal of ground-insulating organic matter, which can affect the long-term soil temperature regime and thus cause thickening of the active layer – the c. 1.5-m thick layer that melts each summer – leading to local subsidence and the formation of very uneven topography known as thermokarst (Swanson, 1981). This effect can be long-lasting. For example, Burn (1998) reported for discontinuous permafrost in southern Yukon Territory, a progressive increase over 38 years in the thickness of the active layer from 1.4 m before wildfire up to 3.8 m at the end of the period. This effect seems to be highly dependent on the amount of organic layer destroyed. Yoshikawa et al. (2003) suggest that if an organic layer >7–12 cm thick remains after wildfire, the thermal impact will be minimal. Some vigorous vegetation types regrowing after fire can actually lead to improved insulation and thus thinning rather than thickening of the active layer (Viereck, 1973).

4. Indirect effects of fire

The increase in hydrological and geomorphological activity following wildfire tends to occur during the ‘window of disturbance’ (Prosser and Williams, 1998), which begins immediately after burning. This period varies in length between locations (Wondzell and King, 2003) and can last from as little as a month to several years or more (e.g. Brown, 1972; Morris and Moses, 1987; Prosser, 1990; Walsh et al., 1992; Shakesby et al., 1994; DeBano et al., 1996; Shakesby, 2000). During this period, the changes brought about by wildfire can affect the hydrology of the burnt area by altering the patterns and quantities of infiltration and overland flow at the small scale and discharge and peakflows and corresponding impacts on channels at large scales. The changes often enhance soil erosion processes, aiding detachment by overland flow and rainsplash and causing accelerated hillslope soil redistribution. Much research effort has been aimed at assessing the quantities and rates of erosion by water during this disturbance phase. Attention has also been drawn to a range of other erosion processes such as landslides, debris flows and gravity sliding of dry particles. Hydrological and geomorphological effects are dealt within separate sub-sections here and consideration is given to current knowledge of the mechanisms, the magnitude of post-fire changes and resulting impacts on landforms.

4.1. Post-fire hydrological effects

4.1.1. Infiltration

The usually accepted view concerning the effect of wildfire on infiltration is its reduction relative to comparable unburnt areas (e.g. Swanson, 1981; Benavides-Solorio and MacDonald, 2001; Martin and Moody, 2001a; Wondzell and King, 2003). Differences in the amount of litter removed as a result of fires of different severity can have a marked effect on the proportions of rainfall available for overland flow. Under simulated rainfall applied at a rate of 62 mm h^{-1} on partially burnt subalpine range, in Ephraim catchment, Utah, Bailey and Copeland (1961) reported that, with a ground cover of vegetation and litter of 60–75%, only about 2% of rainfall contributed to overland flow. For a 37% ground cover, about 14% and, for a 10% cover, about 73% of the rainfall contributed to overland flow. Thus in low severity fires, where much of the litter and ground remain, rainfall storage and infiltration in the only partially destroyed or unaffected litter layer are typically higher than in areas affected by moderate and high fire severity (e.g. Rubio et al., 1997; see also Section 3.2). A number of other causes have been suggested including fusing of the soil surface and soil water repellency (see Section 3.3), the sealing of pores by fine soil and ash particles (Campbell et al., 1977; Morin and Benyamini, 1977; Wells et al., 1979; Lavee et al., 1995; Neary et al., 1999), development of a fungal crust (Lavee et al., 1995), and compaction by raindrop impact (Wells et al., 1979; see also Section 4.1.2). Wildfire may also cause lower infiltration in areas subject to seasonal freezing by increasing the likelihood of the occurrence of frozen soil because of the removal of the insulating organic matter (Campbell et al., 1977).

The most frequently cited change to the soil affecting infiltration is that of soil water repellency explained in Section 3.3. For example, DeBano (1971) found that the infiltration capacity of a water-repellent soil was 25 times lower than for a similar soil rendered hydrophilic by heating, and for the first 5 min of measurement a water-repellent soil exhibited only 1% of its potential infiltration capacity when non-repellent. For extremely water-repellent soils, wetting may not occur at all during rainstorms. This has, for example, been reported for some Portuguese forest soils during 40–46 mm of simulated rain in an hour (Walsh et al., 1998), despite such soils being able to exhibit infiltration capacities of c. 80 mm h^{-1} in the laboratory when rendered non-repellent (Doerr et al., 2003). Water repellency for some of these Portuguese soils has proved so persistent that samples in different

laboratory experiments remained dry beneath a ponded water layer for more than 3 weeks (Doerr and Thomas, 2000).

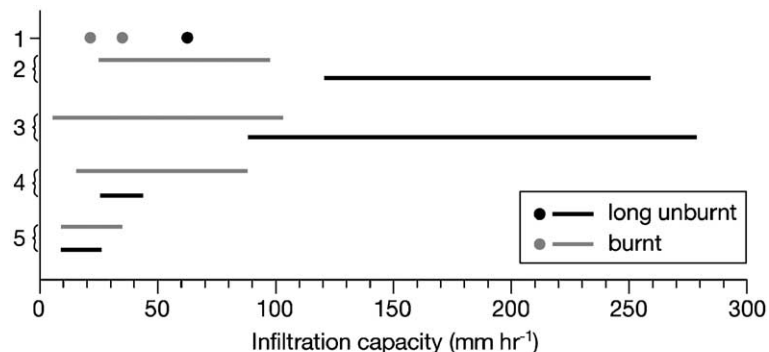
Although often present in long unburnt terrain, where it may be a hydrologically and geomorphologically less relevant characteristic (Shakesby et al., 2000), water repellency can become more prominent in post-fire conditions because of its intensification, changed position in the soil profile and the removal of vegetation and litter cover. Water repellency has thus been implicated as the main cause of reduced infiltration rates on burnt compared with unburnt terrain (Martin and Moody, 2001a), although its impact under field conditions may be difficult to separate from other post-fire changes such as alterations to the litter cover (e.g. Shakesby et al., 1993). Attempts to isolate its impacts using simulated rainfall over bare soils in eucalypt stands, involving surfactants to achieve wettable soil conditions, have shown that water repellency can reduce infiltration 16-fold compared to wettable conditions (Leighton-Boyce, 2002).

Water repellency is generally only fully expressed when soil moisture falls below a critical threshold (Doerr and Thomas, 2000; Dekker et al., 2001). Thus its effects on infiltration are most pronounced following dry antecedent conditions. This also means that burnt forest slopes more exposed to insolation can record lower infiltration capacities than their more shaded, and therefore less dry counterparts (Pierson et al., 2002).

Burning in general, and specifically the presence of post-fire soil water repellency, does not necessarily lead, however, to a marked reduction in infiltration (see, for example, Anderson et al., 1976). Imeson et al. (1992) found that the water-holding capacity of aggregates in moderately water-repellent soils was actually improved after wildfire in oak woodland in north-east Spain by the creation of many water-retaining pores. Only after several years, through compaction and reduced porosity, was infiltration capacity reduced. They concluded that there was little difference in infiltration behaviour between burnt and unburnt soils. Cerdà (1998) found a high infiltration capacity on wettable, dry, ash-covered soils immediately after fire resulting in negligible overland flow. Similar findings were reported by Calvo Cases and Cerda Bolinches (1994) in pine scrub near Valencia, eastern Spain. The examples of infiltration capacities following wildfires and in comparable unburnt terrain given in Fig. 2 illustrate this variation in response: in two out of the five studies, infiltration capacities from burnt and unburnt terrain are virtually indistinguishable. Predicting post-fire infiltration capacity is thus not straightforward (Doerr and Moody, 2004).

4.1.2. Overland flow

In contrast to undisturbed forested catchments where overland flow is typically rarely seen or observed, it is comparatively common in burnt forested terrain (Fig. 3). Overland flow can occur by saturation of the soil up



1. Ponderosa pine, Arizona (Campbell et al., 1977)
2. Ponderosa pine and mixed conifer, Washington (Martin and Moody, 2001a)
3. Pine and eucalypt, Portugal (Shakesby et al., 1993)
4. Pine and oak scrub, Israel (Kutiel et al., 1995)
5. Oak scrub, Spain (Imeson et al., 1992)

Fig. 2. Infiltration capacities measured after wildfire and on comparable unburnt terrain according to various authors. Lines represent ranges of values and points represent individual values.



Fig. 3. Overland flow transporting burnt soil, ash and charred debris during intense rain following wildfire in eucalypt forest in the Victorian Alps, south-east Australia in 2003 (photo courtesy of Rob Ferguson).

to the surface (saturation overland flow) or by water input rates exceeding infiltration capacities (Hortonian or infiltration-excess overland flow). The role of the former in post-fire enhanced geomorphological activity has received less attention than the latter. Reduced infiltration capacity of the soil and destruction of the vegetation and litter to varying degrees depending on fire severity tend to cause enhanced overland flow and runoff (Scott et al., 1998). For example, in burnt and unburnt dry eucalypt-dominated forest in New South Wales, Australia, Prosser (1990) found about 1.5–3 times more overland flow from a burnt compared with a long unburnt plot. He attributed this difference to fire creating at the plot scale a spatially more homogeneous and intense water repellency in the burnt soil. Wells (1981) reported seven times more overland flow in chaparral terrain on burnt compared with unburnt plots and Cerdà and Doerr (2005), using controlled simulated rainfall in Mediterranean pine forest, found that recovering vegetation 3 years after burning reduced overland flow to 18% of that measured on bare soils 6 months after the fire. Robichaud et al. (2000) maintained that if the vegetation and litter cover are reduced to <10%, overland flow can increase by more than 70%. Even more extreme differences in overland flow were found by Soler et al. (1994) for plots on 27–29° slopes in burnt oak scrub in the Catalan Coastal Ranges of Spain. During 18 months, overland flow was 14.6 times higher, even though the fire was of moderate rather than high severity, than on the unburnt control plot. In New South Wales, Australia, Prosser (1990) found markedly different effects for plots in locations of low and high fire severity, the former only producing increased overland flow in the first post-fire rainfall event. Vega and Díaz-Fierros (1987) reported that

after a severe fire in a *Pinus pinaster* stand, the overland flow coefficient (8.5%) was 21 times greater in the first year after fire but 11 times more for a moderate fire compared with an unburnt control (0.31%).

A clear link between overland flow response and fire severity is, however, not always present. Benavides-Solorio and MacDonald (2001) found only a slight difference in overland flow between rainfall simulation plots located in moderate and high fire severity areas and no significant difference between a moderate and low severity burn, even though there were large differences in sediment yield. Soto et al. (1994) found higher overland flow coefficients in the first year after fire for burnt compared with unburnt (3.0%) plots, but there was little difference between the two fire severities (low burn, 6.7%; moderate burn, 5.4%).

A number of authors have noted that higher overland flow coefficients occur following dry than wet periods, which are often attributed to the enhancement or development of soil water repellency (e.g. Shahlaee et al., 1991; Walsh et al., 1994; Soto and Díaz-Fierros, 1998). These increased amounts are generally thought to reflect Hortonian rather than saturation overland flow (Soto and Díaz-Fierros, 1998) because of the presence of water repellency at the soil surface or only a short depth below it, although strictly they may both operate where surface and/or accessible subsurface wettable soil is present (Doerr et al., 2000). In a eucalypt forest in Australia, Burch et al. (1989) reported a 3-fold increase in overland flow coefficient from 5% to 15% after drought. In burnt pine forest in South Africa, saturation overland flow in the surface wettable soil was implicated in explaining an increase in stormflow response to 7.5% compared with 2.2% on unburnt terrain (Scott and Van Wyk, 1990). Walsh et al.

(1994) found a 5–25% higher overland flow response on plots in burnt compared with unburnt eucalypt and pine forest, attributed in part to saturation overland flow resulting from a perched water table overlying dry water-repellent soil.

Some other effects of wildfire (e.g. removal of vegetation and litter cover, rainsplash compaction of the soil, inwash of fines into cracks and pores and development of a surface stone lag) have also been considered important in enhancing overland flow (White and Wells, 1982; Imeson et al., 1992; Shakesby et al., 1996; Walsh et al., 1998; Doerr et al., 2000). Wildfire can also affect snowmelt behaviour with implications for runoff characteristics, although there has been relatively little research into this topic. There is some evidence that, for relatively warm snowpacks, enhanced melting aided by lowering of snow albedo by surface dust from blackened trunks and logs and increased exposure from loss of overstorey (Tiedemann et al., 1979) may lead to greater input of meltwater into the soil compared with unburnt terrain (Swanson, 1981).

Most authors have reported that the impact of burning on overland flow is most acute in early rainfall events following fire, although these may need to be relatively intense to induce overland flow. For example, Soto et al. (1994) noted appreciable overland flow on plots of *Ulex europaeus* scrub only when more than 20 mm had fallen in 24 h. Prosser and Williams (1998) installed plots on eucalypt woodland, south-eastern Australia affected by fire of moderate severity and found that only the largest rainfall event, 68 mm in 24 h, produced substantial overland flow. Both the 24-h and 30-min intensities of the event were considered to have a recurrence interval of about 1 year.

4.1.3. Catchment runoff behaviour

Hydrological responses to fire at the catchment scale have received less attention than at smaller scales largely because of the greater difficulties of installing and maintaining instruments, the long recovery period or 'relaxation time' at this scale (Brunsden and Thornes, 1979), the greater spatial heterogeneity of environmental factors such as geology, topography, vegetation and soils as well as fire extent over large catchments, fire severity and fire history (Miller et al., 2003). The size of the post-fire increases in hydrological response can be up to two orders of magnitude of difference (e.g. Anderson et al., 1976; Scott, 1993), with increases in total flow being often considerably less than those in peak discharge.

Contrasts between pre- and post-fire flow tend to be most marked in humid ecosystems with high pre-fire evapotranspiration (Anderson et al., 1976; Robichaud

et al., 2000). Berndt (1971) observed immediate increases in baseflow following wildfire in north-central Washington. Campbell et al. (1977) found that runoff was eight times greater on a small (8.1 ha) severely burnt catchment of ponderosa pine than on a larger (17.7 ha) unburnt one in the first autumn rains following the Rattle Burn of May 1972 in north-central Arizona, USA. In the following year, water yields from moderately and severely burnt catchments were respectively 3.1 and 3.8 times greater than from the unburnt catchment. In Entiat catchment, Washington, USA, comprising ponderosa pine and juniper forest, a 42% increase in water yield was recorded during the first year after a wildfire (Helvey, 1980). In contrast, the Tillamook Burn in 1933 in Oregon increased total annual flow of two catchments in the first year by only 9% (Anderson et al., 1976). Similarly, Scott (1997) found only a 12% increase in total flow after a wildfire in a pine-afforested catchment in South Africa.

Whilst Scott (1997) found little difference in total flow, he found a 3- to 4-fold increase in stormflow in the first year and still a 2-fold increase in the second year after wildfire. Post-fire peak discharges tend to be more sensitive hydrological responses to wildfire than changes in total flow (Moody and Martin, 2001a), and because of their potentially damaging effects and easier measurement, more attention has been paid to them than to other hydrological changes. On burnt catchments following a storm, the response magnitude tends to be greater and the response time shorter compared with unburnt terrain. Post-fire peak discharge increases tend to be most pronounced when short-duration, high-intensity rainfall of comparatively small volume occurs on steep, severely burnt catchments with shallow, skeletal, water-repellent soils (Robichaud et al., 2000; Neary et al., 2003). The Tillamook Burn in 1933 in Oregon increased the annual peak discharge by about 45% in the first year after the fire in the Trask and Wilson River catchments (370 and 412 km², respectively) relative to the adjacent, slightly burnt Siletz catchment (518 km²) (Anderson et al., 1976). Croft and Marston (1950) found that 5-min rainfall intensities of 213 and 235 mm h⁻¹ led to peakflows in newly burnt catchments that were five times larger than those from adjacent unburnt areas in northern Utah. Neary et al. (2003) describe a 15-min rainstorm with an intensity of 67 mm h⁻¹ after the Coon Creek Fire of 2000 in Arizona, which led to a peakflow more than seven times greater than any peakflow recorded over the previous 40 years. Glendening et al. (1961) reported peakflow increases following wildfire in Arizona chaparral of as much as 450-fold. Peakflows have been found to differ between summer and winter.

For example, for a chaparral site in Arizona, summer peakflows increased 5- to 15-fold, whereas there was no change for winter peakflows, probably reflecting less intense rainfall and repellency (Robichaud et al., 2000). On the other hand, in spring and early summer peakflows would be expected to be increased through enhanced snowmelt, and whilst there are reports of increased peakflows, decreases of up to 50% have also been noted (Anderson et al., 1976). Fire severity can have a substantial effect on peakflow. For ponderosa pine in Arizona, Campbell et al. (1977) found that whereas a wildfire of moderate severity increased peakflow response up to 23-fold, a high severity wildfire increased it nearly 200-fold relative to undisturbed terrain (Table 2). Neary et al. (2003) describe special circumstances that can amplify post-fire peak discharges. Sequential failure of debris dams comprising large woody debris in and adjacent to streams can increase peak discharges by up to 3 orders of magnitude.

The timing of peakflow tends to change following wildfires (Helvey, 1980). Burnt catchments generally respond to rainfall fast, producing more ‘flash floods’ than in comparable unburnt catchments or in the same catchment prior to burning. Water-repellent bare soils and loss of plant and litter cover tend to cause the flood peak to arrive faster as well as being higher (Scott, 1993, 1997; Neary et al., 2003). There may also be more than one peak. For the Yarrangobilly catchment in south-eastern New South Wales, Australia, Brown (1972) noted a sharp peak early in the flood hydrograph, which occurred following rainfall only during the post-fire period. It was followed several hours later by a more rounded one characteristic of those on pre-fire hydrographs (Fig. 4). The most likely cause of the sharp peak is Hortonian overland flow enhanced by litter destruction lowering the surface storage capacity of rainfall and reducing surface roughness together with the water-repellent nature of the soil in the eucalypt forest. Such peaks still occurred, though in increasingly attenuated form 4–5 years after fire.

The temporal decline in post-fire peakflows is typical, although the time taken to return to pre-fire levels varies. Moody and Martin (2001a) noted a decline in peakflow during a 3-year post-fire period in the 26.8-km² Spring Creek catchment in the Front Range of the Rocky Mountains, Colorado, USA vegetated mainly with ponderosa pine and Douglas fir. In 1997, following fire in May 1996, peak discharge dropped from 6.6 m³ s⁻¹ km⁻² after 30 mm of rainfall at an intensity of 50 mm h⁻¹ to only 0.11 m³ s⁻¹ km⁻² by 2000 after a rainstorm of similar size and intensity. Recovery to pre-fire peakflows can, however, take many decades (Robichaud et al., 2000). As vegetation, litter and soil begin to return to their pre-burn state, elevated post-fire streamflow amounts characteristic of the immediate post-fire period decline (e.g. Campbell et al., 1977). The ‘flashy’ character of peakflows declines too but may take longer to return to pre-fire levels (Anderson et al., 1976; Scott and Van Wyk, 1990; Scott, 1997).

4.2. Post-fire geomorphological effects

4.2.1. Soil erosion by water

4.2.1.1. Water erosion processes. The detachment of soil particles by rainsplash or overland flow and their transfer downslope are very sensitive to the kinds of modifications to land surface properties caused by fire (Johansen et al., 2001). In particular, general reductions in the vegetation cover and especially the ground vegetation and litter leave the soil prone to raindrop impact and reduce the opportunities for rainfall storage so that erosive overland flow tends to occur more readily. Many researchers view this as the most important factor leading to increased post-fire erosion (e.g. White and Wells, 1979; Wells, 1981; Dieckmann et al., 1992; Inbar et al., 1998). These effects are generally thought to be related to fire severity as this reflects the amount of destruction of ground cover and affects important

Table 2

The highest peakflows recorded in three water years (1972–1975) in unburnt and burnt Rattle Burn ponderosa pine catchments, Arizona following wildfire in early May 1972 (modified from Campbell et al., 1977)

Unburnt catchment (17.7 ha)		Moderately burnt catchment (4.0 ha)		Severely burnt catchment (8.1 ha)	
Date	Peakflow (m ³ s ⁻¹ ha ⁻¹)	Date	Peakflow (m ³ s ⁻¹ ha ⁻¹)	Date	Peakflow (m ³ s ⁻¹ ha ⁻¹)
(Summer only) ^a	0	2 Sept. 1972	0.002	2 Sept. 1972	0.038 ^b
17 Oct. 1972	0.0007	19 Oct. 1972	0.003	17 Oct. 1972	0.038 ^b
7 July 1974	0.0001	7 July 1974	0.003	7 July 1974	0.019
27 Oct. 1974	0.0002	16 July 1975	0.0009	14 July 1975	0.005

^a For direct comparison with the two burnt catchments, the peakflow is for only the summer period following the fire in May 1972.

^b Flow exceeded capacity of the flume.

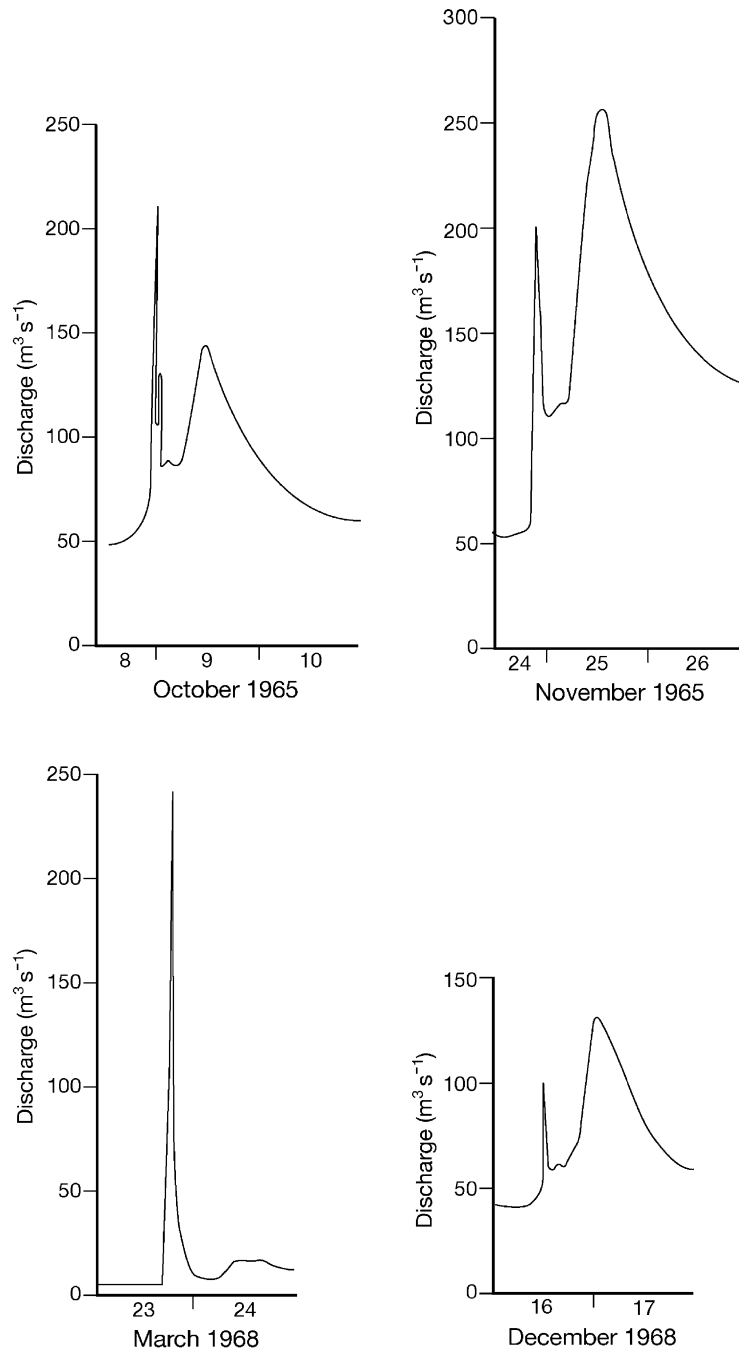


Fig. 4. Typical hydrographs for selected post-fire storms in the Yarrangobilly catchment, New South Wales, Australia following fire in March 1965 (modified from Brown, 1972).

soil properties such as aggregate stability and water repellency (see Section 3.3). After fire, there is usually a ready supply of highly erodible fine ash and charcoal on the soil surface. As described in Section 4.1.1, infiltration capacity of the underlying soil is often reduced after fire, which tends to increase overland flow and thus the likelihood of erosion.

There are a number of additional factors that can affect the process of post-fire soil erosion by water including slope angle and aspect, soil thickness and its spatial variation, spatial distribution of soil thickness, climate (especially rainfall amounts and intensities) and fire severity. For example, slope aspect may influence the amount of burnt plant debris and the rate

of vegetation recovery (Marqués and Mora, 1992) together with the rate of drying out of soil, which in turn can affect its repellency characteristics (Dekker et al., 2001; Pierson et al., 2002) and the amount of soil lost (Andreu et al., 2001). Selective removal of the more easily eroded fine sediment can leave the eroded soil with an increasingly coarse texture, which makes it more resistant to further erosion (Thomas et al., 1999). Long-term redistribution of soil cover within a catchment through repeated fire-induced soil erosion events can also affect catchment sediment yields (Nishimune et al., 2003).

Interception of rainfall by burnt tree surfaces tends to lead to increased sizes of water drops, which often fall on a bare ground surface, thereby tending to enhance the process of rainsplash detachment of soil particles (McNabb and Swanson, 1990). On slopes, rainsplash erosion of fine sediment made available from the fire-induced break-up of aggregates and plentiful ash may take place. Erosion by this process has probably been underrated in many locations because the results can be mistaken for those of sheet erosion. Whereas rainsplash erosion is, however, generally uniform across a slope, sheetwash erosion is more variable in its effect. Where soils are stony, the efficacy of rainsplash erosion is reduced as fines are removed and more stones revealed. Rainsplash can limit erosion by compacting the soil (Meyer and Wells, 1997) and sealing the soil surface by clogging soil pores (Swanson, 1981). On poorly aggregated water-repellent soils, however, rainsplash erosion can be particularly effective. Terry and Shakesby (1993) carried out laboratory tests showing that simulated raindrops impacting on water-repellent soil produced fewer, slower-moving ejection droplets, which moved shorter distances than those from drops impacting on corresponding wettable soils, but which carried larger sediment quantities with each droplet. Under repeated simulated raindrops, the wettable soil surface became sealed and compacted increasing its resistance to splash detachment, whereas the repellent soil surface remained dry and non-cohesive making the soil particles easily displaced by splash even when a repelled surface water film developed. These laboratory results have been replicated using simulated rainfall on repellent, long unburnt soil surfaces in the laboratory (Doerr et al., 2003) and in the field (Leighton-Boyce, 2002), but as yet they have not been verified for natural rainfall on bare, newly burnt surfaces (Shakesby et al., 2000).

Soil particles on bare post-fire soil surfaces can be detached by overland flow as well as by rainsplash. Overland flow is also the main process by which ma-

terial is transported downslope, as sheetflow or as concentrated flow where overland flow is directed by the surface form along specific paths. Wondzell and King (2003, p.77) argue that this is the “dominant mechanism of surface erosion after wildfire”.

4.2.1.2. Sediment redistribution by overland flow and runoff and resulting landforms. Where there is little overland flow despite intense rainfall, post-fire soils of low cohesion can be susceptible to net downslope transfer of soil by rainsplash alone. Where such soils contain stones and/or substantial roots, stone- or root-capped pedestals may develop (e.g. Germanoski and Miller, 1995; Shakesby et al., 2003, in press). Loss of rainfall storage capacity on the vegetation and in the litter combined with, where present, the water repellency of the soil, promote sheetwash, and the development of rills and gullies in some post-fire environments. Sediment transport is often by sheetwash where water flow is not concentrated into small channels (Meeuwig, 1970). Where water does become concentrated it can form extensive rill systems particularly on steep and water-repellent slopes and can develop in response to the smallest of flow concentrations (e.g. Doehring, 1968; Wells, 1981, 1987; Wells et al., 1987) (Fig. 5). According to DeBano (2000a,b), rill formation follows well defined stages on water-repellent burnt soil. During rainfall, the wettable surface soil layer becomes saturated leading to increased pore pressures, decreased shear strength and ultimately failure. It begins to slide downslope and a miniature debris flow develops (e.g. Wells, 1981; Gabet, 2003). Water in the wettable soil layer adjacent to the debris flow, being no longer confined, flows into the debris flow channel and erodes and entrains sediment from the water-repellent and subsequently from the underlying wettable soil. The downcutting process is thought to be self-limiting because once rill downcutting reaches wettable soil, infiltration increases causing flow to diminish.

The importance of fire-induced water repellency in rill production is illustrated by experiments carried out by Osborn et al. (1964). Using surfactant to suppress repellency on some of a series of plots, rills developed only on those plots left untreated (i.e. water-repellent). The concentration of flow necessary to produce rills can occur naturally where slope form tends to focus sheetwash sufficiently or where anthropogenic modifications such as roads and paths lead to concentrated flow (e.g. Scott and Van Wyk, 1990). Following post-fire rain, Atkinson (1984) observed rills up to 60 cm deep developed along walking trails near Sydney. Zierholz et al. (1995) investigating a later fire in the



Fig. 5. Rills formed after the Buffalo Creek Fire in May 1996, south-west of Denver, Colorado, USA (photo courtesy of John Moody).

same region also remarked on the tendency for rills to occur along trails.

Rill networks do not, however, always seem to develop on repellent burnt soils for a variety of reasons including interception of overland flow by cracks, burnt-out root holes and animal and insect burrows (Burch et al., 1989; Booker et al., 1993; Shakesby et al., 2003, *in press*), the rapid recovery from fire-induced reductions in infiltration rates, high antecedent soil moisture and rapid vegetation regrowth (Wondzell and King, 2003). Rill development can also occur on wettable post-fire soils, indicating that repellency is not a pre-requisite for rill erosion.

Given that two key factors leading to gully formation on hillslopes are a reduced protective ground cover and increased overland flow and runoff (Prosser and Slade, 1994), a large literature on post-fire hydrological and geomorphological effects in which gullies feature might be anticipated even though they are rare in undisturbed forest ecosystems (Heede, 1975). Although there are exceptions (e.g. Doehring, 1968; Brown, 1972; Good, 1973; Germanoski and Miller, 1995; Benda et al., 2003; Istanbuluoglu et al., 2002, 2003), the post-fire formation of gullies is relatively infrequently mentioned. Thus, for example, in reviews by authors in the USA on post-fire hydrology and erosion (e.g. Wells et al., 1979; Swanson, 1981; Robichaud et al., 2000), reference is made to rill systems, sheetwash and mass movement but not to gullies. Rather than gullies being formed after fire, it has been argued that gully excavation was limited (Brown, 1972), restricted to enlargement of pre-existing forms (Leitch et al., 1983) or that they were virtually unaffected by post-fire erosion instead acting as conduits for sediment eroded from upslope (Atkinson, 1984; Sha-

kesby et al., 2003). A probable reason for the infrequent reference to post-fire gullies in many cases is simply a matter of semantics, there being no precise size limits to distinguish small gullies from rills, which are by contrast frequently reported. Other possible reasons may be the thinness as well as better cohesion of forest soils (particularly in Europe), their relatively high stone content in steep terrain (McNabb and Swanson, 1990), the absence of soil-loosening tillage (except for forest plantations) which is a major factor in promoting gullies on agricultural land, sediment ‘exhaustion’ in frequently burnt and eroded terrain, or the spatial heterogeneity of runoff-generating and sink zones on burnt hillslopes (Kutiel et al., 1995) so that there is insufficient volume of concentrated overland flow.

Channel systems respond in a complex way to the destruction of vegetation and litter and alterations to soil properties caused by wildfire. Reported responses range from channel aggradation, braiding and the creation of alluvial fans and boulder deposits to headward expansion of channel system, entrenchment, terrace development and channel narrowing (e.g. Laird and Harvey, 1986; McNabb and Swanson, 1990; Florsheim et al., 1991; Germanoski and Harvey, 1993; Meyer and Wells, 1997; Moody and Martin, 2001b; Benda et al., 2003; Legleiter et al., 2003). Variations in sediment supply from tributaries to trunk streams modify the larger fluvial systems and these modifications are transmitted back through the tributaries (White and Wells, 1982). In a burnt chaparral catchment in California, Keller et al. (1997) found that for low-order streams two extreme conditions operated in the first two moderately intense post-fire storms. After the first storm, there were transport-limited conditions resulting in

channel deposition whilst, in the second, sediment-limited conditions had already become established leading to scouring in low-order streams and to considerable downstream sediment transport. Just over 2 years later, most post-fire eroded sediment reaching the stream system had moved some distance beyond the limits of the burnt area where it led to an enhanced flooding risk because of reduced channel capacity.

There are exceptions to this view of rapid transfer of soil through the hillslope-channel system after fire. In a semi-arid burnt catchment in Nevada, Germanoski and Miller (1995) found that there was little export of sediment from hillslopes but substantial entrenchment and resulting erosion of the channels occurred in the first 2 years after fire. Moody and Martin (2001b) had similar findings: they found that approximately 20% of sediment was eroded from hillslopes and 80% from channels in the highly decomposed granite, ponderosa pine and Douglas fir catchment of Buffalo Creek in the Colorado Front Range. They suggested that sediment transport was probably invariably transport-limited after fire but supply-limited before it. Furthermore, they found that post-fire eroded sediment had an extremely long residence time of several hundred years (see Section 5 for further discussion). Benda et al. (2003), investigating post-fire channel environments of the Boise River, Idaho following intense thunderstorms, focused on newly aggraded and enlarged alluvial fans encroaching on channels. They found that the fans led to downstream knickpoints being developed causing an increase in channel gradient on their downstream sides but a decrease in gradient on their upstream sides for up to 4 km. Wide floodplains, side channels and proto-terrace construction were associated with the increased sediment storage near the aggraded fans.

Legleiter et al. (2003) investigated fluvial responses some 12–13 years after the 1988 fires in Yellowstone National Park, USA. They found that the finer-grained material was transported from burnt hillslopes by sheet-flow and rill networks to the channel network over a period of 5–10 years, but that subsequently the still elevated post-fire discharges and decreasing sediment supply had induced a period of channel incision leading to the development of an armoured channel bed layer. This represents, however, an unusual study in its relatively large-scale and long-term observations of fire effects and generally “the fluvial knowledge base remains meager” (Legleiter et al., 2003, p.134).

4.2.1.3. Quantities and rates of water erosion. As regards geomorphic impacts, most wildfire-related literature concerns soil erosion, commonly that by water,

and its visual and quantitative effects. (Rates of erosion involving processes other than the removal of particles by water are considered in Section 4.2.2.) This section is concerned with erosion estimated from natural post-fire rainfall and the reader is referred to Johansen et al. (2001) for a summary of work concerning erosion by simulated rainfall. Whilst soil erosion modelling (e.g. Soto and Diaz-Fierros, 1998; Moody and Martin, 2001b; Miller et al., 2003), which is not dealt with here, has been used to estimate post-fire erosion rates at the hillslope scale, this review considers the measurement of post-fire soil erosion by water involving field monitoring which includes instrumenting soil losses, use of tracers, remeasuring ground level changes and collecting sediment eroded from unbounded slope areas in a variety of sediment traps (Fig. 6) and small bounded plots. Examination of the ‘genealogy’ of erosion rates given in published summary tables and reviews has revealed errors. In the construction of summary tables in this review, therefore, the original sources have been consulted as much as possible to try to avoid the perpetuation of such errors.

There have been reports of erosion amounts that were small relative to other post-fire disturbance effects (e.g. Noble and Lundeen, 1971; Kutiel, 1994; Shakesby et al., 1996, 2000), and net losses from hillslopes have been reported as lower than might have been anticipated (e.g. Germanoski and Miller, 1995; Prosser and Williams, 1998; Shakesby et al., 2003). Others have even reported no signs of post-fire erosion despite high fire temperatures or, exceptionally, slight decreases in runoff and erosion have been recorded (Naveh, 1973; Kutiel and Inbar, 1993; Cerdà, 1998). The typical view, however, is that fire increases overland flow and soil losses on burnt hillslopes relative to undisturbed forested land (e.g. Ahlgren and Ahlgren, 1960; Helvey, 1980; Swanson, 1981; Scott and Van Wyk, 1990; Soto et al., 1994; Soler and Sala, 1992; Zierholz et al., 1995; Robichaud and Brown, 1999; Moody and Martin, 2001b) and also that the amounts tend to be at least in broad terms positively correlated with fire severity (e.g. Prosser and Williams, 1998).

Assessing the relative importance of wildfire as a geomorphological agent and comparing the impacts between locations has proved difficult for the following reasons:

- (1) To assess the impact of fire on erosion, post-fire erosion rates are often quoted as being tens, hundreds or even thousands of times higher than those recorded for comparable undisturbed forest areas. Post-fire erosion of catastrophic proportions is often implied. In some cases, erosion



Fig. 6. Eroded sediment accumulated upslope of a silt fence installed following the Bitterroot Fire in 2000 in Bitterroot National Forest, Montana, USA (photo courtesy of Kevin Spiegel).

may well be serious with respect to medium- to long-term soil degradation, but since, for most mature forests, erosion is negligible or virtually non-existent, this way of expressing the seriousness of post-fire erosion can be misleading. For example, [Inbar et al. \(1998\)](#) expressed post-fire erosion rates that were 100,000 times higher in burnt than in unburnt Mediterranean forest, but in fact they represented a loss of just 1 mm depth of soil, which gives a different geomorphological perspective. (It should be pointed out, however, that the authors considered that even this modest amount might be important from a soil fertility perspective on these thin, degraded soils.) More useful, ideally, would be comparison of overall post-fire erosion amounts with some evaluation of the seriousness of soil loss for a particular environment or a long-term erosion rate from the sedimentary record, but this is rarely done, mainly because of the considerable problems and uncertainties with these approaches. Most so-called tolerable soil loss rates are developed for agricultural land ([Schertz, 1983](#)), which are not appropriate for disturbed forested land. An example of a possibly appropriate tolerable soil loss figure for such environments is that determined for artificially disturbed terrain soil systems by the US Environmental Protection Agency (EPA) which recommends a value of $4.5 \text{ t ha}^{-1} \text{ yr}^{-1}$ ([US EPA, 1989](#)). This value, however, is almost twice the estimated annual renewal rate, for example, for a forested soil in Iowa, USA of $2.7 \text{ t ha}^{-1} \text{ yr}^{-1}$ ([Glanz, 1995](#) cited in [Renschler and Harbor,](#)

[2002](#)), which might be a more useful guide of a critical threshold for tolerable soil loss.

- (2) Erosion rates determined from point measurements (e.g. erosion pins), short slope transects, bounded plots, 'open' plots and entire catchments are often indiscriminately quoted and compared even though there is an approximately inverse relationship between erosion amounts and the scale of measurement ([Scott et al., 1998](#)).
- (3) An appreciation of pre-fire hillslope and channel conditions and the effects of post-fire storm-by-storm data on erosion for a number of years is clearly critical for an improved understanding of wildfire impacts on soil erosion. For a number of logistical reasons, however, it is difficult to install monitoring equipment prior to the first post-fire erosive rainfall and continue to monitor erosion over a period of a year or even longer. As a result, the number of studies with uninterrupted erosion records of this length of time starting in the immediate post-fire period prior to any rainfall is actually quite small.
- (4) Many studies focus on the effects of severe fire followed by large or extreme rainfall events and whilst there is a considerable amount of research dealing with prescribed fire, there is relatively little concerning the effects of moderate to low severity wildfire on erosion ([Wondzell and King, 2003](#)).
- (5) Despite the availability of aids for converting units used to express erosion rates (e.g. [Foster et al., 1981](#)), conversion errors have not been eliminated from the literature.

- (6) Important background information (e.g. slope angles, rainfall characteristics, fire severity) is often not given.

With these problems in mind, a selection of published post-fire hillslope erosion rates, all expressed in terms of tonnes per hectare from wildfire-prone areas in North America, Africa, Australia and Europe subject mainly to fire of high severity is presented in Table 3. Where measurements have been derived from ground-level change measurements and volumes rather than weights of sediment, a standard assumed value for bulk density of 1.0 g cm^{-3} has been used to estimate weight losses, unless the author(s) have suggested a value, in which case it is used. The data are arranged according to the scales and types of measurement: slope transects, ranging from centimetres to metre scale, to bounded plots usually m^2 to tens of m^2 in extent and to tracer measurements and sediment traps of various types (in which the estimated contributing areas are usually measured up to hundreds of m^2 in extent so that an erosion rate per unit area can be derived).

Generally, as would be expected, measured hillslope erosion rates in Table 3 in the first year after fire show a negative correlation with the size of the area considered, with values being higher for slope transects ($27\text{--}414 \text{ t ha}^{-1}$) than for bounded plots ($0.5\text{--}197 \text{ t ha}^{-1}$), which in turn are higher than tracer measurements or sediment traps collecting sediment from relatively large unbounded plots draining into sediment traps ($0.1\text{--}70 \text{ t ha}^{-1}$). Almost certainly, slope transects for ground level change measurements are selected to reflect erosion rather than deposition. Consequently, erosion rates derived from such transects using an estimated bulk density and extrapolated in terms of losses per hectare are higher than those from areally determined soil losses. Since bounded plots are usually placed on planar slopes specifically to control for the effects of micro-topographical variations and to facilitate comparison, sediment storage, which would tend to reduce erosion rates, is limited, although this effect is countered to some extent because plots normally exclude erosive overland flow from upslope. For sediment traps collecting sediment from unbounded areas, their usually much larger contributing areas mean that there is more spatial heterogeneity in terms of infiltration capacity and surface roughness (Kutiel et al., 1995; Pierson et al., 2002) and hence more opportunities for storage, and generally lower erosion rates (Imeson et al., 1992; Pradas et al., 1994; Inbar et al., 1998), as overland flow infiltrates otherwise often water-repellent soil via preferential flow pathways, such as less repellent zones (Dekker et al., 2001), abandoned faunal burrows

(e.g. Booker et al., 1993) and along live roots and burnt-out root holes (e.g. Burch et al., 1989; Ferreira et al., 1997, 2000). Soil redistribution figures derived from comparatively large unbounded areas draining into traps probably provide the best insights into hillslope post-fire losses. Significantly, they tend to suggest mostly relatively modest erosion at this scale. Indeed, the first-year erosion rate for most studies in Table 3 is less than the estimated annual forest soil renewal rate of $2.7 \text{ t ha}^{-1} \text{ yr}^{-1}$ quoted earlier. Thus, whilst there are reports of high soil erosion at the hillslope scale where conditions are particularly conducive (e.g. Lavee et al., 1995), the results in Table 3 support the finding that soil redistribution tends to be high at small scales (point and plot scales) but comparatively low at hillslope and catchment scales (e.g. Imeson et al., 1992; Kutiel and Inbar, 1993; Shakesby et al., 2003).

As regards erosion rates beyond the first year after wildfire, there have been relatively few studies monitoring hillslope erosion continuously for several years following wildfire (Benavides-Solorio and MacDonald, 2001), but these suggest that post-fire erosion during the 'window of disturbance' takes the form of a peak lasting 1–2 years followed by a decline of varying steepness towards pre-fire conditions (e.g. Helvey, 1980; Robichaud and Waldrop, 1994; Inbar et al., 1998) (Fig. 7). During the peak, sediment redistribution tends to be transport-limited (Moody and Martin, 2001b), later becoming supply-limited as easily eroded sediment has been removed and there is greater protection of the surface from an increasing cover of vegetation and litter (Connaughton, 1935; Grigal and McColl, 1975; Megahan and Molitor, 1975; Wells et al., 1979; White and Wells, 1982; Shakesby et al., 1993), and in some locations from litter dam development (e.g. Mitchell and Humphreys, 1987; Diaz-Fierros et al., 1994; Zierholz et al., 1995) and an increasing concentration of stones at the surface in stony soils (Morris and Moses, 1987; Calvo Cases and Cerda Bolinches, 1994; Shakesby et al., 1994, 2002; Legleiter et al., 2003). For example, Paton et al. (1995) in south-east Australia, Inbar et al. (1997) in Israel and Robichaud and Brown (1999) in eastern Oregon, USA (Fig. 8a–c; Table 3) reported declining erosion rates over the first few years after fire such that, after c. 3–4 years, they approached those measured in undisturbed terrain. A similar decline was deduced by Shakesby et al. (1994) for north-central Portugal using space–time substitution to reconstruct the post-fire erosion curve from measurements on erosion plots in pine and eucalypt plantations burnt at different times (Fig. 8d). In the Colorado Front Range, some studies (e.g. Morris and Moses, 1987; Martin and Moody, 2001b) have shown that soil erosion

Table 3

A selection of published post-fire measured rates of hillslope erosion by water based on measurements from slope transects, different types of bounded plots, tracer studies and sediment traps collecting sediment from unbounded areas. Erosion rates relate to the first year after fire unless otherwise stated. Most bounded plots, but not troughs and traps, involved monitoring overland flow amounts as well as sediment

Location	Vegetation	Rainfall ^a (mm)	Fire severity ^b	Slope (°)	Post-fire erosion rate (t ha ⁻¹)	Unburnt erosion rate (t ha ⁻¹)	Notes	Author(s)
<i>(A) Ground height changes</i>								
Arizona, USA	Mixed conifer	n.d.	High	23 33 38	80 226 414	n.d.	Based on steel tape profiles and a bulk density of 1.6 g cm ⁻³	Hendricks and Johnson (1944)
Oregon, USA	Douglas fir	836 ^b	High	18–40	300	n.d.	Steel tape profiles	Sartz (1953)
North-central Portugal	Eucalypt and pine	~800	Medium–high	3–40	27–104	0.005–0.02	Erosion bridge ^c and 16 m ² plots. Measured bulk density of 1.0 g cm ⁻³ used	Shakesby et al. (1994, 2002)
New South Wales, Australia	Eucalypt	953	Medium–high	8	118	n.d.	Erosion bridge ^c . Measured bulk density of 1.0 g cm ⁻³ for surface soil used	Shakesby et al. (in press)
Colorado Front Range, USA	Ponderosa pine and Douglas fir	440	High	n.d.	68 (s.-facing) 41 (n.-facing)	0.28 0.28	Based on stone pedestals and a bulk density of 1.7 g cm ⁻³	Moody and Martin (2001b)
New Mexico, USA	Mixed conifer	825	High	n.d.	46 ^d	n.d.	Erosion pins	White and Wells (1982)
Arizona, USA	Ponderosa pine	240	High	n.d.	49 ^d	n.d.	Remeasurement of >1500 points on grass-seeded soil	Rich (1962)
<i>(B) Bounded plots</i>								
San Gabriel Mts, California	Chaparral	773	High	n.d.	197	n.d.	36 m ² plots	Krammes and Osborn (1969)
Galicia, Spain	Pine plantation	1400	n.d.	3–29	15–170	n.d.	20 m ² plots	Diaz-Fierros et al. (1982, 1987)
Galicia, Spain	Gorse scrub	n.d.	n.d.	17	13	0.7	80 m ² plots	Soto et al. (1994), Soto and Diaz-Fierros (1998)
Western Cape Province, South Africa	Pine plantation	~1500	High	16–32	10–26	n.d.	54 m ² plots	Scott and Van Wyk (1990), Scott et al. (1998)
New South Wales, Australia	Eucalypt	~710	Medium	n.d.	2.5–8.0	n.d.	8 m ² plots	Blong et al. (1982)
North-central Portugal	Eucalypt and pine	678 1472	Medium	19–22	0.5–2.2 (yr 1) 3.2–6.6 (yr 2)	n.d.	16 m ² plots	Shakesby et al. (1996)

San Gabriel Mts, California	Chaparral	559 820 1468 495	n.d.	26	19.1 ^d (yr 1) 2.3 ^d (yr 2) 0.7 ^d (yr 3) 0.01 ^d (yr 4)	0.01	80 m ² plots	Wells (1981)	
<i>(C) Tracer studies</i>									
Colorado, USA	Lodgepole pine and spruce-fir	n.d.	High	n.d.	63 ^d	n.d.	Use of ¹³⁴ Cs and fluorescent dye as particle markers to determine soil particle movement	Striffler and Mogren (1971)	
<i>(D) Sediment traps</i>									
New Mexico, USA	Mixed conifer	825	High	12	70	n.d.	Sediment trap collecting from ~92 m ²	White and Wells (1982)	
North-east Spain	Pine and scrub	675	High	25	3 (n.-facing)	0.01–0.22	Collecting troughs for 200 m ² areas	Marqués and Mora (1992)	
Oregon, USA	Mixed conifer	n.d.	High	31	22 (s.-facing) 2.5	n.d.	Silt fences collecting from areas of 152–636 m ²	Robichaud and Brown (1999)	
Idaho, USA	Felled and burnt mixed conifer	n.d.	Medium–high	17 11 27 (mean)	2.2 ^c 1.1 1.3 ^d	0.6	Collecting troughs for 0.97 km ²	Megahan and Molitor (1975)	
Washington, USA	Mixed conifer	n.d.	High	19–26	0–1.8 (yr 1) 0–3.6 (yr 2)	n.d.	Collecting troughs draining 26–357m ²	Radek (1996)	
Mt Carmel, Israel	Oak and pine	~700	High	n.d.	0.9–3.7 (yr 1) 0.1–0.3 (yr 2) 0.001–0.0008 (yr 3) 0.001–0.0005 (yr 4)	0.00001–0.0006	Collecting troughs for ~200 m ² areas	Inbar et al. (1997)	

^a Average annual rainfall in upright font; total rainfall for periods of less than a year over which soil losses were recorded in italic font.

^b An arbitrary classification of fire severity interpreted from authors' own descriptions.

^c See Shakesby (1993).

^d A bulk density of 1.0 g cm⁻³ is assumed for this paper.

^e Plots half the length of the others.

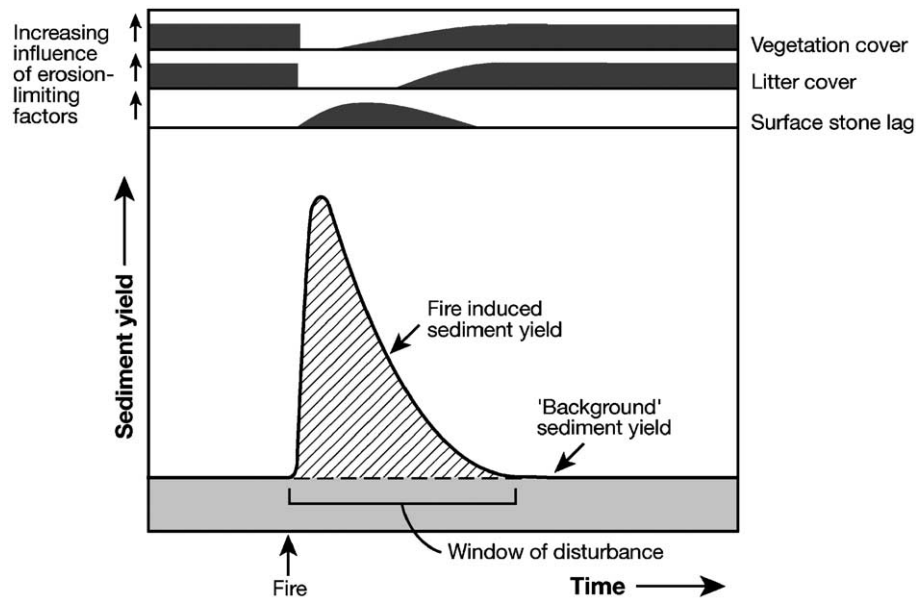


Fig. 7. The hypothetical decline in sediment yield after wildfire and the role of three factors (vegetation cover, litter cover and stone lag development) in reducing erosion rates (based on Swanson, 1981; Shakesby et al., 1994; Prosser and Williams, 1998).

rates return to undisturbed values 3–4 years after burning, and in general this relaxation time is usually thought to last 3–10 years after burning (Rowe et al., 1954; Doehring, 1968; Scott and Williams, 1978; Tiedemann et al., 1979; Helvey, 1980; Wells et al., 1979; Robichaud et al., 2000), although periods as short as 3 months (Rich, 1962) and as long as 14 years have been reported (DeBano et al., 1996). This decline, however, can be altered in response to an uneven pattern of rainfall such that the erosion peak may be delayed (e.g. Shakesby et al., 1996; Cerdà, 1998).

Individual high-intensity rainstorms can account for appreciable quantities of post-fire erosion. The following examples illustrate this effect. In the Monserrat area of south-east Spain, two rainfall events of 170–183 mm and 35–37 mm accounted for 99% of total post-fire erosion over 16 months (Marqués and Mora, 1992). In eucalypt forest near Sydney, Australia, Atkinson (1984) found the equivalent of a year's loss of soil after one rainfall event of 16.5 mm lasting 45 min, Leitch et al. (1983) estimated a loss of 22 t ha⁻¹ after 21 mm of rain on small plots in burnt eucalypt-dominated forest in the Victorian Central Highlands of Australia and Blong et al. (1982) found that during just 2 days with rainfall intensities of > 30 mm h⁻¹, 40% of the first year's erosion occurred on a 4 m × 2 m plot on a 12° slope in the northern suburbs of Sydney. Cannon et al. (2001b), in burnt terrain near Los Alamos, New Mexico, USA, measured hillslope sediment yield and runoff concentrations of 0.23–0.81 kg L⁻¹ in response to maximum 30-min rainfall intensities

of 31 mm h⁻¹. In Bitterroot Valley, Montana, USA, Robichaud (2002) reported relatively high soil losses of 2–40 t ha⁻¹ following short-duration, high-intensity rainfall with maximum 10-min intensities of 75 mm h⁻¹, compared with losses of only 0.01 t ha⁻¹ following long-duration, low-intensity rainfall.

With some exceptions (e.g. Prosser, 1990; Shakesby et al., 2003), few studies have focused on monitoring hillslope soil losses following wildfires of different severity. Some, however, have incorporated a chance wildfire into a field assessment of the effects of prescribed fire on erosion. For example, Wohlgemuth (2003) compared soil losses collected in sediment traps in chaparral before and after wildfire and prescribed fire. He found that post-fire sediment fluxes for the latter were only a quarter to a third of the former.

At the catchment scale, assessing wildfire-induced increases in sediment yields has proved more difficult than at the hillslope scale for the following reasons. First, the wildfire may only burn part of the catchment or, if affecting the whole of it, it may destroy the instruments installed for long-term sediment yield monitoring. Second, if instrumentation of burnt and unburnt catchments of similar characteristics immediately after fire proves feasible, it is difficult to be sure that they were comparable in an unburnt state. Some examples of post-fire sediment yields for different sized catchments are given in Table 4. The greater capacity for storage of sediment at the catchment- compared with the hillslope-scale is a major factor in causing sediment yields recorded at

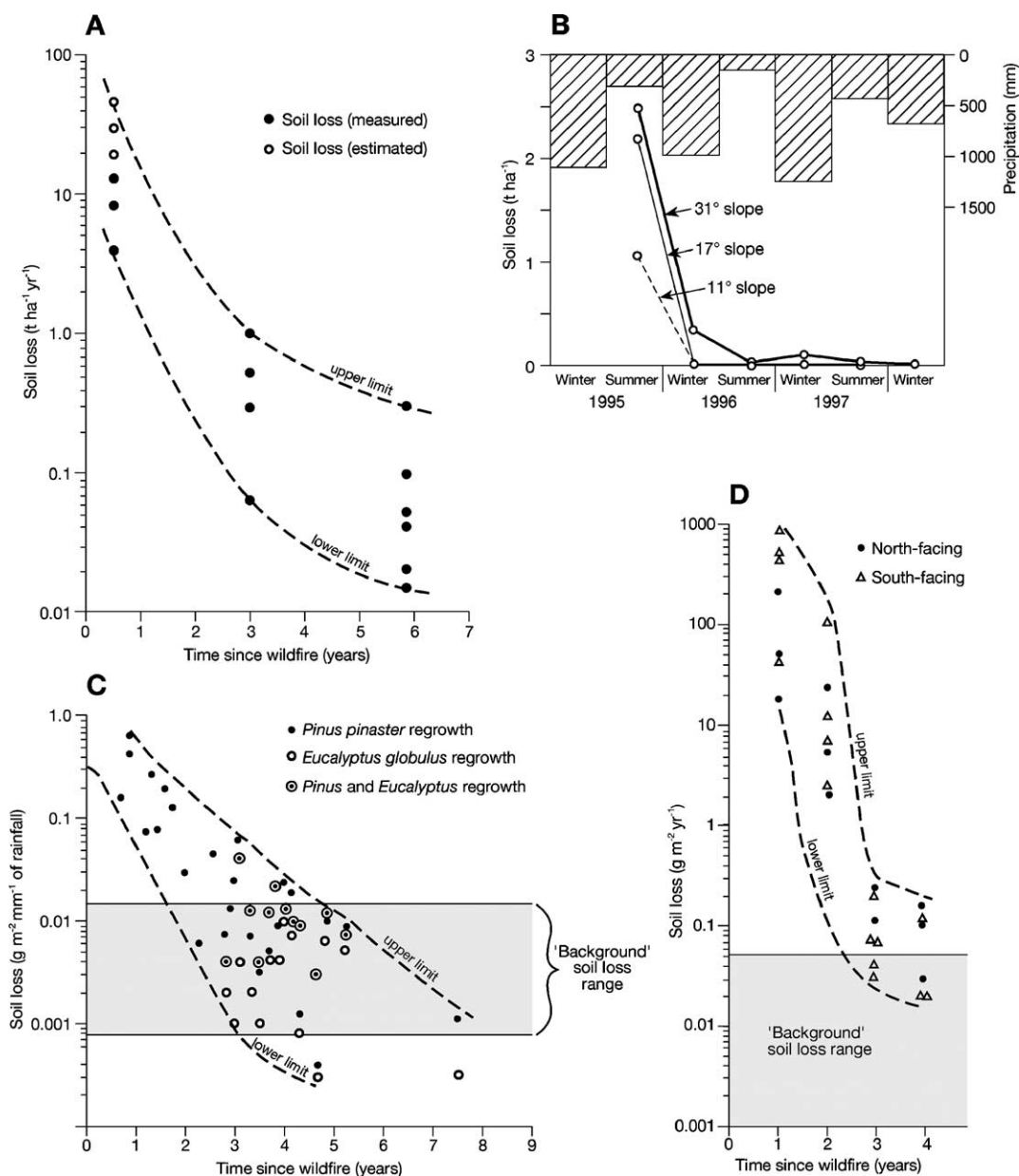


Fig. 8. Examples of the post-fire decline in soil loss. A) New South Wales, Australia (modified from Paton et al., 1995); B) mixed conifer, Eastern Oregon, USA (modified from Robichaud and Brown, 1999) (note arithmetic scale for soil loss, inclusion of rainfall amounts and short timescale compared with other examples); C) pine and eucalypt plantations, north-central Portugal (modified from Shakesby et al., 1994); and D) pine and oak scrub, Mount Carmel, Israel (data from Inbar et al., 1997).

gauging stations to be lower, sometimes considerably lower than those measured, for example, on hillslope erosion plots or estimated from slope transects (Kutiel and Inbar, 1993; Osterkamp and Toy, 1997; Prosser and Williams, 1998). In one of the few documented series of studies of simultaneous nested measurements of erosion in wildfire-affected terrain at both scales, Scott and Van Wyk (1990) found for a steep, pine-afforested 200-ha catchment in South Africa that the measured rate of

sediment delivery at the catchment gauging station of 7.83 t ha⁻¹ during the first year after fire was as much as half that recorded on small hillslope plots over the same period. This may be exceptionally high for a catchment of this order of size and the figure of <1 t ha⁻¹ as reported by Helvey (1980) for a larger catchment (Table 4) may be more typical. Defining what may be typical even for a specific catchment with its set of distinctive climatic, geological, geomorphic, pedologi-

Table 4

A selection of published post-fire measured rates of sediment yields for catchments. Sediment yields relate to the first year after fire unless otherwise indicated

Location	Vegetation	Rainfall ^a (mm)	Fire Severity ^b	Slope ^c (°)	Area (ha)	Post-fire sediment yield (t ha ⁻¹)	Unburnt sediment yield (t ha ⁻¹)	Notes	Author(s)
Burnt Mesa, New Mexico, USA	Mixed conifer	825	High	5	0.41	47 ^d	n.d.	–	White and Wells (1982)
			Moderate	7	0.08	10	–		
			Light	5	0.14	32	–		
Arizona, USA	Ponderosa pine	737	High	n.d.	8.1	4.8 ^c	0.003	Bedload and suspended sediment	Campbell et al. (1977)
			Moderate	n.d.	4.0	0.005 ^c			
Washington, USA	Mixed conifer	580	High	16	514	0.12	0.008–0.100	–	Helvey (1980)
				15	563	0.26			
				15	473	0.40			
Western Cape Province, South Africa	Pine plantation	1296	High	7	200	7.83	n.d.	Bedload and suspended sediment	Scott et al. (1998)

^a Average annual rainfall in upright font, rainfall for measured period in italics.

^b An arbitrary classification of fire severity based on authors' descriptions.

^c Channel or catchment slope.

^d A bulk density of 1.0 g cm⁻³ is assumed for this paper.

^e Measured over 6 months.

cal and floral characteristics, however, is difficult not least because short-term monitoring may miss infrequent, large events when most geomorphological impact occurs (Wondzell and King, 2003). Higher sediment yields (10–47 t ha⁻¹) like those reported by White and Wells (1982) (Table 4) relate to very small catchments (<0.5 ha) and illustrate the effect of catchment size on post-fire sediment delivery and thus more opportunities for sediment storage in large catchments (Schumm, 1977; Wells, 1981; Osterkamp and Toy, 1997).

Although the delivery ratio of hillslope to catchment sediment yields has been infrequently measured, there have been a number of reports of increased post-fire sediment yields and sediment concentrations in catchments compared with pre-fire yields in the same catchment or a nearby undisturbed catchment judged to be comparable in the post-fire period (e.g. Rich, 1962; Storey et al., 1964; Campbell et al., 1977; Scott and Williams, 1978; Wells et al., 1979; Burgess et al., 1981). The highest sediment concentrations after fire were 335 times higher than for a similar pre-fire discharge for the 227-km² Yarrangobilly catchment in south-eastern Australia (Brown, 1972). In the 98-ha Dog Valley experimental station near Reno, Nevada, Copeland (1965) recorded 9.6 t ha⁻¹ of sediment yield in a single exceptional post-fire storm, whereas in a nearby unburnt catchment there was scarcely any sediment eroded in

the same storm. Scott and Williams (1978) reported a loss of 22.8 t ha⁻¹ for an 857-ha burnt catchment in the San Gabriel Mountains, California following a 10-day storm with a 24-h maximum rainfall of 267 mm and a recurrence interval of >100 years.

Wildfire can cause raised sediment yields at the catchment scale for some considerable time after fire where conditions are suitable. This is exemplified by burnt chaparral in the western USA where sediment supply by dry ravel processes (see Section 4.2.2.1) to channels may be large even when post-fire vegetation recovery is well advanced. Helvey (1980) reported an 8- to 10-fold increase in sediment yields compared with pre-fire conditions 7 years after fire for small catchments c. 5–6 km² in extent. He also explored the relationship between sediment concentration and discharge and found that sediment concentration at a given flow rate was higher during the rising stage of the hydrograph than during the falling stage. He concluded that during high flow conditions large quantities of fresh, loose sediment became available for transport, but during low-flow conditions the availability of such sediment and the carrying capacity of the stream were both reduced (Fig. 9).

Long-term storage of sediment in and near channels can also help to sustain high post-fire sediment yields. For example, Rice (1974) estimated for a burnt chap-

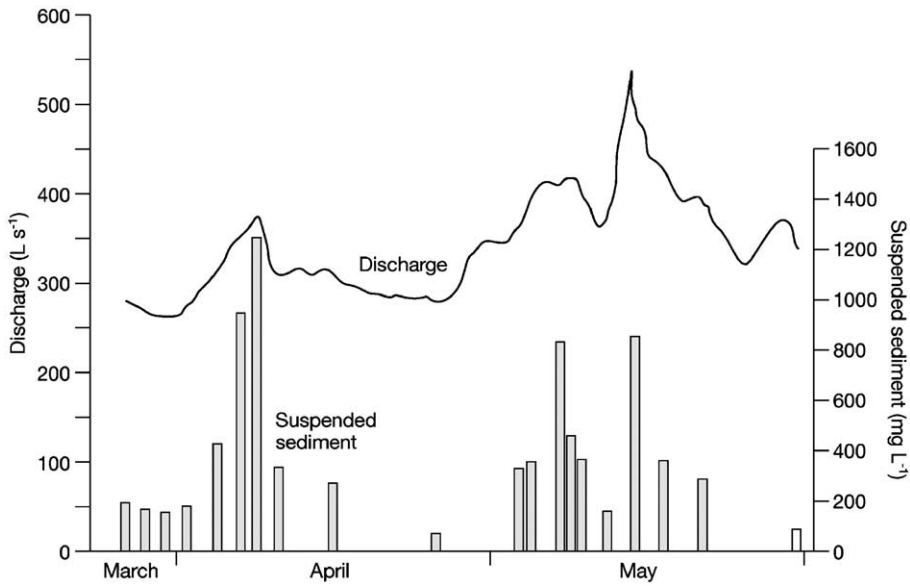


Fig. 9. The relationship between post-fire sediment concentration and discharge for the 56.3 km² Burns catchment, north-central Washington, USA during snowmelt runoff in 1972 (modified from Helvey, 1980).

arral catchment in California that 74% of the sediment came directly from scour of residual sediment in the channel with 22% from rills and gullies and only very small quantities from wind, dry ravel and landslides, although the latter two processes were regarded as the long-term source of the channel sediment. The unusual nature of the hillslope processes (particularly dry ravel)

in such highly erodible terrain must be borne in mind, but high yields also occur in other environments. For example, relatively high post-fire sediment losses were also recorded in suspended sediment yields by Scott and Van Wyk (1990) for a steep 200-ha pine catchment in South Africa (Fig. 10) subject to high-severity fire which induced soil water repellency. The change in the

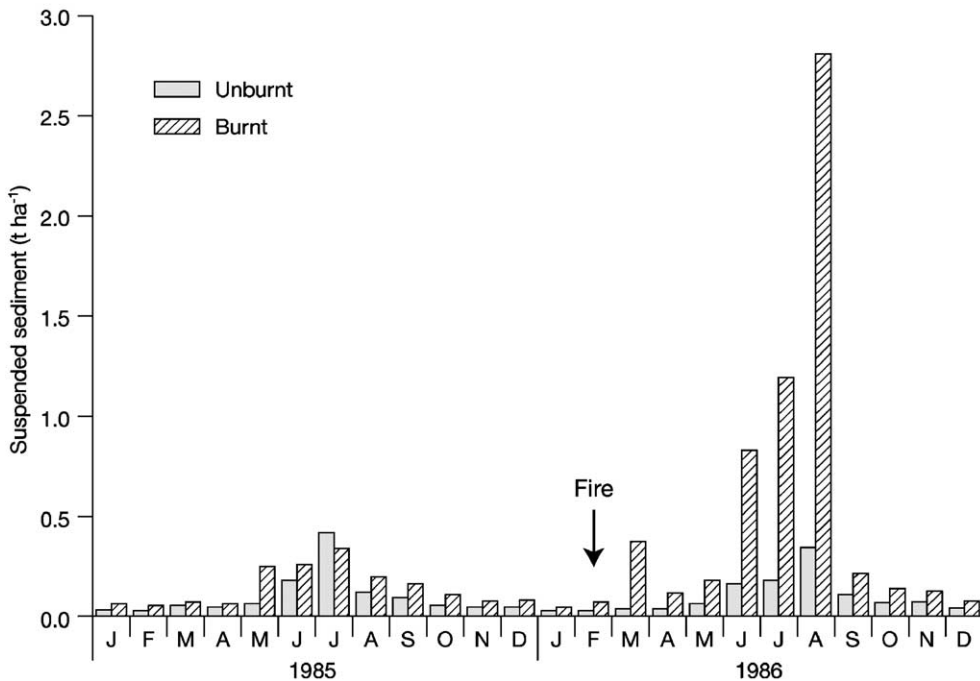


Fig. 10. Monthly suspended sediment yields for paired catchments (Bosboukloof, 201 ha, burnt; Lambrechtsbos-B, 65 ha, unburnt), Western Cape, South Africa before and after a high-severity wildfire in February 1986 (modified from Scott and Van Wyk, 1990).

suspended sediment characteristics after fire between the control and the burnt catchment in Fig. 10 is clear with losses in the latter in August 1986 alone approaching 3 t ha^{-1} , more than five times as much as recorded in the former. Similarly, Brown (1972) found peak suspended sediment concentrations of 143,000 ppm after severe wildfire in eucalypt-dominated forest in the catchment drained by Wallace's Creek in south-east Australia, which represented a 20-fold increase over the highest values recorded before the fire.

There have been some attempts to consider the contribution of fire-induced accelerated erosion to long-term sediment yields in fire-prone terrain. For example, Swanson (1981) compared the long-term fire impacts of chaparral scrub and Douglas fir — western hemlock forest on catchments in western USA. He estimated for the former a 30-fold increase in sediment yield in the first year after fire, recovery over 10 years and a fire frequency of 25 years (Fig. 11). He suggested that fire-induced sediment would represent 70% of the long-term sediment yield. In contrast, for the latter he estimated no more than a 5-fold post-fire increase in sediment yield, a recovery period of 20–30 years and a 200-year fire return period. In this case, fire-induced sediment yield would represent only about 25% of the long-term figure. These fire frequencies seem to be short by comparison with long-term records in western USA, which indicate recurrence intervals of 500–1000 years (Meyer et al., 1995; Rhodes and Davies, 1995), but they may prove appropriate for estimating future fire impacts, *even in these areas*, because of possibly increased fire frequency caused by changing climate and future management strategies (FAO, 2001).

4.2.2. Mass movement processes

Although most literature on post-fire erosion focuses on detachment and transport of sediment by water, increased soil redistribution by a range of mass move-

ment processes and agents can also be important. These include dry ravel, debris flow, other mass movement processes, wind erosion and bio-transfer of sediment (Wells et al., 1979; Swanson, 1981; Dragovich and Morris, 2002).

4.2.2.1. Dry ravel. Dry ravel (or ravel) is a post-fire erosion process widely referred to in review literature as well as in individual research reports, but to the authors' knowledge, it has only been reported from the western USA. It refers to the rapid dry particle-to-particle sliding of debris (derived mainly from weathered bedrock but also including organic matter) under the force of gravity (Florsheim et al., 1991). It tends to occur more readily on steeper slopes. For example, Mersereau and Dyrness (1972) found that rates of movement were four times more on 38° compared with 31° slopes. Because drying out of the soil is important, the process also preferentially occurs on slopes receiving the greatest insolation (i.e. south-facing in the northern hemisphere) (Wondzell and King, 2003). Most post-fire erosion directly or indirectly depends on water, but accelerated post-fire erosion by this process (and by wind) may occur during as well as after fire, although only the latter has been monitored. In chaparral areas in California and Arizona, it occurs on slopes underlain by granitic or coarse-grained sandstone rocks. These areas experience Mediterranean climate with relatively low annual rainfall totals and pronounced dry seasons when dry ravel usually occurs, although DeBano et al. (1979) also observed it occurring between storms in the wet season. Without fire, erosion by this process is relatively high on steep chaparral slopes (Scott and Williams, 1978; Wells, 1986) with half or more of hillslope erosion estimated to occur by this process (Krammes, 1965; Rice, 1974). Anderson et al. (1959) found pre-fire losses of as much as $224\text{--}4300 \text{ kg ha}^{-1}$ from dry ravel, rates that reportedly increase

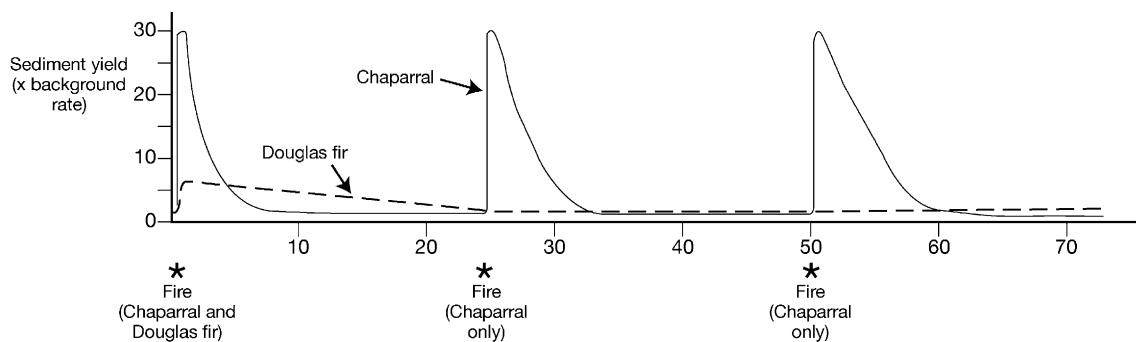


Fig. 11. Idealised variation in sediment yield during several fire rotations for a steep-land chaparral and Douglas fir western Cascade mountain system in the USA (modified from Swanson, 1981). The vertical scale shows by how many times the estimated sediment yield is raised above the background (i.e. pre-fire) rate. Fire return periods are shorter and post-fire sediment yields higher for chaparral than for Douglas fir.

9- (Krammes, 1960) to 13-fold (Rice, 1982) after fire. Two-thirds of the total amount eroded in the first year after fire in the Oregon Coast Range occurred in the first 24 h after fire (Bennett, 1982). In unburnt chaparral terrain, much of the mobilised sediment is stored upslope of vegetation stems and litter, but following their destruction by fire, it tends to move relatively short distances (Wondzell and King, 2003) and to accumulate in irregularities on hillslopes and in channels during dry conditions (Fig. 12) from where it tends to be transported downstream later in wet conditions (DeBano and Conrad, 1976; Florsheim et al., 1991). Since vegetation is slow to establish on slopes with high rates of ravel, erosion by this process can continue for more than a decade (McNabb and Swanson, 1990), although where vegetation recovery has been rapid, rates of movement can be reduced to near zero by the second growing season (Mersereau and Dyrness, 1972).

4.2.2.2. Debris flows. Although described in earlier literature (e.g. Mersereau and Dyrness, 1972; DeBano et al., 1977; Swanson, 1981; Wells et al., 1987), de-

tailed consideration of fire-related debris flows largely dates from the 1990s and with some exceptions (e.g. Conedera et al., 2003) relates mostly to the USA (e.g. Meyer and Wells, 1997; Cannon et al., 1998, 2001a,b; Cannon, 2001; Meyer et al., 2001; Cannon and Gartner, 2005) (Figs 13 and 14). In burnt areas, large-scale debris flows can be initiated through a process of progressive accumulation of sediment (so-called ‘progressive sediment bulking’) by overland flow and rill erosion in steep upper basin slopes, followed by deep incision on lower slopes (Meyer and Wells, 1997), which also leads to sediment entrainment (Parrett, 1987). Meyer and Wells (1997), working in Yellowstone National Park, further suggested that addition of fine-grained post-fire surface sediment eroded from hillslopes to the generally coarser-grained channel material was important both in developing debris-flow conditions and in maintaining flow mobility. On Storm King Mountain, Colorado, Cannon et al. (2001a) found that although soil-slip scars did form on post-fire slopes, their volumetric contribution to debris flows was relatively small, and that progressive sediment entrainment was the primary source of the

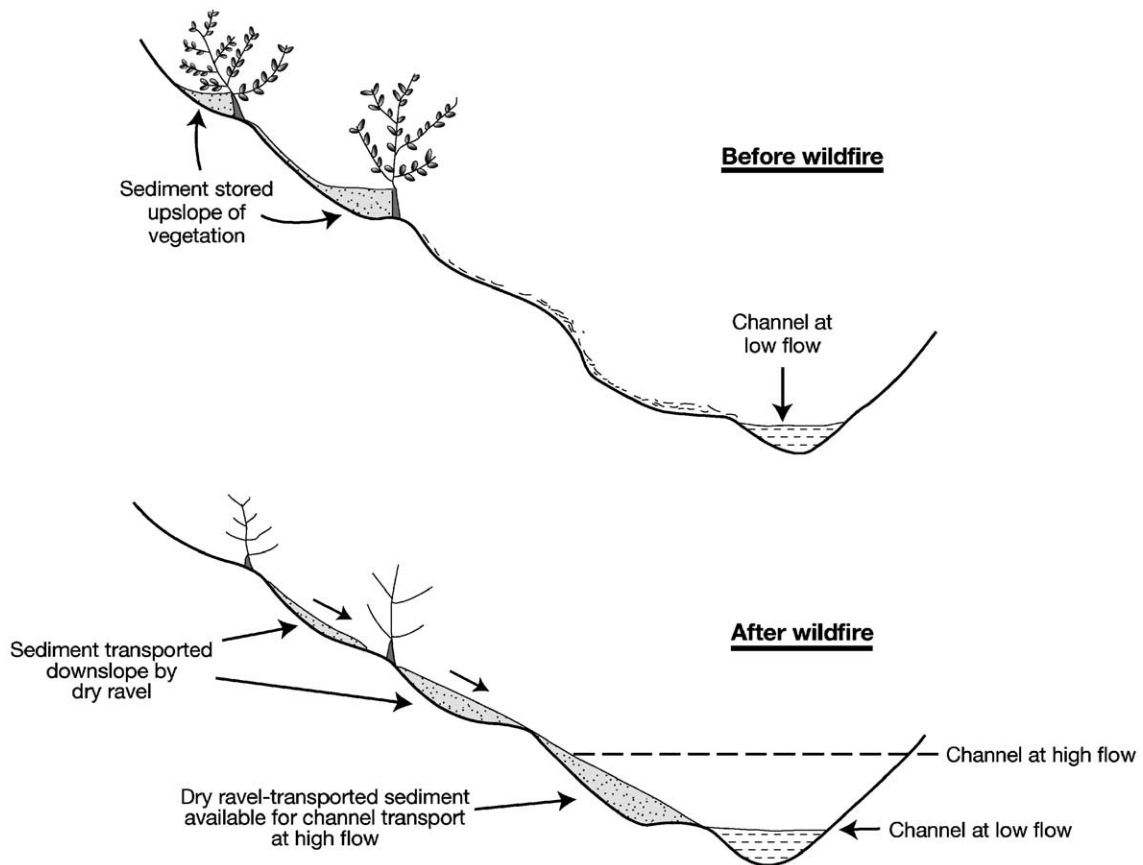


Fig. 12. Hillslope profile showing the effect of wildfire on vegetation and downslope transfer of sediment to stream channel by dry ravel (modified from Florsheim et al., 1991).



Fig. 13. Debris-flow track, Sula Complex, Bitterroot, Montana, USA (photo courtesy of Susan Cannon).

material. Debris flows generated through this mechanism most frequently occur in response to short-recurrence, short-duration rainstorms. Two other mechanisms of debris-flow initiation have been proposed. Debris flows may develop after fire downslope of shallow landslides during heavy rainfall and are referred to as soil-slip debris flows (Campbell, 1975). An alternative debris-flow initiation process particularly favoured by post-fire soil conditions has also been proposed (Johnson, 1984; Wells, 1987). Wells (1987) suggested that failure of a saturated layer of wetttable soil only a few millimetres thick overlying a subsurface water-repellent zone was important. Material from these tiny soil-slips formed rills and moved downslope as shallow, narrow debris flows in the manner described by Gabet (2003).

Debris flow magnitude and frequency have been examined in relation to wildfire. Wells et al. (1987) observed numerous small debris flows in the first-order basin in Santa Ynez Mountains, California burnt in 1985. According to Florsheim et al. (1991), these debris flows were triggered by fire and probably acted as important means of delivering to streams colluvial material built up by dry ravel in steep, low-order basins. Others have also considered that small-scale debris flows were triggered by fire (Wells, 1987; Weirich, 1989). On the other hand, large debris flows in burnt areas may be unrelated to fire events or require particular conditions. Florsheim et al. (1991), considering chaparral fire in Ventura County, California, noted a discrepancy between the frequency of fires and large debris flows: fire occurred every c. 30–65 years, where-



Fig. 14. Aerial view of post-fire debris-flow track and fan, Kuskonook, British Columbia, Canada (photo courtesy of Susan Cannon).

as the recurrence interval of large debris flows is “at least hundreds of years” (p. 510) according to radiometric evidence. Clearly fire was not the main factor responsible and the authors speculated about earthquakes as possible triggers in this tectonically active region. In granitic mountains of central Idaho, USA, Meyer et al. (2001) compared short-term measurements and long-term estimates of sediment yields by examining alluvial stratigraphy and the measurements by other researchers. In the period 7600–6800 years BP, there were frequent fires but modest sedimentation rates similar to those of today. Elevated Holocene average sediment yields, however, imply episodes of accelerated yield. They hypothesised that elevated debris-flow and flood activity caused by intense summer thunderstorms or spring storms with rapid snowmelt in post-fire periods following large, stand-replacing fires must have occurred.

There have been few attempts to estimate the rates of erosion per unit area by post-fire debris flow. From 34 debris flow-producing basins burnt at least over 50% of their areas in the western USA listed by Gartner et al. (2004), however, the average and median catchment-wide amounts of lowering were respectively 23 and 15 mm, which represent 230 and 150 t ha⁻¹ if conservatively a bulk density of 1.0 g cm⁻³ is assumed. These figures have been derived from basins <0.5 km² to 20 km² in area and seem to indicate substantial post-fire erosion by debris flow when compared with the water erosion figures given in Tables 3 and 4. It should be borne in mind, however, that debris flow generation requires substantial water erosion and the material is derived from various processes including hillslope overland flow and erosion, concentrated flow and flooding in channels as well as debris flow (S.H. Cannon, personal communication).

Although the examples discussed are debris flows on recently burnt terrain, fire is not always viewed as the primary trigger. For example, Cannon (2001) examined recently burnt basins in Colorado, New Mexico and California and found that following storm rainfall debris flows were present in only 37 of 95 basins examined. She suggested that specific geomorphological and geological conditions could indicate a susceptibility to post-fire debris flow. She also found that large-calibre debris flows produced from basins underlain by sedimentary rocks and gravel-dominated flows from decomposed granite occurred in basins with non-water-repellent soils, whereas gravel-dominated flows required repellency. Thus, although a factor determining the type of debris flow, water repellency was not critical to the process. The same conclusion about the lack of signifi-

cance of repellency in causing debris flows has been noted elsewhere (e.g. Meyer and Wells, 1997; Cannon and Reneau, 2000; Cannon et al., 2001a).

4.2.2.3. Shallow landslides. These can be important post-fire erosion features. On steep chaparral slopes they may account for about half of the erosion (Rice, 1974; DeBano et al., 1979; Robichaud et al., 2000). They occur relatively infrequently being dependent on high-magnitude, low-frequency rainfall events. Consequently, the importance of landsliding as an erosive process has reportedly tended to be underestimated and until recently relatively few studies were concerned with its measurement (cf. Rice et al., 1969; Rice and Foggin, 1971). For chaparral fires, DeBano et al. (1979) argued that post-fire soil loss is at first dominated by water erosion. For any landslide-prone area, reduced infiltration caused by soil water repellency and the binding action of any pre-fire root systems delay the process until some time after fire, typically when the soil becomes saturated during heavy storms and snowmelt, when these root systems have decayed and before new ones have become well established (e.g. Reneau and Dietrich, 1987; Gray and Megahan, 1981; Meyer and Pierce, 2003). In Payette River basin, Idaho, landslides were commonest 4–10 years after burning (Gray and Megahan, 1981). This decay of the root system may take 20 years (Meyer et al., 2001), so that slopes can remain susceptible to landsliding for a considerable period after fire. Where fire-induced channel scour and erosion occur they may cause removal or steepening of channel banks possibly reactivating the toe zones of pre-existing landslides (Wondzell and King, 2003). Rice (1974) estimated that the volume of sediment eroded by landslides on a chaparral area burned 9 years earlier was more than 18 times that on a comparable area unburned for 50 years.

4.2.2.4. Other mass movement processes. Soil creep can be enhanced by increased soil wetness following removal of litter and vegetation. Given that fire affects slopes to relatively shallow depths, however, large, deep-seated (>2 m depth to the failure surface) mass-movement forms, such as slumps and earthflows, are not usually thought to be directly related to fire (Swanson, 1981). In cold climates, increased susceptibility to freeze–thaw action, enhanced snow accumulation and snowmelt following wildfire can all cause enhanced erosion. Consumption of surface organic matter and litter by fire can leave the minerogenic soil more prone to freeze–thaw processes than under pre-fire conditions (Swanson, 1981). These processes can pro-

vide highly erodible material for subsequent removal by overland flow. White and Wells (1982) reported a 3- to 8-fold increase in water erosion over a 3-month period following upward movement of fine soil particles induced by freeze–thaw. Increased freeze–thaw activity can also lead to downslope frost-creep of soil, which involves the ratchet-like downslope movement of soil in frost-affected areas. This process tends to be most effective in temperate climates because of the higher frequency of freeze–thaw cycles compared with colder climates. On steep slopes, burning may cause increased snow avalanche activity (Winterbottom, 1974) in response to reduced anchoring of snow to the slopes by vegetation and altered snow accumulation and melting patterns in avalanche-initiation areas (Swanson, 1981). Such avalanches may entrain soil and rocks and scour and uproot trees (McNabb and Swanson, 1990).

4.2.3. Wind erosion

Erosion by wind in forested areas is usually rare (Wondzell and King, 2003), but burnt matter is highly susceptible to transport by this process. There have been some observations but few measurements. For example, Atkinson (1984), describing the effects of the 1983 wildfire in Royal National Park south of Sydney, found that wind had caused ash and sand to be built up on the windward side of obstacles to a depth of 8 cm. Blong et al. (1982) noted the downslope transport mostly of scorched leaves by wind on erosion plots. Wind-blown sand draped over an uneven burnt soil in exposed ridge-top locations on recently burnt terrain some 5 months after the December 2001 fires near Sydney was reported by Shakesby et al. (2003) and Whicker et al. (2002) undertook measurements near Carlsbad, New Mexico and found that erosion was 3 times greater over several months and 70 times greater during strong winds for recently burned compared with unburned sites. Regional wind and the convective forces generated during wildfire doubtless cause considerable redistribution of material but understandably it has not been measured.

4.2.4. Bioturbation and bio-transfer

Bioturbation can lead both directly and indirectly to appreciable downslope transfer of sediment after forest fire. Post-fire activity by, for example, small mammals, termites, lyrebirds and ants can all contribute to enhanced erosion in the sandstone terrain of south-eastern Australia. Ant mounding alone can provide an excess of highly erodible material for slopewash by a factor of 2–30 times (Humphreys and Mitchell, 1983). Dragovich and Morris (2002) maintained that ants directly con-

tributed 36% of the total sediment moved downslope on plots installed in areas subject to a range of fire severities and referred to this process as bio-transfer. In contrast, Shakesby et al. (2003, in press) maintained that in similar terrain the mounds and tunnel systems of the ant species *Aphaenogaster longiceps* probably acted more effectively in limiting the erosive effect of overland flow by increasing surface roughness and providing important sinks on water-repellent soil, thus tending to reduce rather than promote erosion.

5. Reconstruction of wildfire patterns and effects in history and prehistory

Understanding of the long-term geomorphological effects of wildfires has undoubtedly improved since Swanson (1981, p.414) remarked that the “effects of fire on long-term landform development are unknown” but the literature is still relatively limited in terms both of number of studies and geographical coverage. Although there is evidence of catastrophic fire and related soil erosion far back in geological history (Nichols and Jones, 1992; Scott and Jones, 1994; Scott, 2000), most literature on palaeowildfires has concentrated on the Holocene (the last c. 11,500 years) and on the late Holocene in particular (e.g. Rhodes and Davies, 1995). By no means all studies focus on the hydrological and geomorphological effects, but several indicate a strong link between fire-induced accelerated sediment yield and northern-hemisphere climatic perturbations, even though the magnitude of these changes was relatively small by long-term geological standards. Three periods in the Holocene have been of particular interest. First, analysis of charcoal, pollen and other fire proxies from lake records indicates that the Pacific Northwest and locations in the Rocky Mountains, USA experienced their highest fire activity during the early Holocene (11,000–7000 cal yr BP) (Millsbaugh et al., 2000; Whitlock et al., 2003). Second, the Medieval Warm Period lasting from about AD 900 to 1300 and with an optimum at AD 1090–1230 (Meyer et al., 1992) caused markedly accelerated sedimentation rates in geographically diverse parts of the USA, including Chesapeake Bay on the east coast (Brush, 1994), the northern Rocky Mountains and Pacific Northwest (Whitlock et al., 2003). Research by Meyer et al. (1992, 1995) in north-eastern Yellowstone National Park represents the only detailed investigation of long-term fire-related geomorphological response on a basin-wide basis known to the authors. Here, the Medieval Warm Period appears to have been drought-prone, with higher temperatures, reduced winter precipitation and intensified summer

convective–storm activity (Meyer et al., 1995). The geomorphological response was debris flow removal of sediment from recently burnt steep slopes building up foot-slope alluvial fans. Incision of main streams along narrow active floodplains occurred, leaving the former wet-phase floodplain as a terrace (Fig. 15). On the floor of Chesapeake Bay, Brush (1994) found marked increases in charcoal and accelerated sedimentation rates at this time (Fig. 16). Importantly, pollen evidence here showed a shift from wet-to dry-loving plants confirming dry conditions.

Third, the succeeding cool period known as the ‘Little Ice Age’ and lasting from c. AD 1300 to 1900, when temperatures were depressed by up to 2 °C in the northern hemisphere (Grove, 1988), shows an inconsistent response in terms of fire frequency and sedimentation rates. The marked reduction in charcoal occurrence and sedimentation rate from Chesapeake Bay (Fig. 16) prior to the influence of European colonisers accords with the change to cold winters and cool, wet summers normally associated with this event. The evidence for a distinct geomorphological response is less clear in Yellowstone. This period was only represented here by a relatively narrow main river terrace, implying modest increases in precipitation. Meyer et al.

(1995) concluded that the period was not uniformly cool and wet in this location and that a minor increase in runoff may have resulted largely from reduced evapotranspiration and more precipitation as snow rather than a wetter climate.

Overall, Meyer et al. (1995) found fire-related sedimentation occurring every 300–450 years in north-eastern Yellowstone during the late Holocene and the presence of strong 1300-year climate-driven alternations between fire-related sedimentation and terrace tread formation. In all, fire-related debris-flow and probable fire-related sediments made up an estimated c. 30% of the volume of alluvial fan materials during the Holocene, with a substantial part of the remaining volume of sediment probably comprising reworked or non-diagnostic fire-related sediment (Meyer et al., 1992). Such sediment redistributed within a fire-prone landscape may have a long residence time. For example, following the Buffalo Creek fire in May 1996 in the Colorado Front Range, USA, Moody and Martin (2001b) estimated that the steady state residence time in alluvial fans at the mouths of tributaries and along the main channels was about 300 years for stored sediment in two catchments of 26.8 km² and 122.4 km² in extent. With a fire recurrence interval of 20–

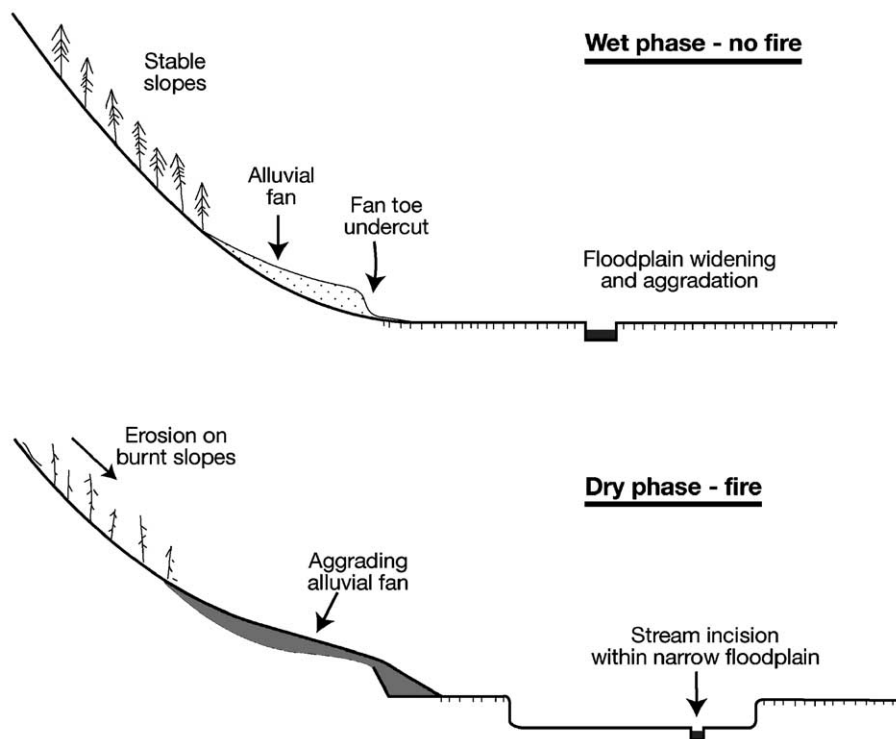


Fig. 15. Geomorphological response of hillslopes and floodplains to wet (top) and dry (bottom) phases during the Holocene in Yellowstone National Park, USA (interpreted from Meyer et al., 1992).

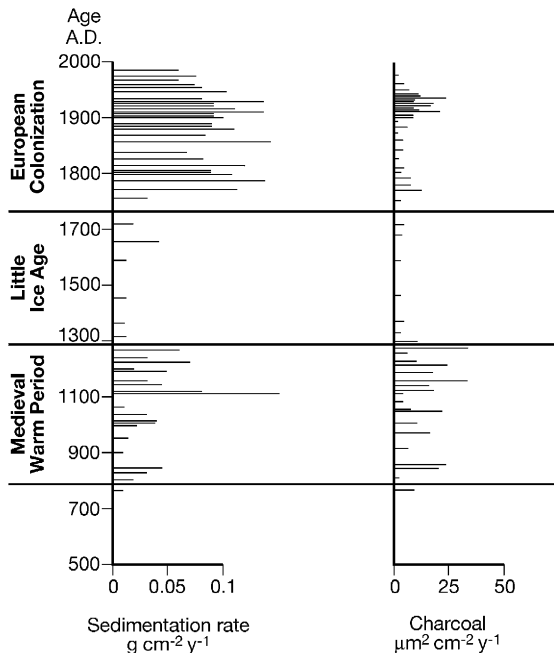


Fig. 16. Sedimentation rates and charcoal influxes for a sediment core from the Nanticoke River, which drains into Chesapeake Bay, USA (modified from Brush, 1994).

50 years, they surmised that many of the resulting depositional features might become long-lived phenomena, which would form landscape features affecting sediment transport in the next post-fire period of heightened hydrological and geomorphological activity.

Periods of erosional quiescence in the past can also be useful in assessing the impacts of fire. In the granitic mountains of central Idaho, Meyer et al. (2001) found a limited geomorphological response in the period 7600–6800 cal yr BP with similar quantities of sediment export to those monitored today, which extrapolated over the Holocene underestimated sedimentation rates found for the whole of this period. They reasoned that both these low-erosion episodes must have been counterbalanced by climate-induced, fire-related erosional events of accelerated sediment yield in order to account for the much higher average sediment yields found for the entire Holocene in the area.

Fire regime does not, however, seem to be closely linked to geomorphological response everywhere. Prosser (1990) investigated denudation in Wangrah Creek in the Southern Tablelands of New South Wales, Australia. He noted an increase in fire frequency beginning 3000–4000 years ago resulting from intensified Aboriginal burning, but found no associated increased rate of denudation or widespread alluviation, as argued previously by Hughes and Sullivan (1981). Swanson (1981)

reported research by Cooke and Reeves (1976) suggesting that in analysing 13 factors contributing to arroyo incision in the American Southwest between 1850 and 1920, fire regime affected by Anglo-Americans seemed to be irrelevant.

That there is nevertheless strong evidence of a link in many areas between climate change, wildfires and increased erosion has clear implications for future landscape effects of global warming in fire-prone terrain and this has not gone unrecognized (e.g. Meyer et al., 2001), not least because of an expected increase in the area affected (McCarthy et al., 2001). Whitlock et al. (2003) simulated summer soil moisture anomalies for 6000 years BP and for AD 2050–2059 for the northern USA compared to the mid-twentieth century. The first two periods indicate slightly drier conditions than the latter, leading the authors to conclude that the higher fire incidence in prehistoric times could be repeated in the middle of this century. Arguably, only by means of a long-term record will it be possible to separate the climate-induced impacts on fire frequency from those caused by land management changes. Some research studies carried out in the USA have alluded to the difficulties of looking for a ‘normal’ fire frequency over the last 200 years or so which spans the period of European expansion in the USA (Clark, 1988; Meyer and Pierce, 2003; Whitlock et al., 2003). Meyer and Pierce (2003) point out that it must be borne in mind that pre-European settlement tree densities and fuel loads in North American forests relate to cooler ‘Little Ice Age’ conditions, so that such densities should not be viewed as the pre-European settlement norm. The potential positive feedback effects of increased forest fire frequencies on global warming add complications. The Mauna Loa records of CO_2 concentrations in the atmosphere indicated that 2002 and 2003 were the first 2 years on record with increases > 2 ppm (Keeling and Whorf, 2004), which might well be linked to dieback following large fires in the northern hemisphere during these years, particularly in Europe. Whether or not this has been the case remains to be established. What is clear, however, is that fire frequencies and severities will vary in the future, and with them the short- and long-term hydrological and geomorphological impacts of wildfire.

6. Conclusion: retrospect and prospect

The role of wildfire as a potent hydrological and geomorphological agent is now widely recognized amongst the scientific, environmental and policy-making communities. Whilst the volume of literature and

our understanding of the different hydrological and geomorphological impacts of fire have all increased, there is an uneven coverage both in terms of topic and geographical area.

Current knowledge of the rates of rock weathering by fire is based on relatively few observations, but they suggest that it may be a significant means of rock breakdown in some fire-prone locations. The need for well-founded assumptions of long-term weathering rates when interpreting cosmogenic-isotope dates of the duration of rock exposure in fire-prone terrain, however, will doubtless lead to further investigations of this process. In contrast, many studies have identified, mostly under controlled conditions, potentially hydrologically and geomorphologically significant fire-induced changes in the soil and the temperature thresholds at which they occur. The temperature reached and its duration during a fire are seen as key variables in understanding the hydrologically and geomorphologically important soil changes. In the absence of direct soil temperature measurements during wildfires, a fire severity index based on the above-ground destruction of biomass is often used as a proxy. The reliability of such an index for predicting fire impacts at and below the soil surface is in need of improvement (cf. [Hartford and Frandsen, 1992](#)). The problem is compounded by there being no standardized classification of fire severity. The challenge remains, therefore, somehow to devise an alternative, more soil- and ground-oriented, and preferably, universally applicable fire severity classification for use in predicting the hydrological and geomorphological consequences of a wildfire.

In addition to the removal of vegetation and litter, the most widely reported fire impact on soil relevant to hydrology and erosion is the change to its water repellency characteristics, although other, less often cited changes to the soil affecting, for example, texture, aggregate stability, and macropore or root characteristics may be just as critical. Knowledge of the impacts of repellency on infiltration, overland flow and erosion has advanced considerably over the last two decades. It has been based in large part on laboratory analysis, on the one hand, and inference about its influence from measuring and monitoring these post-fire hydrological and geomorphological processes, on the other. There are, however, relatively few reliable data on fire-induced changes to water repellency and much of the progress has been made in the field of non-fire-related repellency. On fire-prone terrain, water repellency may be present but hydrologically and geomorphologically ineffective and as a spatially and temporally often variable property,

although it can also be induced, enhanced or eliminated by fire. Unravelling the complexities and understanding how repellency interacts with different post-fire biomass and soil characteristics at different scales and establishing its relative importance in conjunction with other fire-induced changes to soil characteristics are important goals for the future, given the potential impacts this soil property can have on overland flow, runoff and erosion.

Significant recent advances have been made concerning measurement and understanding of the range of hydrological and geomorphological impacts of wildfire. The geographical range of observations and measurements of these impacts has been extended beyond the North American, and particularly the western USA 'heartland' of earlier research into fire impacts, to Europe (especially the northern Mediterranean Basin), and to Israel, southern Africa and Australia. From the range of terrain types studied, a picture of regionally and locally distinctive post-fire responses is emerging. Thus, for example, the process of dry ravel is very important where it occurs on steep chaparral terrain in the western USA, but it seems to be essentially restricted to this region. The widespread introduction of plantations of highly flammable tree species (pines and eucalypts) to formerly intensively used land in the Mediterranean leads to post-fire hydrological and geomorphological changes of a different character to those affecting more wildland-type areas of, for example, western USA or Australia. It is also apparent that regional differences in climate, geology and vegetation can also be overprinted by localized small-scale effects (e.g. post-fire bioturbation activity, litter dam formation, the formation of debris dams in stream channels), adding to the local distinctiveness of post-fire responses. There remain, however, types of fire-prone terrain (e.g. boreal forest, seasonally dry tropical forests) for which there is little information on fire-induced hydrological and geomorphological impacts.

In the last 20 years or so, the range of hydrological and geomorphological processes identified as being enhanced by wildfire has increased. In addition to, for example, rainsplash detachment, surface wash, dry ravel, snow avalanching and frost creep, there have been detailed observations of landslides, debris flows and the downslope transfer of sediment by faunal activity. As regards hydrological impacts, there is more emphasis in the literature on the impact of wildfire on soil erosion by water than on the timing and quantity of runoff and the effects on stream channels. Despite the logistical difficulties, there is a practical need to understand fluvial responses in order to improve prediction of

potential post-fire flooding problems and impacts on channel stability as well as the implications for water supply.

There is now a large body of data concerning post-fire monitoring of soil erosion by water mainly at small scales, but it could be used to better effect if the following problems could be solved:

- (1) A means of gauging the seriousness of post-fire erosion in the many forested landscapes where erosion is negligible is clearly required. Some development of the tolerable soil loss rate developed for erosion from agricultural land might be applicable. The weak knowledge, however, about soil renewal rates under natural conditions makes judging the degradational implications of wildfire difficult, because of the sometimes long time-scales involved in the post-fire relaxation time until pre-fire conditions similar to those pertaining prior to fire are reinstated. There is also the related matter of the impact of the preferential loss of fine particles on soils subject to post-fire erosion.
- (2) There is a relatively poor understanding of the impact of soil erosion at scales larger than small plots. Most erosion monitoring has been carried out at small scales (transects, small plots and, rarely, hillslope measurements) with relatively few studies carried out at larger scales. Yet, rates based on the former are often quoted as if they were figures representative of catchments or even regional sediment yields despite the known negative relationship between scale and soil loss. At larger scales, therefore, the post-fire geomorphological impact should often be viewed as a matter of soil redistribution rather than of simple net loss. The development and application of sediment source tracing techniques and of cosmogenic radionuclides to post-fire terrain could help to complement our knowledge of small-scale processes by providing potentially a catchment-wide overview of soil redistribution over diurnal to decadal timescales.
- (3) In connection with erosion monitoring, erosion rates are expressed in a variety of units, leading to errors and misunderstandings. It would be helpful if, wherever possible, a standard notation were used. The use of tonnes per hectare is virtually a lingua franca for soil erosion experts, and provided the scale of measurement is made clear, its universal use would help comparison of the results of erosion studies.

Interest in and understanding of post-fire erosion other than by water alone (especially landslides, debris flows) have increased within the last two decades, but have been largely restricted to the USA. The extent to which such phenomena are absent from other post-fire environments because regolith and slope conditions are not conducive or they have simply not been reported remains to be established. In addition, the geomorphological significance of landslides and debris flows in landscape development is currently difficult to judge as few estimates of erosion rates by these processes have been made.

Finally, in addition to the accumulation of fuel loads due to fire suppression in some parts of the world, an important resource in helping prediction of the hydrological and geomorphological consequences wildfire frequency and severity is the Holocene sedimentological record, which can aid in formulating a long-term understanding of the links between changes in climate, fire behaviour and post-fire hydrological and geomorphological impacts. Currently, research into future hydrological and geomorphological impacts caused by wildfires lags behind that of interest in the prediction of the occurrence of wildfires. Further work on the impacts of palaeowildfires will improve the basis for predicting these future impacts.

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