HYDROLOGICAL PROCESSES
Hydrol. Process. **22**, 4415–4435 (2008)
DOI: 10.1002/hyp.7045

Rain event properties in nature and in rainfall simulation experiments: a comparative review with recommendations for increasingly systematic study and reporting

David Dunkerley*
School of Geography and Environmental Science, Building 11, Clayton campus, Monash University, Victoria 3800 Australia

Abstract:

In hydrology and geomorphology, less attention has been paid to rain *event* properties such as duration, mean and peak rain rate than to *rain* properties such as drop size or kinetic energy. A literature review shows a lack of correspondence between natural and simulated rain events. For example, 26 studies that report event statistics from substantial records of natural rain reveal a mean rain rate of just 3·47 mm h^{-1} (s.d. 2·38 mm h^{-1}). In 17 comparable studies dealing with extreme rain rates including events in cyclonic, tropical convective, and typhoon conditions, a mean maximum rain rate (either hourly or mean event rain rate) of 86·3 mm h^{-1} (s.d. 57·7 mm h^{-1}) is demonstrated. However, 49 studies using rainfall simulation involve a mean maximum rain rate of 103·1 mm h^{-1} (s.d. 81·3 mm h^{-1}), often sustained for >1 h, exceeding even than of extreme rain events, and nearly 30 times the mean rain rate in ordinary, non-exceptional, rain events. Thus rainfall simulation is often biased toward high rain rates, and many of the rates employed (in several instances exceeding 150 mm h^{-1}) appear to have limited relevance to ordinary field conditions. Generally, simulations should resemble natural rain events in each study region. Attention is also drawn to the raindrop arrival rate at the surface. In natural rain, this is known to vary from <100 m^{-2} s^{-1} to >5000 m^{-2} s^{-1}. Arrival rate may need to be added to the list of parameters that must be reproduced realistically in rainfall simulation studies. Copyright © 2008 John Wiley & Sons, Ltd.

KEY WORDS rain event; rainfall rate; rainfall intensity; rainfall simulation; drop arrival rate

Received 19 September 2007; Accepted 28 February 2008

INTRODUCTION

Rain event properties such as duration and intensity may influence the partitioning of water among interception, evaporation, infiltration, ponding, and overland flow (Hawke *et al.* 2006; Struthers *et al.* 2007). They also exert a fundamental control on water-driven erosion processes such as splash, sheetflow and rillflow (Watung *et al.* 1996; Mermut *et al.*, 1997; Parsons and Stone 2006). For surface geomorphic processes such as splash dislodgment and soil particle entrainment, there have been extensive efforts to explore the relevant properties of rain (kinetic energy, momentum, storm depth, EI_{30}, etc. See van Dijk *et al.* (2002) for a review of rainfall intensity—kinetic energy relationships). Likewise, in seeking to understand water losses to canopy interception, extensive work has examined above-canopy rainfall, and patterns of sub-canopy throughfall, stemflow, and leaf drip, in relation to storm size, intensity, and other properties such as time of day (Keim *et al.*, 2006). Very similar enquiries have arisen regarding the fate of agrochemicals applied to plant foliage, where wash-off is a concern, and in urban flash flooding studies. Geochemical and nutrient balance studies, including studies of phosphorus export from agricultural soils, also have a major concern with storm properties (Angermann *et al.*, 2002). The agricultural sector has also generated studies of rainfall events in relation to water use efficiency. Greater soil evaporative losses from rainfall, and reduced water use efficiency, are expected in regions that have a higher frequency of small rainfall events (Sadras 2003).

However, while the literature contains a wealth of studies where rain event properties are shown to be significant in infiltration, generation of overland flow, and erosion, there has been less discipline-specific exploration of the definition and analysis of rain event properties, and their geographical and temporal variability. This information is really essential in order to develop a correct view of the rain event properties of different storm types and geographical regions (e.g. tropical convective rain, dryland thunderstorm rain) and their impact on environmental processes. In this paper it will be argued that the absence of this perspective is particularly evident in a comparison of the kinds of rain event properties employed in rainfall simulation experiments with the properties of natural rain events. Simulated rain is widely used as a way of controlling and standardizing experimental conditions, but it has been argued (Agassi and Bradford, 1999) that a lack of correspondence between natural and simulated rain introduces some doubt about the validity of data and conclusions arising from some rainfall simulation studies. Reproducing natural rain event properties in rainfall simulation studies requires that we have relevant knowledge

---

* Correspondence to: David Dunkerley, School of Geography and Environmental Science, Building 11, Clayton campus, Monash University, Victoria 3800 Australia. E-mail: david.dunkerley@arts.monash.edu.au

of the corresponding natural properties. There has been a tendency to focus on issues such as drop size and kinetic energy (van Dijk *et al.*, 2002), at the expense of a corresponding concern with event intensity and duration properties. Furthermore, there are rain event characteristics whose significance has not been properly explored, and which may prove to offer significant explanatory power. An instance noted later is the drop arrival rate at the ground, expressed in impacts per unit area and time. Some relevant field-based data are available on this event characteristic, but little exploration of its importance has been undertaken even though physical intuition suggests that drop arrival rate could be a parameter of relevance to the understanding of splash, surface seal formation, canopy interception, and other environmental processes.

In addition to the rationales just presented, further investigation of rain event properties is also justified by the likelihood of changes in various rain event properties as global and regional environmental change develop in coming decades. Changes in atmospheric composition are foreshadowed to result in an increase in extreme events, including rainfalls (Milly *et al.*, 2002; Palmer and Räisänen, 2002; Chauvan and Denvil, 2007), as well as greater occurrence of drought and a shift to generally more arid climate in some regions (e.g. see Seager *et al.*, 2007 on the instance of the south-west USA). Geomorphologists have opportunities to contribute to the prediction and analysis of the consequences in a range of landscape contexts (Goudie, 2006). A study of storm precipitation in the USA based on a 30-year database (1972–2002) containing more than 1500 recording raingauge records shows regional differences among temporal trends, but a widespread tendency for 15-min rain rates to increase towards the present day (Palecki *et al.*, 2005). The consequent need for adaptive management of urban drainage systems has been highlighted (Trefry *et al.*, 2005; Guo 2006). Potential consequences for hydrology and for soil erosion have also been highlighted (Angel *et al.*, 2005). Nearing (2004) estimated that soil loss rates would increase by 1·7% for each 1% increase in annual rainfall total. Increases in extremes of rain intensity may provide a significant component of any future increase in soil loss related to rainfall environment. This is very relevant to the use of simulated rain in the study of soil erosion, discussed later, and in particular to the storm intensities and durations that form the basis of rainfall simulation studies. A sound awareness of rain event intensities, and of the changes envisaged for coming decades, may guide work exploring further the implications for infiltration, production of overland flow, and erosion.

This paper therefore presents a brief review of the nature of rain events based on recent hydrologic, erosional and other studies including those that have used rainfall simulation as a research tool. The review cannot hope to be comprehensive, given the vast scope of the published literature. The examples are drawn primarily from work published within the last 10–15 years. The fundamental questions that are addressed are as follows:

1. How are natural rain events described in the literature, particularly in terms of rainfall intensity (rainfall rate) and event duration?
2. Are these characteristics of natural rain events being adequately represented in rainfall simulation research?
3. Are there additional rain properties that might be relevant to understanding hydrologic and erosional processes at the surface of the Earth (e.g. drop arrival rate), but which have not yet been adequately explored and reproduced in simulation experiments?

THE NATURE OF GLOBAL PRECIPITATION

It is useful to begin with some brief remarks on the nature of rain globally (snowfall and related cryosphere processes are not considered). In the absence of net changes in average atmospheric moisture content, rain balances the upward evaporative flux from the land and oceans (Trenberth 1998). Global average rain amounts are 2–3 mm day^{-1} (Chiu and Chang 2001; Shin *et al.*, 2001). Over land, rain is characteristically an afternoon phenomenon, while over the oceans it falls predominantly during the night and very early morning (Chiu and Chang 2001). While the upward evaporative flux is relatively steady, the rate of rainfall exhibits characteristic bursts separated by rainless periods lasting from minutes to weeks or months. The structure of rain intervals and the intervening rainless gaps has been analysed in a number of studies, and the temporal pattern shows some scale-invariant properties that reveal self-organized criticality (Peters and Christensen, 2002). Rain has been described therefore as a relaxation process, akin to the episodic release of accumulated strain in seismic events (Peters and Christensen, 2006). Given the burst-like occurrence of rain, it is clear that global evaporation is not balanced by a steady downward rain flux. Indeed, given that rain may only occur for 2–5% of the time (Jones and Sims, 1978), the average rain depth in individual events may be 20–50 times the global mean daily rain amount. Several factors are needed for the arrival of large rain depths and intensities: these include rapid inflows of moisture-laden air, large rates of decrease of cloud moisture content and small intra-storm evaporative losses during droplet descent to the surface (Smith *et al.*, 2001). An analysis of raingauge records with 1-min resolution has shown that there are variations in rain properties associated with broadly distinct climatic zones (Jones and Sims, 1978). These include maximum 1-min rates of 68 mm h^{-1} for mid-latitude continental interiors, but up to 238 mm h^{-1} for maritime subtropical conditions. Patterns of rain event intensity are examined in more detail below, since this is one of the key rain event parameters of relevance to hydrologic and erosional processes.

HOW INTENSE IS NATURAL RAIN?

The question of rainfall intensity is a central one, since this parameter influences so many relevant outcomes for

DOI: 10.1002/hyp

the landscape (e.g. energy delivery, the formation of saturation overland flow, initiation of sediment dislodgment and transport, raindrop impact crust and seal development on soil surfaces).

In terms of preliminary terminology, it is necessary to distinguish the concept of 'instantaneous rainfall rate', I (often simply referred to as the 'rainfall intensity'), and 'mean rainfall rate', R, based on the extrapolation of an instantaneous rate through a time period longer than would be likely to experience truly steady rainfall conditions, or reported when the total rain depth in a rain event (or other defined time interval, such a minute, hour, or day) is divided by the event duration to express the mean conditions during the event, thus:

$$R_e = D_e/T_e$$

or

$$R_t = D_t/T_t$$

where R_e is the mean rainfall rate in an event of total depth D and duration T. The subscript 'e' indicates that the calculation is based on the depth and duration of an event, while the subscript 't' indicates that the calculation is based upon time-averaging or extrapolation through a defined interval such as 6 min, 30 min, 1 h, 24 h, etc. In the soil erosion literature in particular, the use of I instead of R_t for rain rate defined by time-averaging is firmly established; in particular, a great deal of research has focused on the explanatory power of the EI_{30} parameter, which refers to the maximum rain rate that can be identified by tallying the total rain depth in any 30-min period of a rain event, multiplied by the kinetic energy delivery per mm depth (Wischmeier and Smith, 1978). It is recommended that the term 'intensity' be limited to what are effectively instantaneous rates, i.e. with integration times of 10 s or less. Clarity in the reporting of rain rates, and a distinction between R and I, is often missing in the published literature, where it may not be apparent whether mean 'intensities' expressed in mm h^{-1} are actually R_e or R_t results, and in the case of the latter, over what time period the rain rate was tallied. The use of the expression 'rain rate', with an indication of how the rate was arrived at, will assist in building a clearer reporting of rain event properties in the experimental literature. Hereafter reference to intensity (I) will mean a near-instantaneous rainfall rate, and $I_{\max}$ the largest such rate encountered in a period of rain. Where it was reported, the associated integration time will be noted.

Two other broad terms that appear in the literature include 'conditional' rainfall rates (R_c), given by the ratio of cumulative rainfall to the duration of rain, and the 'unconditional' rainfall rate (R_u), which is the ratio of the cumulative rainfall to the duration of raingauge operation (Tokay *et al.*, 2003). Unconditional daily rain rates are sometimes reported (Nicholls and Kariko, 1993), but since they include rainless intervals, these are of limited use in studies of hydrologic and erosional processes. The definition of the start and end of rain is important in defining rain rates on an event basis, and is problematic with tipping-bucket rain gauges. This is because a rain event starting or ending with low-intensity rain may take a considerable time to fill a gauge bucket (or indeed may fail to do so at the end of a shower), leaving an interval of time during which rain fell but from which no tip event was recorded. As Tokay *et al.* (2003) show, this can leave an uncertainty of more than 2 h in both the start and end times of a rain event. Thus, the true average rain rate in an event where only a central hour of 10 mm was recorded, but where an additional (but not separately resolved) 0·2 mm fell in two preceding and two following hours, would not be 10 mm h^{-1} but rather just 2·08 mm h^{-1}.

The inclusion of rainless gaps within a longer rain event clearly also has the potential to alter the mean rain rate that would be recorded for that event. The literature exhibits a range of rules for the inclusion of rainless gaps within what are reported as single, continuous events, and for the length of dry interval that would result in two separate events being tallied. The approaches used to define rain events will be explored further in a separate paper (Dunkerley, Submitted). However, the range of rainless intervals permitted within what is counted as a single 'event' range from 3 min (Vilar and Burgueño, 1995), 15 min (Cattan *et al.*, 2006), 30 min (Cosgrove and Garstang, 2005), 60 min (Pawlina, 2002) to 90 min (Chahinian *et al.*, 2005). Clearly, apparatus and approaches used in defining rain events can have a significant effect on reported mean rain rates, as has been argued by Bracken *et al.* (2008), Cattan *et al.* (2006), and others. Cattan *et al.* (2006) have argued for the adoption of precise definitions of both rain events and surface runoff events (they suggest no cessation of flow >5 min duration as a criterion for continuous runoff, in their bounded runoff plots of 3000 m^2), so that the occurrence and duration of runoff events can be linked to sound measures of the causative rain events. Care is required in reading the literature, since some authors (e.g. Cutrim *et al.*, 2000) define an event as a 'storm' (a period of rain bounded by dry intervals of at least 1 h) together with the dry interval (an 'interlude') that follows it. This leads to low event rain rates, but higher storm rates.

Gaps in rain or overland flow are of great relevance to many hydrologic and geomorphic processes. Dry gaps within rain events allow evaporation from wet plant surfaces to offset prior filling of interception stores, and so tend to increase overall interception losses in the event. They may also increase the time for the canopy to become fully wetted, and hence reduce the time through which intra-storm evaporative losses can proceed at the maximum rate (Carlyle-Moses, 2004). Gaps in the occurrence of rain and/or overland flow may allow infiltration to partially empty surface ponding sites (Aryal *et al.*, 2005), with renewed detention of flow once overland flow re-commences. During such gaps, the extent of near-surface soil saturation may also decline, with further effects on the production of overland flow once flow resumes.

Classifications of rainfall rates

Descriptive terms are needed for reporting rain rates. Among several classifications of rain rate, that of Tokay and Short (1996) (Table I) is representative. This categorizes rain rates exceeding 5 mm h^{-1} as 'heavy', as does a classification presented by Laakso *et al.* (2003). A rain rate of 5–10 mm h^{-1} is also often seen as a 'dividing line' between stratiform rains and convective rains (Tokay and Short, 1996; Pawlina, 2002). Some authors refer to rain rates <5 mm h^{-1} as 'weak', and >5 mm h^{-1} as 'strong' (Pawlina, 2002; Dairaku *et al.*, 2004). There is considerable variation in the use of descriptive terms, some of which undoubtedly reflects regional variation in rainfall climate, but some of which may reflect a misapprehension of the true range of rain rates. Moszkowicz (2000) indicates that 1·2 mm h^{-1} would be regarded as a moderate rain rate in Europe, while Toba and Ohta (2005) describe 2·3 mm h^{-1} as a high rate. Others have employed different criteria, such as mean $R > 50$ mm h^{-1} adopted to define 'heavy' storms (Vilar and Burgueño, 1995). Hunsche *et al.* (2007) adopted a three-way classification of their simulated rainfall: light rain (0·5 mm h^{-1}), heavy rain (5 mm h^{-1}) and torrential rain (48 mm h^{-1}). Also for rainfall simulation purposes, Tejada and Gonzalez (2007) employed $R = 60$ mm h^{-1}, which they described as 'low intensity', and $R = 140$ mm h^{-1}, which they described as 'high intensity'. As the ensuing discussion will suggest, there is scope for a classification that extends to much higher rain rates than listed in Table I (perhaps adopting categories such as 'torrential' or 'exceptional'), since rain rates sustained at 50–100 mm h^{-1} are known (Table II). Nevertheless, as the data in Table III illustrate, many published reports do indeed confirm the dominance of rain rates in the 'light' and 'moderate' categories of Tokay and Short (1996). These will be discussed further below. The next two sections briefly highlight the rain event characteristics from about 65 published reports selected to represent a wide sample of global locations. The analysis begins with reports of high rain rates, and the section following that considers low rain rates.

Intensity–duration–frequency (IDF) curves

Relationships among intensity, event duration, and the recurrence interval are often established from large regional rainfall data sets (Koutsoyiannis *et al.*, 1998; Gerold and Watkins, 2005) in order to rate the relative rarity or commonness of particular combinations of rain rate and event duration. Discussion of these procedures lies beyond the scope of this paper. However, they clearly establish that intensity and duration are related, and that increasing the value of one or both is associated with a decline in the mean frequency with which such events occur. In the majority of field studies in soil erosion and the production of overland flow, insufficient historical rainfall records are available to allow the construction of IDF curves. Moreover, the main purpose of this paper is to illustrate the kinds of intensities reported in the literature, without extending the treatment to cover formal analysis of the rarity or commonness of the rain events. Many of the high rain rate events reported in Table II are reported as rare or extreme, while many of the low rain rate reports in Table III reflect more common rain event characteristics. Therefore, reading these tables provides a general guide to the rain rates reported in studies in hydrology, geomorphology, and related fields. While the IDF approach views event duration and rain rate as existing as continuous variables, the actual precipitation mechanisms that result in rain (such as stratiform or convective) precipitation evidently tend to produce rain events with distinct properties.

Table I. The rain rate classification of Tokay and Short (1996)

Descriptive rain rate category	Rain rate R (mm h^{-1})
Very light	<1
Light	$1 \leq R < 2$
Moderate	$2 \leq R < 5$
Heavy	$5 \leq R < 10$
Very heavy	$10 \leq R < 20$
extreme	$R \geq 20$

Published reports of high rain rates

There are many accounts of very large daily rainfall depths, often sustained over several days. Dunkerley *et al.* (Submitted) cite several published studies of 24 h totals of 400–600 mm, often from reports dealing with cyclonic or other intense rains, and with the resultant flash flooding. Exceptional though such daily rain amounts are, when expressed as average hourly rain rates, they are less striking: 600 mm d^{-1} is the equivalent of a sustained 25 mm h^{-1}. Thus, what is really exceptional in many of these large events is the persistence of the heavy rain through so many successive hours. World record point rainfalls have been listed by Barcelo *et al.* (1997), and these include 1825 mm in 24 h at Reunion Island in the Indian Ocean, which is the equivalent of a sustained rain rate of 76 mm h^{-1}.

Although high daily rain depths do not imply equally high rain rates, some very high values of what are in effect intensities have certainly been reported. The world record 1-min total listed by Barcelo *et al.* (1997) is 31·2 mm (equivalent to 1872 mm h^{-1}). (These authors also report that during sustained heavy rain on Reunion, a 0·5 mm capacity tipping bucket gauge typically tips at the remarkable rate of once every 10 s). From 49 years of rainfall observations collected in Barcelona, Spain, Vilar and Burgueño (1995) reported an $I_{\max}$ of 586 mm h^{-1}. This is virtually an instantaneous rate, since the data were collected with a Jardí rain rate recorder with a response time of 10 s. The highest near-instantaneous rain rate known to the writer was reported by Nystuen and Amitai

DOI: 10.1002/hyp

Table II. Compilation of rain rates from 35 published reports in which high or extreme rain rates are recorded. See text for comments

Location	Rain rates (including integration time) mm h^{-1}	Source of data
Brunei	$I_5 = 175\cdot 2$	Dykes and Thornes (2000)
Alicante, Spain	$I_2 = 143\cdot 7 - 293\cdot 6$ for 5–50 year ARI	Garcia Bartual and Schnedier (2001)
Alicante, Spain	$I_5 = 106\cdot 8,\ 221\cdot 7$	Garcia Bartual and Schnedier (2001)
NE Spain	Average $R_e = 91\cdot 8$ $I_{30} = 170$	Ramos *et al.* (2004)
NE Spain	$I_1 > 250$ $I_5 = 180$	Martinez-Casanovas *et al.* (2002)
Spain	$I_{15} = 76$	Bochet *et al.* (2006)
Barcelona, Spain	$I_{max} = 586$	Vilar and Burgueño (1995)
Normandy, France	$I_6 = 105$ $R_e = 6\cdot 8$	Cerdan *et al.* (2002)
Brazil	$I_{10} = 100\cdot 6$ $I_{60} = 57\cdot 9$	Germer *et al.* (2005)
Taiwan	$I_{60} = 103$	Cheng *et al.* (2005)
Maryland USA	$I_{15} = 102$	Smith *et al.* (2005)
South China (Guangdong Province)	$I_{10} = 162$ $I_{60} = 73$	Woo *et al.* (1997)
Missouri USA	$I_{60} = 76$	Winston and Criss (2002)
France	$I_{60} = 50$ (sustained for 6 h)	Gaume *et al.* (2004)
Taiwan (Typhoon Herb, 1996)	$I_{60} = 96\cdot 5$	Lin and Jeng (2000)
Pyrenees, Spain	$I_{10} = 153$	Gutierrez (1998)
Pyrenees, Spain	$I_{30} > 150$	Alcoverro *et al.* (1996)
Kalimantan, Indonesia	I_{60} reaches 20–25 mm h^{-1}	Asdak *et al.* (1998)
Borneo	Maximum $I_5 = 73\cdot 7$ Intensity >50 mm h^{-1} for 0·09% of rain time	Bidin and Chappell (2006)
Sarawak, Malaysia	Maximum event mean intensity (8 min duration) = 68·2 mm h^{-1} Average $R_e = 14\cdot 9$ mm h^{-1}	Kume *et al.* (2006)
Europe	Annual maximum I_{15} at many stations <10 mm/15 min–11 mm/15 min (40–44 mm h^{-1})	Rauch and Toffol (2006)
French Alps	Summer storms often reach 50 mm h^{-1}	Cras *et al.* (2007)
Cairngorms, Scotland	Maximum $I_{30} = 33\cdot 5$	McEwen and Werritty (1988)
Valencia district, Spain	$I_5 > 120$ mm h^{-1} at least once per year Average intensities 3–4 mm h^{-1}	Belmonte and Beltrán (2001)
Odra River, Poland	Mean intensity during 2 h rain = 96·7 mm h^{-1}	Dubicki *et al.* (2005)
Greece	Maximum I_{30} from 31 rain events = 56·4 mm h^{-1}; 55% of max. I_{30} values <10 mm h^{-1}	Valmis *et al.* (2005)
Cebu, Phillipines	Highest mean event intensity 44·7 mm h^{-1}; highest $I_5 = 142\cdot 3$ mm h^{-1}	Fornis *et al.* (2005)
United Kingdom	Mean event intensities (extreme events) 200–300 mm h^{-1}	Hand *et al.* (2004)
Hong Kong	Maximum intensities 30 mm h^{-1}–50 mm h^{-1}	Li and Lai (2004)
Southern France	Maximum $I_1$6–149 mm h^{-1} (from 28 events)	Chahinian *et al.* (2005)
Texas USA	Supercell rain: mean event intensity over 2·75 h was 203 mm h^{-1}	Smith *et al.* (2001)
Malaysia	I_1 maximum 200 mm h^{-1}; I_{60} maximum 74 mm h^{-1} (264 events)	Ziegler *et al.* (2006)
Tanzania (Kwalei catchment)	Maximum I_{30} 2·0–42·8 mm h^{-1}	Vigiak *et al.* (2006)
Spain	Maximum $R_{t5} > 120$ mm h^{-1} (flash flood of 10 June 2000)	Amengual *et al.* (2007)
Taiwan	Maximum 70 mm h^{-1} (in typhoon Xangsane, November 2000)	Chen *et al.* (2006)

(2003) from rain measured in Florida from the sound generated by drops striking water, and recorded using an acoustic sensor submerged at a depth of 1·5 m. This system has very high response rate, since there is no time delay related to water flow though a funnel or siphon. The maximum 5-s I_{max} listed by these authors is 1600 mm h^{-1}.

More commonly reported are rain rates over periods of 1–15 min, and these are clearly time-integrations (average rain rates), not true intensities. Barcelo *et al.* (1997) list the world record 15-min total as 198·1 mm. A range of rain rates has been reported using integration times from 1 min to 1 h (Table II). Table II shows that R_{t60} rates in the range 50–100 mm h^{-1} are encountered

Table III. Compilation of rain rates from 31 published reports of low or moderate rain rates. See text for comments

Location	Rain rates (including integration time)	Source of data
Puerto Rico	80% of hourly intensities $\leq 2{\cdot}5$ mm h^{-1}	Scatena (1990)
Brazil	Mean intensity 6·7 mm h^{-1} Maximum $R_{t10} = 100{\cdot}6$ mm h^{-1} Maximum $R_{t60} = 57{\cdot}9$ mm h^{-1}	Germer *et al.* (2005)
Japan	Mean intensities 1·2 mm h^{-1}–1·4 mm h^{-1}	Iida *et al.* (2005)
Brazil (near Manaus)	Mean intensity 5·2 mm h^{-1}	Lloyd (1990)
Washington State USA	Mean intensity 0·25–2·0 mm h^{-1}	Pypker *et al.* (2005)
Ecuador	Mean intensity 1·1 mm h^{-1} (based on 672 events)	Fleischbein *et al.* (2005)
Malaysia	Mean intensity 7·8 mm h^{-1}	Konishi *et al.* (2006)
Amazonia	Mean intensity 3·8 mm h^{-1}	Zeng *et al.* (2000)
Germany	Mean intensity 0·59 mm h^{-1} (winter) Mean intensity 0·93 mm h^{-1} (summer)	Hörman *et al.* (1996)
Georgia USA	Mean event intensity 1·8 mm h^{-1} (based on 140 rain events) Maximum recorded: 14·4 mm h^{-1}	Bryant *et al.* (2005)
Japan and Siberia	Mean intensity 0·6–2·3 mm h^{-1}	Toba and Ohta (2005)
Borneo	$I_5 < 4{\cdot}8$ mm h^{-1}	Bidin and Chappell (2006)
West Java, Indonesia	Mean intensity 4·3 mm h^{-1}	Van Dijk and Bruijnzeel (2001)
Fukuoka, Japan	Mean intensity 3·6 mm h^{-1} Maximum recorded: 24·2 mm h^{-1}	Sato *et al.* (2004)
China	Three rain regimes recognized, having max. $I_{30} = 8{\cdot}4$ mm h^{-1}, 15·6 mm h^{-1}, and 6·6 mm h^{-1}	Wei *et al.* (2007)
Tigray uplands, Ethiopia	Mean storm intensity <4 mm h^{-1} 90% of rain received at <12 mm h^{-1}	Deschemeeker *et al.* (2006)
Tibetan Plateau	Mean intensity 1·27 mm h^{-1}	Yang *et al.* (2007)
South-east Spain	Most events have intensity <10 mm h^{-1}, and often <5 mm h^{-1}	Bracken *et al.* (2008)
Thailand	90% of rain events <5 mm h^{-1} >50% of events <1 mm h^{-1}	Dairaku *et al.* (2004)
Greece	10 278 events in 49 years yielded mean event rate $R_e = 5{\cdot}0$ mm h^{-1}	Vilar and Burgueño (1995)
Florida USA	Mean event intensity $R_e = 12$ mm h^{-1} (summer), 6 mm h^{-1} (winter) (197 rain events). 92% of events <24 mm h^{-1}	Cosgrove and Garsteng (1995)
Palau, Micronesia	Intensities generally <1 mm h^{-1}	Kubota *et al.* (2006)
Finland	Mean intensity 0·9 mm h^{-1}	Laakso *et al.* (2003)
China	60% of rain events had $I_{10} < 5$ mm h^{-1}	Li *et al.* (2004)
Switzerland (Mt Rigi)	Mean event rain rate 0·8 mm h^{-1}	Volken and Schumann (1993)
USA, several states, plus Marshall Islands	Various mean event rain rates, 1·19–6·33 mm h^{-1}	Guy *et al.* (1990)
Palestine	I_{60} range 2·4–11·2 mm h^{-1}	Hamad *et al.* (2006)
Colombia	Mean intensity 5·46 mm h^{-1}; only ∼2% of event intensities >5 mm h^{-1}	Tobón Marin *et al.* (2000)
Japan	I_{30} range 4–24 mm h^{-1} (mean event intensity, 13 events, 2·48 mm h^{-1})	Haga *et al.* (2005)
Puerto Rico	Mean $R_e = 3{\cdot}0$ mm h^{-1} (based on 80 rain events)	Schellekens *et al.* (2000)
China	Mean $R_e = 5{\cdot}3$ mm h^{-1}	Zhang *et al.* (2006)

during typhoon rains, intense convective storms, etc., and are often reported in studies of flash flooding. For shorter integration times, such as R_{t5}–R_{t10}, rain rates of 150–180 mm h^{-1} have been reported. However, these are relatively uncommon to exceptional events in most regions. The consequences of these very high rain rates are often extraordinary, and such rain rates are widely reported in the literature on extreme events. The mean of 17 studies reporting hourly or event mean R data (Table II) is 86·3 mm h^{-1} (standard deviation 57·7 mm h^{-1}). However, the compilation of rain rates in Table III, many of which are derived from raingauge records spanning years, suggests that in many areas, the vast bulk of rain events exhibit much lower rain rates, falling within the range spanned by the rain rate categories of Tokay and Short (1996).

Published reports of low rain rates

A number of the studies reported in Table II, showing high rain rates, are from studies of extreme events, or report high rain rates only over short integration times (R_{t5}, R_{t10}, R_{t30}, etc). Owing to their highly visible and often damaging effects, extreme events have been studied frequently. However, there is a substantial literature from field studies where rainfall has been recorded for a period of many years, resulting in a description of ordinary rain event properties (Table III). These results demonstrate that in a wide range of geographic and climatic environments, mean rain rates are low, often lying below 10 mm h^{-1}. In many cases, the mean rain rates lie below 5 mm h^{-1}, and some are as low as 0·6–0·9 mm h^{-1}. From 867 rain intervals at Careiro in the Amazon, where the mean annual rainfall exceeds 2000 mm, Cutrim *et al.* (2000) report a mean rain rate of just 2·4 mm h^{-1}. A second example is provided by the mean rain rate from 10 278 events in Greece, where the mean rain rate was 5·0 mm h^{-1} (Vilar and Burgueño 1995). From 12 years of rain event data from near Madrid, Marques *et al.* (2008) reported a geometric mean R_{t30} of 3·8 mm h^{-1}. From records of 517 rain events in central Amazonia (delimited by a 3 h dry interval), Cuartas *et al.* (2007) established a mean event rain rate of 7·13 mm h^{-1}. Frequency distributions of rain rates in longer-term studies (see Figure 11 in Bracken *et al.* (in press)) show that the distribution is strongly positively skewed, and the majority of events therefore have a rain rate well below the mean rate. Thus, Dairaku *et al.* (2004) reported from Thailand that 90% of rain events exhibited $R_e < 5$ mm h^{-1}. Similarly, from Puerto Rico, 80% of hourly intensities are <2·5 mm h^{-1} (Scatena, 1990), and at a site in Borneo, even 5-min rainfall rates (R_{t5}) corresponded to 4·8 mm h^{-1} or less, and for only 0·09% of the record did 5-min rates exceed the equivalent of 50 mm h^{-1} (Bidin and Chappell, 2006). From this short compilation of rain rate data, it is clear that average rain rates (even in tropical environments) can often lie in the range 0·5–5 mm h^{-1}. The overall mean of 26 studies reporting mean rain rates (Table III) is 3·47 mm h^{-1} (standard deviation 2·38 mm h^{-1}). While such low rates are challenging to simulate (discussed later), their widespread occurrence warrants greater attention, particularly in studies of processes like canopy interception, effects on soil surface properties, infiltration, and the fate of rain within ecosystems. More intense events, when they occur, generally last for only a small fraction of the total rain time. As will be shown later, much less experimental work has been done under low rain rates than under high or extreme rain rates. It is important to consider how this may have influenced our view of the significance of key environmental processes such as soil surface sealing or canopy interception on foliage.

Factors responsible for regional variation in rain event properties

The global patterns of daily rainfall are reasonably well known from a number of ground-based and remotely-sensed climatologies (Browning, 1990; Legates and Willmott, 1990; Hume, 1992). These have shown the high daily totals of the regions such as the zone of tropical convergence, and demonstrated the high spatial variability that is associated with mountainous terrain. In detail, there are regional variations in daily rainfalls whose origins are unclear. For example, Jackson (1988) has shown that northern Australia receives fewer rain days and higher mean daily rain rates than global means would suggest. This is explained partly by the prevalence of tropical lows and cyclones in this region, and Jackson notes that in terms of rain rates, a distinction can be made between cyclonic and non-cyclonic areas within the tropics. However, the controls on rain rates and event durations at sub-daily time scales are poorly known, partly owing to a lack of short period rain rate data from key regions such as the tropics (Jackson, 1988). Houze (1997) pointed out that tropical precipitation is the product of a continuum of modes of uplift from convective to stratiform, with the high rain rates of vigorous convection giving way to lower rates as the vigour of the uplift decays. Thus, tropical rain can include significant stratiform contributions at relatively low rain rates. Spatial and temporal variation in the occurrence of convective and stratiform rain have also been highlighted as controls on rain rates elsewhere. For example, for sub-Saharan Africa, Schumacher and Houze (2006) have shown that convective rain rates average 14 mm h^{-1} but stratiform rates average only 1·9 mm h^{-1}. In this region, convective activity that characterizes the pre-monsoon season is progressively replaced by stratiform rain during the monsoon. Globally, the behaviour of mesoscale convective systems exerts a key control on rain rates (Mohr *et al.*, 1999; Dai, 2001), but the influence of larger features such as planetary waves has also been shown. For example, Ding and Reiter (1982) demonstrated that blocking highs over East Asia can direct depressions southward over China, resulting in prolonged and heavy rainfall. Temporal variations in atmospheric circulation, teleconnections such as ENSO and local landsurface properties including topography, all drive variations in rain rate and duration. There remains much to be learned about the spatio-temporal behaviour of rain events in global environments (Dai *et al.*, 2007).

The relationship between rain event duration and event rain rate

As the results in Table II suggest, rain rates become higher when measured over shorter intervals of time. Thus, more intense bursts of rain tend to be relatively short-lived within longer rain events, and it would generally be expected that $R_{t1} > R_{t5} > R_{t10} > R_{t60}$ etc., within a particular long rain event.

However, this is a different issue from the relation between total event duration T_e and event rain depth, or, in other words, between event duration and mean event rain rate R_e.

A number of studies have demonstrated that in fact, longer events can exhibit higher mean rain rates than shorter events, as well as higher R_{t30} and R_{t60} rates, which is somewhat counter-intuitive. For example, this was established from rain event data in Palau (Kubota *et al.*, 2006), from the Baltic coast (Peters and Christensen, 2002), from a 49-year record from Barcelona (Vilar and Burgueño, 1995), and from Italy (Telesca *et al.*, 2007).

DOI: 10.1002/hyp

The trend is also shown in the rain rate, event depth, and duration data presented on 31 events by Hammad *et al.* (2006). Their results show that as rain event depth increased from about 5 mm to 50 mm, R_{t60} increased from about 2·5 mm h^{-1} to 10–12 mm h^{-1}. In parallel, event duration rose from <3 h to >18 h. Where it was explicitly examined, studies have suggested that the exponent in the relation between event rain depth D_e and event duration time T_e, $D_e \propto T_e^n$, lies in the range 1·2–1·7. Further data on this exponent would be valuable, although the results obtained are very much dependent on the definition of 'event', and on the length of rainless gaps that are permitted within an event. The relation seen in the 10 278 rain events analysed by Vilar and Burgueño (1995) shows great scatter, but nonetheless exhibits a positive correlation coefficient of 0·25 between D_e and R_e. Other data illustrate the reverse trend. For example, Drufuca and Zawadzki (1975) reported a correlation of −0·65 between D_e and R_e, from 10 years of Canadian tipping bucket gauge data. A stronger focus on the definition of 'event', and on the start, end, and dry gap conditions that define them, will be needed to extend the analysis of the connections between event duration and mean rain rate and to refine the embodiment of this relation into field experimentation using simulated rain. The criteria adopted in defining rain events are critical to inter-comparisons among published results, and need to be specified in all published work where event data are reported.

Distributions of rain event durations have been reported in a number of studies. There does not appear to be a widely-used classification of rain event durations; rather, events are often classified in terms of the depth of rain delivered in 24 h. As an example, Gochis *et al.* (2007) use the categories 'light event' (0 mm < depth < 10·0 mm), 'moderate event' (10 mm < depth < 50·8 mm) and 'heavy event' (>50·8 mm). Durations are not generally classified, but field data are widely available. For example, in Thailand, Dairaku *et al.* (2004) found that 95% of events lasted <5 h; over half lasted about 1 h. Only 5% of events lasted >6 h, but these long events contributed 30% of total rainfall. Balme *et al.* (2006) found that at Niamey, in the Sahel, 50% of the total annual rainfall was received from events lasting 3·9 h or less. In Malaysia, median rain event duration was reported as 1·38 h (Ziegler *et al.*, 2006). In Colombia, Tobon Marin *et al.* (2000) found that event durations lay in the range 20 min–13 h. In Japan, Haga *et al.* (2005) found event durations from 2–46 h. For Australian sites, Heneker *et al.* (2001) reported mean event durations as follows: Sydney 4·10 h (12 076 events); Melbourne 3·60 h (17 592 events); Brisbane 7·33 h (11 107 events). For Georgia USA, Bryant *et al.* (2005) reported event durations from 30 min–34 h, and a mean event duration of 2·65 h. For various field sites in Japan, Toba and Ohta (2005) reported mean event durations of 7·8 h–28·3 h. In tropical Puerto Rico, mean event duration was 3·57 h (Schellekens *et al.*, 2000). From 672 rain events in Ecuador, Fleischbein *et al.* (2005) established a mean duration of 9·15 h. In Brazil, Germer *et al.* (2005) found that 44% of rain events lasted <1 h. In a similar finding, Hjelmfelt (1978, in Lloyd, 1990) found that 50% of events lasted <1 h. Similar durations were reported by Cutrim *et al.* (2000).

These examples show that there is considerable variability in reported event durations. Clearly, some events last >1 day, while many last <1 h. At the short duration end of this range, the issue of how rain events are delimited becomes critical, and without a greater uniformity of approach, it is difficult to draw meaningful conclusions. Currently, there is a range of criteria in use for delimiting events. In work on canopy interception, for example, events are often only separated if there is a dry intervening time (typically 6 h) considered to be sufficient for the canopy to dry completely before the next rain event.

It would also be valuable to have data on inter-event times (the 'interludes' of Cutrim *et al.*, 2000), but these are also very dependent on the definition that is adopted for an event. One conclusion that will become clear later in this paper is that rainfall simulation experiments often target the short end of the distribution of event durations. Very rarely do simulated events last more than 1·5 h, and frequently, their duration is just 20–30 min.

The range of intensities within a rain event

Considerable variations in rain rate occur within rain events. The event for which Vilar and Burgueño (1995) reported $I_{\max} = 586$ mm h^{-1} lasted about 40 min, and had R_e of 57 mm h^{-1}, so that the $I_{\max}$ was about $10 \cdot 3 \times R_t$. A second example can be drawn from Jameson and Kostinski (2002), who reported a rain event having a mean rainfall rate of 12·85 mm h^{-1} but $I_{\max}$ of about 93 mm h^{-1} (so that $R_e = 7 \cdot 2 \times I_{\max}$), based upon 1-s data from a video disdrometer. Nine rain events recorded in an experimental watershed on the island of Honshu, Japan, showed R_e values averaging 3·18 mm h^{-1}, but with peak 30-min intensities averaging 12·0 mm h^{-1}. On average $R_{t30} = 5 \cdot 36 \times R_e$, but in individual storms the value reached $8 \cdot 24 \times R_e$ (Sidle *et al.*, 2005). Five days of heavy rain in Normandy, France, were reported by Cerdan *et al.* (2002). On 9 May 2000, for example, the mean event rain rate was 6·8 mm h^{-1}, but the maximum 6-min rate, R_{t6}, was 105 mm h^{-1}; thus $R_{t6} = 15 \cdot 4 \times R_e$. In five events, the mean ratio was $R_{t6} = 9 \cdot 4 \times R_e$. These instances and other suggest that within a rain event, it is reasonable to expect short bursts of rain approximately an order of magnitude more intense than the mean R value.

There is a growing realization that intra-storm intensity variations of the kind just noted can have a range of effects on hydrologic and erosional responses at the soil surface. The evidence comes primarily from rainfall simulation experiments, where pre-determined intensity variations can be incorporated into rain events. An example is provided by the rainfall simulation experiments on soil trays carried out by Frauenfeld and Truman (2004). In runs otherwise identical, they employed either constant rainfall rate (57 mm h^{-1}) or a variable rate delivering an intensity of about 160 mm h^{-1} at a peak in the second quartile of the storm duration, preceded by rising intensity and followed by declining intensity. Both treatments delivered the same total water depth. The rain event properties were based upon data for natural storms in the field area, but all events were held to a fixed duration of 70 min.

Although the rain event properties did not change total infiltration or surface runoff volumes, they did result in changes in the amount of soil lost from the sample trays. The changes were related to differences in the time and extent of surface seal formation among the steady or variable intensity events. Interrill erodibility was found to vary with the properties of the rain event.

Large $I_{\max}$ bursts within a rain event exhibiting a lower mean R are significant in terms of both overland flow production and sediment entrainment (Parsons and Stone, 2006). Likewise, studies are progressively demonstrating that rain rate fluctuations significantly affect nutrient and agrochemical runoff arising in rainfall simulation studies (Potter *et al.*, 2006; Frankline *et al.*, 2007; Truman *et al.*, 2007). Using simulated storms with a pattern of changing rain rate matched to average local storm characteristics, Truman *et al.* (2007) demonstrated that sediment and carbon losses in a variable rain rate event were considerably larger than arose in constant-rain-rate events of the same total rain depth. Runoff rates peaked earlier in the runoff event, and at a higher flow rate. Results such as these suggest that the use of constant rain rates thus has the potential to yield results that have reduced correspondence with what would arise under natural rain events. The distinction between $I_{\max}$ and mean R is important because of the widespread finding that large proportions of the total rainfall in an event are received within a small fraction of the event duration (this appears to apply regardless of event duration) (Huff, 1967). From rainfall records for Illinois, Huff (1967) introduced the classification of rain events into four groups, which he referred to as quartiles, depending upon whether the most intense rainfall occurred in the first, second, third, or fourth quarter of the rain event. (The most common events were those of first and second quartiles; fourth quartile events were the least common). Additionally, there may be no good relationship between rain event depth and the maximum rain rate within it. For example, in the nine rain events studied by Sidle *et al.* (2005), the maximum R_{t30}, 23·6 mm h^{-1}, was recorded in the rain event with the third smallest rain depth; the overall correlation was very low, with $r^2 = 0{\cdot}06$.

The environmental impact of rain events with low rain rates

The foregoing shows that although short, high rain rate intervals occur, the mean R_e is commonly far lower. It has been shown that soil erosion and runoff can be generated from rain of quite low R as well as from more intense events. For example, Fraser *et al.* (1999) report field data from cultivated fields in south-west England that establish that overland flow resulted from rain exceeding just 0·8 mm h^{-1}, and that rain of 0·8–1·0 mm h^{-1} caused suspended sediment losses of 14 kg ha^{-1} h^{-1}. More intense rainfalls ($R > 9$ mm h^{-1}) were associated with larger losses. Fraser *et al.* (1999) pointed out that while uncommon, heavy rains caused the export of considerable amounts of phosphorus, but noted that the contribution made by less intense but longer events was also significant. Palestinian plot studies by Hammad *et al.* (2006) showed that runoff and soil erosion were both likely to occur in rain events in which $R_{t30} > 4$ mm h^{-1}, quite a low rate. Low rain rate thresholds for overland flow and erosion have been reported elsewhere (Hussein and Othman, 1988; James and Alexander, 1998; Kumar *et al.*, 2002). For erosion on burned hillslopes, Robichaud *et al.* (2006) reported significant erosion for events in which R_{t10} exceeded 13 mm h^{-1}, and for which R_e values (not reported) would have been considerably lower. Moderate rain rates can also result in flash flooding and associated landscape change. For example, flash flooding in Morocco has been shown to arise from rain events with peak R_{t30} rain rates of just 16 mm h^{-1} (and less on an hourly basis) (Fink and Knippertz, 2003). Similarly, under post-fire conditions, extraordinary runoff conditions can result in flooding from rain rates with peak R_{t30} values equivalent to only 20–32 mm h^{-1} (Moody and Martin, 2001), which is heavy rain but not exceptional. On the other hand, from the post-fire landscape, Moody and Martin (2001) report specific discharges of up to 50 m^3 s^{-1} km^{-2} or 100 m^3 s^{-1} km^{-2}, which are truly exceptional, signalling highly efficient overland flow generation and conveyance from heavy rain.

The commonness of events having low rain rates raises a number of interesting issues. How is the production of overland flow or the entrainment of soil distributed along the spectrum from small to large rain events? This is a question of magnitude–frequency analysis. Marques *et al.* (2007) have commented on this. They pointed out that low rain rate events are much more common than extreme rain rate events, and therefore, that their cumulative erosional consequences should not be overlooked. De Ploey *et al.* (1991) approached this issue via the calculation of an index of 'cumulative erosion potential', and illustrated their analysis using data from several locations in Europe and East Africa, while Bergkamp *et al.* (1999) attempted a magnitude–frequency analysis of water redistribution on hillslopes in Spain. There is likely to be geographic variation in the position in the magnitude–frequency spectrum where rain events achieve most soil dislodgement, production of overland flow, etc. Nicolau (2002), for example, investigated overland flow on artificial slopes in a Spanish coalfield. Because the materials displayed low infiltration capacities, most rain events resulted in overland flow, whose volume was therefore more dependent on rainfall depth than rain rate. On highly permeable materials, where only heavy events could trigger overland flow, these events would dominate. Thus, it has to be remembered that variations in soil type, surface crusting, plant and litter cover, etc., may result in larger shifts in the dominant rain event magnitude than does the changing rainfall climate.

A second question is how the success of magnitude–frequency analyses applied to erosion or the production of surface runoff depend on the definition adopted for rain event or the integration time used for tallying rain rates. In other words, how would explanatory power vary if instead of adopting daily (24 h) rain depths and mean rain rates, as done by de Ploey *et al.* (1991), analysis was based on I_{t5}, I_{t30}, I_{t60}, etc? For a site in New Mexico where rain was delivered by short but intense convective storms, Hastings *et al.* (2005) concluded that intense intra-event bursts of high rain rate, lasting perhaps 1–2 min, may drive both surface runoff and erosion. In this situation, soil loss might best correlate with a measure like I_{t5}. This is an area where additional analyses are required, and it is clear that the use of daily rainfall records to stand for rain events will not prove successful in such cases. Controlled experiments using rainfall simulation may provide a valuable tool here, but will require that the emphasis on high rain rates be moderated, with additional systematic exploration of processes in lower rain rate events. For example, using rainfall simulation combined with erosion monitoring on the same field plots under natural rain, Nolan *et al.* (1997) found better agreement between the natural and simulated rain soil loss when using a 'low' (60 mm h^{-1}) rain rate than when simulated rain was applied at a 'high' rate of 140 mm h^{-1}. It would appear that research has not yet answered some of the questions raised here, and doing so will require a greater awareness of the range of natural rain event properties, particularly in experimental work using rainfall simulation. This topic is addressed next.

RAINFALL SIMULATION

Rainfall simulation provides a convenient tool that allows rain events of nominated properties to be created. In hydrologic and soil erosion studies, it often proves desirable to extend field observations made under natural rain in a controlled way, through the use of simulated rain. Simulation studies may be carried out in the field or in the laboratory. They offer a number of well-known advantages, including the ability to create replicate rain events of identical properties, while varying another parameter of interest, such as foliar cover or soil type, and to control arbitrarily event properties such as intensity and duration. However, several authors have pointed out that key rain event properties employed in rainfall simulation lack complete correspondence with the comparable measures of natural events (Agassi *et al.*, 1999; Mathys *et al.*, 2005; Kinnell, 2005). There is a wide range of approaches as well as spatial scales of experimentation. For field plots, for example, areas range from <0·1 m^2 (Casermeiro *et al.*, 2004) to >300 m^2 (Lusby and Toy, 1976). Experimental rain event durations range from <5 min (Casermeiro *et al.*, 2004) to 6–9 h (Armstrong *et al.*, 1998; Scherrer *et al.*, 2007). Given the wide familiarity of the technique and approach, further details are not included here. Apparatus and methods have developed from early work where hand-operated watering cans were used (Costin and Gilmour, 1970), and new designs continue to appear (Clarke and Walsh, in press).

Rain event properties in rainfall simulation studies

An inspection of the published literature shows that the distribution of rain rates used in rainfall simulation studies (Table IV) more closely resembles the extreme rates of Table II, than the more common lower rain rates of Table III. Some quite extreme rain rates have been adopted, including 838 mm h^{-1} (Scarbrough *et al.*, 2004). Other sustained high rain rates include 227 mm h^{-1} for 30 min, adopted to ensure runoff generation from plots in Utah (Barger *et al.*, 2006). This rain event was estimated to be a 500–1000 year event for this area, although considerable uncertainty must attach to such estimates. The use of extreme rain event properties is partly accounted for by the common means of generating artificial rain using spray nozzles operating under a water pressure designed to yield realistic drop sizes and drop energy values. However, this results in rain rates in simulations typically exceeding those of natural rain (Hanke *et al.*, 2004), unless marked intermittency is introduced by using oscillating or pulsed nozzles. The rainfall simulation literature appears to be dominated by studies that have R set to 60–100 mm h^{-1}, and in some cases even higher, which would fall into the 'exceptional' or 'torrential' categories mentioned earlier, and certainly beyond the 'extreme' category of Tokay and Short (1996). Examples include the work of Liu *et al.* (2004), who used R of up to 200 mm h^{-1}, and Schiettecatte *et al.* (2005a) who adopted 188 mm h^{-1}. The average of the maximum rain rates adopted in 49 experimental studies listed in Table IV is 103·1 mm h^{-1} (standard deviation 81·3 mm h^{-1}). This exceeds the mean of exceptional natural rain rates (Table II), and is nearly 30 times higher than the overall mean rain rate of the non-exceptional events (Table III). There is thus little doubt that these are extreme rain rates when judged against many natural events. Often, however, the drop size and kinetic energy for simulated rain fail to reach the levels typical of natural rain. Thus, the erosivity of the high rain rates in simulated rain would not be as high as natural rains at the same rate (Marcos *et al.*, 2000; Kuhn, 2007).

High rain rates are often used in combination with simulated rain events lasting several h, but still not matching the durations of many natural events. The

Table IV. Compilation of rain rates from 40 published studies employing rainfall simulation. See text for comments

Location	Rain rates (including integration time)	Source of data
Laboratory (South Africa)	45 mm h^{-1} for 60 min	Mills and Fey (2004)
Laboratory (Israel)	36 mm h^{-1} for fixed 60 mm depth	Mamedov *et al.* (2002)
Senegal	70 mm h^{-1} for 30 min	Ndiaye *et al.* (2005)
California	43 mm h^{-1} for 2·5 h	Brady *et al.* (2006)
Laboratory (France)	40 mm h^{-1} and 70 mm h^{-1}	Cousin *et al.* (2005)
Australia	80–100 mm h^{-1}	Herngren *et al.* (2005)
Spain	30–117·5 mm h^{-1} for 30 min	Arnaez *et al.* (2007)
Germany	36 72 mm h^{-1} for 15–30 min	Biemelt *et al.* (2005)
France	60 mm h^{-1} for 1 h	Le Bissonnais *et al.* (2007)
Australia	8 mm h^{-1} and 80 mm h^{-1} for 20 min	Dougherty *et al.* (2004)
Laboratory (USA)	57 mm h^{-1} for 70 min (plus intensity changes in variable storm)	Frauenfeld and Truman (2004)
Tunisia	73 mm in 70 min (= 62·5 mm h^{-1}) variable intensity; maximum $I_5 = 110$ mm h^{-1}	Hamed *et al.* (2002)
Laboratory (Germany)	0·5 mm h^{-1}; 5 mm h^{-1}; 48 mm h^{-1}	Hunsche *et al.* (2007)
Ecuador	20 mm h^{-1}, 27 mm h^{-1}, 50 mm h^{-1}, 70 mm h^{-1}, 90 mm h^{-1}, 120 mm h^{-1} (composite storm: 15 min at each intensity)	Poulenard *et al.* (2001)
Spain	60 mm h^{-1} and 140 mm h^{-1} for 45 min	Tejada and Gonzalez (2007)
Spain	120 mm h^{-1} for 2 h	De Luis *et al.* (2004)
Laboratory (USA; Pacific Northwest)	20 mm h^{-1}, 60 mm h^{-1}, 250 mm h^{-1}, 420 mm h^{-1} applied successively	Keim *et al.* (2006)
Liguria, Italy	Fixed 76 mm h^{-1} for 60 min	Rulli *et al.* (2006)
Uruguay	Fixed 70 mm h^{-1} for 20 and 45 min	Victoria *et al.* (1997)
Laboratory (USA)	68 mm h^{-1} for 1 h	Green *et al.* (2000)
Laboratory	30 mm h^{-1} for 3 h	Römkens and Prasad (2006)
Laboratory	75 mm h^{-1} for 2 h and 150 mm h^{-1} for 1 h	Zhou *et al.* (2003)
Waikato New Zealand	60–111 mm h^{-1} for total depths of 120 mm and 190 mm	Müller *et al.* (2004)
Switzerland	50–100 mm h^{-1} for 3–6 h	Scherrer *et al.* (2007)
India	48·5–136·8 mm h^{-1} for fixed 50 min event	Mandal *et al.* (2005)
Laboratory (Greece)	28 mm h^{-1} for 4 h	Dimoyiannis *et al.* (2001)
Spain	40 mm h^{-1} for 30 min	Seeger (2007)
Tunisia	200 mm h^{-1} for 8–10 min	Schiettecatte *et al.* (2005b)
USA (New Mexico; Colorado)	97–440 mm h^{-1}	Martin and Moody (2001)
Northern Ethiopia	90–120 mm h^{-1} for 10 min	Aerts *et al.* (2006)
Ecuador, Appalachians, Puerto Rico	100 mm h^{-1} for 30–35 min	Harden and Scruggs (2003)
Canada (Prairies)	100 mm h^{-1} for 1 h	Martz (1992)
United States (Indiana)	100 mm h^{-1} for 30 min	Smith *et al. (* 2007)
United States (New Mexico)	100 mm h^{-1} and 133 mm h^{-1}, duration not stated	Neave and Rayburg (2007)
Australia (Victoria)	100 mm h^{-1} for 30 min	Sherdian *et al.* (2007)
Spain	80 mm h^{-1} for 40 min	Ramos and Martínez-Casanovas (2006)
Laboratory (Canada)	40 mm h^{-1} and 100 mm h^{-1} for 2 h	Mermut *et al.* (1997)
Laboratory (United Kingdom)	46·4–170·8 mm h^{-1} for 35 min	Parsons and Stone (2006)
Laboratory (Hawaii, USA)	65 mm h^{-1} for 1 h	Wan *et al.* (1996)
Tunisia	200 mm h^{-1} for 15 min	Schiettecatte *et al.* (2005b)
Laboratory (China)	188 mm h^{-1} for 30–60 min	Schiettecatte *et al.* (2005a)
Laboratory (China)	100 mm h^{-1} for 70 min	Pan and Shangguan (2006)
Laboratory (Australia)	97–124 mm h^{-1} for 30 min	Asadi *et al.* (2007)
Laboratory (Australia)	100 mm h^{-1} for 20 min	Loch and Foley (1992)
Laboratory (China)	90 mm h^{-1} for 1 h	Zhou and Shangguan (2007)
Laboratory (United Kingdom)	5 mm h^{-1} for up to 9 h	Armstrong *et al.* (1998)
Laboratory (France)	40 mm h^{-1} and 70 mm h^{-1} for up to 20 h	Issa *et al.* (2004)
Laboratory (Hawaii)	12·9–107 mm h^{-1} for 40 min	Watung *et al. (* 1996)
Laboratory (France)	25–150 mm h^{-1} for 45 min	Huang *et al.* (2001)
Laboratory (United Kingdom)	3–12 mm h^{-1} for 1–2 h	Holden and Burt (2002)

3–6 h events of Scherrer *et al.* (2007) were noted above. Other instances include 102 mm h^{-1} sustained for 3 h (Ziegler and Sutherland, 1998), and 80 mm h^{-1} for 1·5 h (Polyakov and Lal, 2004). On the other hand, some workers adopted very brief events, such as the 5 min erosion tests at 56·6 mm h^{-1} used by Casermeiro *et al.* (2004) to judge relative protection against erosion afforded by various kinds of plant.

The selection of R and T for rainfall simulation experiments

What explanations can be found for the high rain rates that dominate rainfall simulation studies? Several observations arise from the literature, and are considered briefly here (refer to Table V).

Partial correspondence with natural events

In some cases the intensity and duration are adopted to correspond with extreme natural rain events of known recurrence interval. For example, the valued adopted by Pan and Shangguan (2006) was adopted to correspond to an event having a 10-year return period. Le Bissonnais *et al.* (2007) adopted R of 60 mm h^{-1} for 1 h, noting that events of such size were frequent at their French field sites. Victoria *et al.* (1997) reported that their adoption of $R = 70$ mm h^{-1} was guided by their knowledge of natural rain events in Uruguay, but also influenced by the values of R that others had adopted in simulation experiments. Mandal *et al.* (2005), in India, adopted $R = 48{\cdot}5$ mm h^{-1}, 89·2 mm h^{-1}, and 136·8 mm h^{-1} to correspond with local rain rates of 1, 2 and 20 year ARI (average recurrence interval), but employed a fixed event T of 50 min, which removes some of the correspondence with natural events. Similarly, Kitchen and Goonetilleke (2007) adopted rain rates of 65 mm h^{-1}, 86 mm h^{-1}, 115 mm h^{-1} and 156 mm h^{-1} to correspond with local events having 1, 2 and 10 year ARI at the durations used. In a similar way, Wallbrink and Croke (2002) adopted 45 mm h^{-1}, 75 mm h^{-1}, and 110 mm h^{-1} to correspond with 2, 10, and 100 year ARI events in south eastern NSW, Australia. In Liguria, Italy, Rulli *et al.* (2006) adopted $R = 76$ mm h^{-1} for a fixed 1 h event, noting that in this area, the 20 year ARI, 1 h event, has an intensity of 100 mm h^{-1}, which exceeds the value of R adopted. There is a tendency in rainfall simulation studies to allow rain event duration to be set largely without regard to natural storm durations. For example, Arnaez *et al.* (2007) made a study of erosion in Spanish vineyards, and based their experimental rainfall intensity on an analysis of 25 years of local rainfall records. They varied R under simulated rain from 30 mm h^{-1} to 117·5 mm h^{-1}, but all events were of fixed 30 min duration. This still involves a

Table V. Examples of the drivers of simulated rain event properties from 15 published studies using rainfall simulation as a research tool. See text for comments

Experimental rain rate and event duration	Rationale for adoption of rain event properties	Source of data
60 mm h^{-1} for 1 h	In order to trigger incision of rills (laboratory)	Gómez and Nearing (2005)
100–105 mm h^{-1} for 25–60 min	In order to generate highly sediment-laden flows from Marl slopes, France	Oostwoud Wijdenes and Ergenzinger (1998)
100 mm h^{-1} for 30–35 min	In order to generate surface runoff	Harden and Scruggs (2003)
140 mm h^{-1} for 30 min	In order to reveal different hydrologic behaviour or shrub and non-shrub plots	Schlesinger *et al.* (1999)
70 mm h^{-1} for 45 min	To generate surface runoff	Victoria *et al.* (1997)
188 mm h^{-1}	To ensure soil consolidation during duration of experiment	Schiettecatte *et al.* (2005a)
50·8 mm h^{-1} for 45 min	To yield sufficient erosion for differentiating among responses at experimental sites	Lusby and Toy (1976)
97–440 mm h^{-1} for 15–20 min	To exceed soil infiltration rate, and allow this to reach steady state	Martin and Moody (2001)
50–70 mm h^{-1} for 1 h	To ensure that runoff was generated	Elliott *et al.* (2002)
102 mm h^{-1} for 3 h	High rain rate plus long rain duration to ensure that whole plot length contributed to interrill runoff and sediment transport.	Ziegler and Sutherland (1998)
12·5 mm h^{-1} (though study area mean rain rates were known to be 3–4 mm h^{-1})	To maximise the runoff from plots, and yield sufficient runoff water for chemical analysis	Heathwaite and Johnes (1996)
203 mm h^{-1} for 30 min	So that all plots would reach terminal infiltration rate	Thurow *et al.* (1988)
227 mm h^{-1} for 30 min (estimated to be a 500–1000 year event)	To ensure that runoff was shed from all plots	Barger *et al.* (2006)
120 mm h^{-1} for 15 min (though max recorded R_{t15} for study area was 61 mm h^{-1})	To be consistent with prior research and to ensure adequate volumes of runoff	Deluca *et al.* (1998)
64 mm h^{-1}, durations not reported	Sufficient time for runoff rates to stabilise	Bertol *et al.* (2003)

degree of unreality, since rain events with a peak intensity of say 100 mm h^{-1} would undoubtedly involve periods of lower intensity during which the soil surface could drain, whereas lower intensity natural events would probably have been longer than 30 min, as noted earlier.

Purposefully high rain rates to generate measurable effects. Sometimes high rainfall rates are purposefully chosen not primarily to mimic natural events, but rather to achieve a measurable result (Table V). For example, Harden and Scruggs (2003) deliberately used high simulated rain intensities to exceed the infiltration rates of soils and so generate runoff. Schlesinger *et al.* (1999) used simulated rain at 140 mm h^{-1} for 30 min, an intensity that they note is not representative of rain in their New Mexico study area, having not been reached even by the R_{1t} of natural rain events. They used such intense rain in order to differentiate shrub and non-shrub plots. In a similar way, Victoria *et al.* (1997) adopted $R = 70$ mm h^{-1} expressly to generate surface runoff from the soils that they were studying. Müller *et al.* (2004) adopted $R =$ 60–111 mm h^{-1} (noted to greatly exceed corresponding 10-year natural R values) in order to ensure measurable surface runoff. Schiettecatte *et al.* (2005a) adopted the extreme R of 188 mm h^{-1} in their experiments with the express objective of ensuring that the soils tested were fully consolidated by the experimental treatments. However, under unnatural conditions, the value of the experimental findings to landscape evolution seems to be called into question if natural rain events do not give rise to the effect shown in the rainfall simulation work. Despite this, some workers who adopt very high rain rates still argue specifically that their conclusions can be extrapolated to the natural processes in the landscape (Müller *et al.*, 2004). In this context, it is interesting to note that some hydrologic effects are restricted to events with higher rain rates. An example is the reduction in runoff and soil loss from small plots containing shrubs (Bochet *et al.*, 2006). In this Spanish study, differences in runoff fraction among different shrub taxa were only manifested above a rain rate of 20 mm h^{-1}. Moreover, some authors have demonstrated how using too high an intensity can lead to erroneous estimations of key hydrologic properties (see Léonard *et al.*, 2004 for comments on this in relation to the role of biopores in modifying soil water uptake rates).

Table V indicates that in a number of studies, considerations of statistical analysis, runoff water sample volumes needed for analysis, need to ensure runoff from all sites in comparative studies, and similar constraints, often drive the selection of rain event properties. Even if operational considerations underlie the choice of rain event properties, the position of these in the spectrum of natural rain events in the study area should still be reported.

Other reasons. Sometimes rain rates are chosen that are known by the researchers to exceed corresponding natural rates, without a clear rationale being offered. For instance, Watung *et al.* (1996) used R up to 107 mm h^{-1} for 1–2 h, a value that they reported would not be matched even in a 100-year ARI, 1 h event. Similarly, in a study of vineyard soil erosion, Battany and Grismer (2000) adopted a fixed 60 mm h^{-1}, 40-min storm, despite noting that rain events in the area are of 'relatively low intensity' and that a 100-year, 40-min event would deliver less than half this rain amount. Similarly, Chaplot and Le Bissonnais (2003) adopted $R = 30$ mm h^{-1} despite their field data only including $R < 8$ mm h^{-1}, and generally $R < 2$ mm h^{-1}. Scherrer *et al.* (2007) used both high rain rates and long event durations, which they acknowledge to lie beyond 1000 year ARI for their field area. Presumably, in some of these studies, the choice of R was constrained by the available apparatus for rain simulation.

However, in many cases, no justification is provided for the selection of either the applied rain intensity or the rain event duration (Dimoyiannis *et al.*, 2001; Mills and Fey, 2004; Ndiaye *et al.*, 2005; Hawke *et al.*, 2006). Indeed, this points to one of the issues raised by many studies using simulated rain: the intensity and event duration appear to be selected independently of each other, without regard for the natural values of the parameters, or the fact that in natural rain events the two are correlated. For example, Vilar and Burgueño (1995) analysed 10 278 rain events, and found the average duration to be about 30 min. However, the distribution of event durations was strongly positively skewed, and the most common class of events was those that lasted from about 10 to 15 min. Likewise, duration and average rain rate were strongly correlated, not independent. Events of long duration tend to have higher total depth and higher R. These are examples of rain event properties that in many cases related to the climatological and meteorological situation of the field sites concerned. Reunion Island, referred to earlier, is renowned for its intense and long-duration rains. Other rain systems, such as dryland convective storms, are typically of shorter duration, but probably have a range of regionally distinct intensity–duration–spatial scale properties. Mid-latitude convective rain cells commonly involve heavy rain over spatial scales of only 5–10 km, and lie within broader regional rain fields. They last typically 30 min, and have high central rain rates (up to 50–100 mm h^{-1}), which decline exponentially with distance from the cell centre (Rebora and Ferraris, 2006).

Arbitrary extension or truncation of rain duration is often done to satisfy data requirements, e.g. to allow runoff rate to stabilize, or for a similar experimental objective. For a given rain rate, duration equates to event size (total rain depth). In relation to the experimental study of a phenomenon like surface runoff, event size is clearly relevant to the filling of surface depression stores, and so, to the time to initiation of surface runoff, and the fraction of total event time for which surface runoff was active. These parameters in turn would undoubtedly affect measured soil or solute loss. Therefore, experiments (particularly those where there is an intention to quantify and interpret soil losses) must be designed with both rain rate and rain duration in mind, and ideally, both should be scaled using relationships defined for the field area in question in the research. Only in this way will it be possible to attach the appropriate level of significance to the experimental findings. This point has recently been emphasized by Marques *et al.* (2007).

OTHER SIGNIFICANT RAIN EVENT PROPERTIES

Although intensity, drop size, and kinetic energy properties of real and simulated rain have been widely discussed in the literature, there are other properties that seem generally to be neglected. An example is the density of drop impacts per unit area at the surface, which Hosking and Stow (1987) called the 'raindrop arrival rate'. This is not among the parameters usually listed as relevant to the design of rainfall simulation apparatus and experiments, and was not mentioned in reviews by Riezebos and Seyhan (1977) or Agassi and Bradford (1999). Detailed studies of the shear forces generated by rapid flows ('lateral jets', Ghadiri and Payne, 1981; Terry, 1998) displaced outward from drop impact points have shown that these may be critical in overcoming the rupture strength of soils. Consequently, the density of drop impacts may play a role in lateral jet behaviour via mechanisms such as the destructive interaction of lateral jet flows when adjacent drop impacts are very close in space and time. An additional concern arises therefore in relation to the kind of rainfall simulation apparatus used, particularly for studies whose focus is soil erosion and flow mechanics. Dripformers, for example, may release drops that repeatedly fall to the same ground point below, so that parts of the soil surface escape drop impact. Fans or mechanisms to move the dripper system or the sample tray may be employed to diminish this effect (Agassi and Bradford, 1999). Spray systems seem much more likely to generate drops that would arrive at all parts of the soil surface below the spray nozzles, but it remains unclear whether drop arrival rate is realistically reproduced in this case.

An example of experimental work using rainfall simulation where drop impact density is relevant is provided by the investigation of the effect of rainfall on the properties of shallow sheetflows (Guy *et al.*, 1990). This study used quite high simulated rain rates (nominally 45 mm h^{-1}, 140 mm h^{-1}, and 180 mm h^{-1}) generated by a single nozzle sprayer. The interaction of arriving drops with the shallow overland flow, and the conversion of the vertical trajectory of the water drops to downslope flow, seem very likely to be influenced by the areal density of drop impacts. This would modify the mean flow path distance over which the flow could accelerate following a drop impact, for example, before another impact was experienced. To the author's knowledge, there are no existing data on this potential effect, and published data should be interpreted with this in mind. Other studies of overland flow behaviour have also neglected the potential role of drop arrival rate (Liu *et al.*, 2004). Some experimental results where drop arrival rate was considered as an explanatory variable are presented below.

Natural and experimental raindrop arrival rates

Various kinds of optical imaging-based and disdrometer data have provided data on raindrop arrival rates at various locations. Using an optical disdrometer with a 5 cm diameter orifice, Hosking and Stow (1987) found raindrop arrival rates from a few hundred to several thousand drops $m^{-2}\ s^{-1}$ in rainfall near Auckland, New Zealand. Using a raindrop camera, Smith and De Veaux (1992) reported raindrop arrival rates at Coweeta, North Carolina, in the range $<50\ m^{-2}\ s^{-1}$ to $>5000\ m^{-2}\ s^{-1}$, with considerable short-term variation during rain. Smith and Krajewski (1993) reported additional data from North Carolina, confirming drop arrival rates of 100–5000 $m^{-2}\ s^{-1}$. With the same optical raindrop camera, Smith and De Veaux (1994) reported data from several parts of the USA, as well as from the Marshall Islands. Their drop arrival rate data range from around 400 $m^{-2}\ s^{-1}$ in Alaska and Oregon, to nearly 1800 $m^{-2}\ s^{-1}$ in the Marshall Islands. Foley and Silburn (2002) presented calculated drop arrival rate data based on a generalized relation established by Mueller (1962) in Florida:

$$\text{Drop arrival rate} = 50R$$

where drop arrival rate is in drops $m^{-2}\ s^{-1}$ and R is in mm h^{-1}. This yields a raindrop arrival rate of 250 $m^{-2}\ s^{-1}$ at 5 mm h^{-1} and 1500 $m^{-2}\ s^{-1}$ at $R =$ 30 mm h^{-1}.

Drop arrival rate under simulated rain

Foley and Silburn (2002) successfully simulated the Mueller (1962) raindrop arrival rates quite well using an oscillating spray nozzle system, and less well using a drop-forming rainfall simulator with drippers set out on a 25 mm grid. In laboratory experiments with soils, they demonstrated that infiltration rate and crust permeability were both influenced by raindrop arrival rate, and in some cases this parameter provided higher explanatory power than did rainfall rate R. Foley and Silburn (2002) speculated that drop arrival at the surface might drive infiltration partly by hydraulic pressure created at locations struck by arriving drops. If this is correct, then greater attention should be given to raindrop arrival rate, especially perhaps on exposed agricultural soils. However, no other studies using simulated rain appear to have reported data on the drop impact frequencies that were involved. Nevertheless, this is an important aspect of rain viewed as a point-process at small spatial scales (Uijlenhoet and Sempere Torres, 2006) that deserves wider attention, especially when simulated rain is used in erosion experiments. With spray-nozzle rainfall simulation systems, it is not easy to estimate the areal density of drop impacts, but there are methods by which this can readily be done, including an adaptation of the flour pellet method, filter paper method, and disdrometers.

Dunkerley (2008) reported a study of canopy interception in dryland shrubs made using simulated rain from a grid of drop-forming modules. His apparatus resulted in areal drop impact frequencies of about 56 impacts s^{-1} m^{-2} at 10 mm h^{-1}, increasing to about 104 impacts s^{-1} m^{-2} at 20 mm h^{-1}. These results are comparable with the lower end of the Coweeta data presented above, but notably lower than the frequencies for more intense rain at Miami, Florida, reported by Foley and Silburn (2002). The areal density of drop impacts under simulated rain from a drip-forming apparatus seems likely generally to be lower than the areal density under natural rain (Foley and Silburn, 2002), because drip-formers produce water drops of fixed size, while rain of the same rain rate distributes the equivalent water volume into a range of drop sizes, including many small droplets. (Whether this is a significant difference is not clear, as the more abundant small droplets would have smaller impact energy. Thus, dripper module simulated rain might well be more erosive than natural rain of the same rain rate but higher drop arrival rate).

Given that the raindrop arrival rate is to some extent fixed by the geometry of drop-former based rain simulator apparatus, it is worthwhile noting some of the design details of these devices. Published studies (Table VI) reveal a significant range of areal density of drip-formers, ranging from 400 drippers m^{-2} to >4000 drippers m^{-2}. The density of drip points appears to be a function of the apparatus available, or what is convenient to manufacture, and appears not to have been motivated by a concern to correctly reproduce the raindrop arrival rates of natural rain. However, as indicated in the data from Dunkerley (2008) presented above, the grid spacing of drop formers tells only a part of the story: the other parameter that must be known (and controlled) is the rate of drop release. The drop release rate may only be ~1 per 10 s, so that the drop arrival rate at the surface might be only 10% of the value suggested by the spacing of the drop-formers. For studies using drop-formers, it should be straightforward to collect the relevant data by timing the interval between drop releases from a small sample of drippers, but this has generally not been done in published studies. For example, Blackburn *et al.* (1974) used a density of about 1494 drip-formers m^{-2} in their mobile infiltrometer, but the rate at which each drip-former released a water drop was not reported, so that the drop arrival rate cannot be estimated. Liu *et al.* (2004) used drip-formers to create rain rates of 15–200 mm h^{-1} but do not specify areal drop arrival rate or spacing of drippers. In neither of these studies was the rate of droplet release from the drippers reported, and therefore once again, the droplet arrival rate at the ground cannot be worked out. Clearly, even with this relatively controllable type of rain simulator, more attention to drop arrival rate is needed. In the study by Dunkerley (2008), it was reported that the drop release rate varied with simulated rain rate. Drop release occurred from drippers every 14·9 s at 10 mm h^{-1}, but drops were released every 7·9 s at 20 mm h^{-1}. Data of this kind are needed so that the results from simulation studies employing drop-formers can be inter-compared. At the same time, there seem to be sound reasons for considering that drop arrival rates ought to reflect those of natural rains.

Table VI. Details of the drop-forming array from 18 studies where rainfall simulation was used as a research tool. See text for comments

Description of dripper array (density of dripper array, plus grid layout, if this was specified)	Source of data
4789 m^{-2} (91 drippers over 190 cm^2)	Harden and Mathews (2000)
2500 m^{-2} (on a 2 cm × 2 cm grid)	Roth and Helming (1992)
2222 m^{-2} (on a 2 cm × 2 cm grid)	Shainberg *et al.* (2003); Mamedov *et al.* (2000)
1600 m^{-2}	Hignett *et al.* (1995)
1500 m^{-2} (on 2·5 × 2·5 cm grid)	Kinnell (2005)
1500 m^{-2} (on a 2·5 × 5 cm grid)	Dunkerley (2008)
1494 m^{-2}	Blackburn *et al.* (1974)
1456 m^{-2}	Wan *et al.* (1996)
1344 m^{-2}	Wilson and Seney (1994)
1322 m^{-2} (on a 2 cm × 2 cm grid)	Abu-Zreig (2006)
1254 m^{-2}	Foster *et al.* (2000)
1131 m^{-2} (on a 2 cm × 2 cm grid)	Clarke and Walsh (2007)
1077 m^{-2}	Foster *et al.* (2000)
1024 m^{-2} (256 drippers over 0·25 m^2)	Joel and Messing (2001)
1000 m^{-2}	Heathwaite and Johnes (1996)
725 m^{-2}	Cerda and García-Fayos (2002)
400 m^{-2} (on a 5 cm × 5 cm grid)	Zhang *et al.* (2000)

DISCUSSION AND CONCLUSIONS

The foregoing review has demonstrated that rainfall simulation studies have often overlooked ordinary rain event properties and instead have tended to adopt extreme rain rates. Only a very few studies have adopted rain rates <10 mm h^{-1} (Table IV) and all of these have been laboratory experiments. Although low rain rates may be more challenging to simulate, their widespread occurrence warrants greater attention, particularly in studies of processes like canopy interception, the alteration of soil surface properties, infiltration, and the fate of rain within ecosystems. More intense events, when they occur, generally last for only a small fraction of the total rain time. It seems probable that the bias toward extreme rain rates has influenced our sense of their role in key processes such as soil surface sealing or canopy interception on foliage. Certainly, rainfall simulation provides a valuable tool for exploring processes in extreme events, which may be more difficult to study otherwise, owing to their relative rarity. Much has been learned from well-designed experimental work with simulated rain, and there is no doubt that extreme rain events sometimes yield dramatic effects on the landscape. This does not, however, diminish the argument that we should seek a balanced understanding of the nature and effects of rain events across the whole range of event magnitudes and frequencies.

The technical challenges in exploring processes in events of low rain rate is significant. There will need to be new approaches to generating rain rates of 1 mm h^{-1} to 10 mm h^{-1} while at the same time correctly reproducing the range of drop sizes, kinetic energy, and drop arrival rate. The foregoing review has highlighted other rain event properties that warrant more attention: the intra-event variation in rain rate (and there may be associated changes in drop size and energy, which will pose additional technical challenges for simulation apparatus), and the proper linking of event duration and event rain rate.

Two recommendations arise from the widespread relevance of storm and rain properties to environmental processes, some instances of which were mentioned earlier:

(1) It is vital to analyse and report the relevant storm properties, whether in natural or simulated rain, when accounting for observed patterns of soil loss, nutrient loss, overland flow, etc. Only in this way can the relative roles of storm and soil properties be disentangled, when analysing the controls on soil loss from time-series records gathered from experimental plots. In experimental studies where rain event properties are of relevance to the interpretation of results, all key properties (event definition used, rain rate, fluctuations in rain rate event duration, and drop arrival rate) should be reported wherever possible.

Only by focusing more attention on rain events and their properties will we be able to develop a fuller understanding of how rain events should be identified to yield the most explanatory power in studies of streamflow generation, soil erosion, and other processes. The foregoing discussion presented examples showing that for some purposes, high resolution is needed in rainfall data processing (e.g. to reveal the effect of short, intense bursts on soil dislodgment). To understand the generation of streamflow, where hillslope and channel routing of flow attenuate such short bursts of rain, longer integration times in the analysis of rain rates are probably more appropriate. More research focusing on the significance of rain event properties and their links to landscape processes seems likely to yield valuable insights. It may be that no single criterion for recognizing a rain event will prove sufficient; rather, different criteria may emerge from studies addressing particular hydrologic or geomorphic processes.

(2) One primary reason for adopting the use of rainfall simulation as a research tool is to reproduce in a controlled way the behaviour expected in the natural environment. This has been widely noted in the literature on rainfall simulation, but with a notable emphasis on the correctness of raindrop size distributions, kinetic energy delivery per mm of rain depth, etc. Less attention appears to have been paid to correctly reproducing other event properties, including duration, mean rain rate, and the temporal pattern and magnitude of rain rate fluctuations. Other properties seem to have received little attention, including the density of droplet impacts per unit area and unit time ('raindrop arrival rate') at the soil surface. A secondary application of rainfall simulation is the exploration of mechanisms without the intention to represent the rain event characteristics of any particular region. This was the approach taken by Pan and Shangguan (2006), for example, in a generalized laboratory study of the role of grassy vegetation in affecting soil loss from sloping soil pans. However, even where general principles are being explored, the results have diminished value if the imposed rain event properties do not lie within the range commonly experienced at field sites where the results are intended to find application. Thus, researchers need to have a working knowledge of rain event characteristics in their study area and elsewhere, even in the exploration of general principles. This paper has indicated a need for experimental studies to base simulated rain event properties on those delineated for the region being investigated, or, if no long-term records are available, on the event properties described from comparable rainfall environments elsewhere.

A context within which the study of rain events can be placed is the exploration of the magnitude–frequency spectrum as manifested in soil erosion, the production of surface runoff, etc. The bias towards extreme rain rates in rainfall simulation experiments could be evaluated more fully in the light of a better understanding of this spectrum. Marques *et al.* (2007) emphasized some of these points in a study of soil erosion under low rain rate rainfall simulation. As noted earlier, these authors emphasize that even if soil loss is less spectacular than the losses triggered by rarer, high rain rate events, this may be offset by the relatively frequency of lower rain rate events. They recommended that increased attention be paid to low rain rate events, especially where the climatic context suggests that these might be important hydrologically and geomorphically.

If all studies of hydrologic and geomorphic processes driven by rain and rain event properties can include some common methodologies and objectives, then the advancement of these disciplines will be assisted. These objectives should include deeper exploration of rain event properties and the explanatory power that can arise from their more systematic inclusion in experimental designs. This would support at least three further valuable objectives: (1) To resolve more fully the spectrum of rain event properties, and to uncover any distinct landscape, hydrologic or geomorphic process links to low rain rates, moderate rain rates, high rain rates, etc. (2) To document more systematically how such event properties vary regionally and across environmental gradients, and to acquire baseline data against which future change can be judged. (3) To draw attention to changing rain event properties as global and regional environments change, as a means of understanding potential consequences for important landscape processes including interception, infiltration, runoff production, and the erosive work of rain and surface flows.

ACKNOWLEDGEMENT

I thank the two anonymous reviewers whose constructive criticisms resulted in an improved paper.

REFERENCES

Abu-Zreig M. 2006. Control of rainfall-induced soil erosion with various types of polyacrylamide. *Journal of Soils and Sediments* **6**: 137–144.

Aerts R, Maes W, November E, Behailu M, Poesen J, Deckers J, Hermy M, Muys B. 2006. Surface runoff and seed trapping efficiency of shrubs in a regenerating semiarid woodland in northern Ethiopia. *Catena* **65**: 61–70.

Agassi M, Bradford JM. 1999. Methodologies for interrill soil erosion studies. *Soil and Tillage Research* **49**: 277–287.

Alcoverro J, Corominas J, Gómez M. 1999. The Barranco de Arás flood of 7 August 1996 (Biescas, Central Pyrenees, Spain). *Engineering Geology* **51**: 237–255.

Amengual A, Romero R, Gómez M, Martín A, Alonso S. 2007. A hydrometeorological modelling study of a flash-flood event over Catalonia, Spain. *Journal of Hydrometeorology* **8**: 282–303.

Angel JR, Palecki MA, Hollinger SE. 2005. Storm precipitation in the United States. Part II: soil erosion characteristics. *Journal of Applied Meteorology* **44**: 947–959.

Angermann T, Wallender WW, Wilson BW, Werner I, Hinton DE, Oliver MN, Zalom FG, Henderson JD, Oliviera GH, Deanovic LA, Osterli P, Krueger W. 2002. Runoff from orchard floors—micro-plot field experiments and modelling. *Journal of Hydrology* **265**: 178–194.

Armstrong ASB, Tanton TW, Rycroft DW. 1998. The effect of ped size, simulated rainfall duration and frequency on the leaching of salts from clay topsoils. *Agricultural and Water Management* **37**: 133–143.

Arnaez J, Lasanta T, Ruiz-Flaño P, Ortigosa L. 2007. Factors affecting runoff and erosion under simulated rainfall in Mediterranean vineyards. *Soil and Tillage Research* **93**: 324–334.

Asadi H, Ghadiri H, Rose CW, Rouhipour H. 2007. Interill soil erosion processes and their interaction on low slopes. *Earth Surface Processes and Landforms* **32**: 711–724.

Asdak C, Jarvis PG, van Gardingen P, Fraser A. 1998. Rainfall interception loss in unlogged and logged forest areas of central Kalimantan, Indonesia. *Journal of Hydrology* **206**: 237–244.

Balme M, Vischel T, Lebel T, Peugeot C, Galle S. 2006. Assessing the water balance in the Sahel: impact of small scale rainfall variability on runoff Part 1: rainfall variability analysis. *Journal of Hydrology* **331**: 336–348.

Barcelo A, Robert R, Coudray J. 1997. A major rainfall event: the 27 February—5 March 1993 rains on the southeastern slope of Piton de la Fournaise massif (Reunion Island, southwest Indian Ocean). *Monthly Weather Review* **125**: 3341–3346.

Barger NN, Herrick JE, Van Zee J, Belnap J. 2006. Impacts of biological soil crust disturbance and composition on C and N loss from water erosion. *Biogeochemistry* **77**: 247–263.

Battany MC, Grismer ME. 2000. Rainfall runoff and erosion in Napa Valley vineyards: effects of slope, cover and surface roughness. *Hydrological Processes* **14**: 1289–1304.

Belmonte AMC, Beltrán FS. 2001. Flood events in Mediterranean ephemeral streams (ramblas) in Valencia region, Spain. *Catena* **45**: 229–249.

Bergkamp G, Cerda A, Imeson AC. 1999. Magnitude-frequency analysis of water redistribution along a climate gradient in Spain. *Catena* **37**: 129–146.

Bertol I, Mello EL, Guadagnin JC, Zaparolli ALV, Carrafa MR. 2003. Nutrient losses by water erosion. *Scientia Agricola* **60**: 581–586.

Bidin K, Chappell NA. 2006. Characteristics of rain events at an inland locality in north eastern Borneo, Malaysia. *Hydrological Processes* **20**: 3835–3850. DOI: 10·1002/hyp6160.

Biemelt D, Schapp A, Kleeberg A, Grünewald U. 2005. Overland flow, erosion, and related phosphorus and iron fluxes at plot scale: a case study from a non-vegetated lignite mining dump in Lusatia. *Geoderma* **129**: 4–18.

Blackburn WH, Meeuwig RO, Skau CM. 1974. A mobile infiltrometer for use on rangeland. *Journal of Range Management* **27**: 322–323.

Bochet E, Poesen J, Rubio JL. 2006. Runoff and soil loss under individual plants of a semi-arid Mediterranean shrubland: influence of plant morphology and rainfall intensity. *Earth Surface Processes and Landforms* **31**: 536–549.

Bracken LJ, Cox NJ, Shannon J. 2008. The relationship between rainfall inputs and flood generation in south-east Spain. *Hydrological Processes* **22**: 683–696. DOI: 10·1002/hyp.6641.

Brady JA, Wallender WW, Werner I, Mostafazadeh Fard B, Zalom FG, Oliver MN, Wilson BW, Mata MM, Henderson JD, Deanovic LA, Upadhaya S. 2006. Pesticide runoff from orchard floors in Davis, California, USA: a comparative analysis of diazinon and esfenvalerate. *Agriculture, Ecosystems and Environment* **115**: 56–68.

Browning KA. 1990. Rain, rainclouds and climate. *Quarterly Journal of the Royal Meteorological Society* **116**: 1025–1051.

Bryant ML, Bhat S, Jacobs JM. 2005. Measurements and modeling of throughfall variability for five forest communities in the south eastern US. *Journal of Hydrology* **312**: 95–108.

Casermeiro MA, Molina JA, de la Cruz Caravaca MT, Hernando Costa J, Hernando Massanet MI, Moreno PS. 2004. Influence of scrubs on runoff and sediment loss in soils of Mediterranean climate. *Catena* **57**: 91–107.

Cattan P, Cabidoche YM, Lcas JG, Colz M. 2006. Effects of tillage and mulching on runoff under banana (*Musa* spp.) on a tropical Andosol. *Soil and Tillage Research* **86**: 38–51.

Carlyle-Moses DE. 2004. A reply to R. Keim's comment on "Measurement and modelling of growing-season canopy water fluxes in a mature mixed deciduous forest stand, Southern Ontario, Canada". *Agricultural and Forest Meteorology* **124**: 281–284.

Cerda A, García-Fayos P. 2002. The influence of seed size and shape on their removal by water erosion. *Catena* **48**: 293–301.

Cerdan O, Le Bissonnais Y, Couturier A, Bourennane H, Souchere V. 2002. Rill erosion on cultivated hillslopes during two extreme rainfall events in Normandy, France. *Soil and Tillage Research* **67**: 99–108.

Chahinian N, Moussa R, Andrieux P, Voltz M. 2005. Comparison of infiltration models to simulate flood events at the field scale. *Journal of Hydrology* **306**: 191–214.

Chaplot VAM, Le Bissonnais Y. 2003. Runoff features for interrill erosion at different rainfall intensities, slope lengths, and gradients in an agricultural loessial hillslope. *Soil Science Society of America Journal* **67**: 844–851.

Chauvan F, Denvil S. 2007. Changes in severe indices as simulated by two French coupled global climate models. *Global and Planetary Change* **57**: 96–117.

Chen H, Dadson S, Chi TG. 2006. Recent rainfall-induced landslides and debris flow in northern Taiwan. *Geomorphology* **77**: 112–125.

Cheng JD, Huang YC, Wu HL, Yeh JL, Chang CH. 2005. Hydrometeorological and landuse attributes of debris flows and debris floods during typhoon Toraji, July 29–30, 2001 in central Taiwan. *Journal of Hydrology* **306**: 161–173.

Chiu LS, Chang ATC. 2001. Diurnal cycle of oceanic precipitation from microwave radiometry. In *Proceedings of International Geoscience and Remote Sensing Conference (IGARSS) 2001*, Sydney, Australia. *Scanning the present and resolving the future,* 9–13 July 2001, Institute of Electrical and Electronic Engineers.

Clarke MA, Walsh RDP. 2007. A portable rainfall simulator for field assessment of splash and slopewash in remote locations. *Earth Surface Processes and Landforms* **32**: 2052–2069. DOI: 10·1002/esp.1526.

Cosgrove CM, Garstang M. 1995. Simulation of rain events from rain-gauge measurements. *International Journal of Climatology* **15**: 1021–1029.

Costin AB, Gilmour DA. 1970. Portable rainfall simulator and plot unit for use in field studies of infiltration, runoff and erosion. *Journal of Applied Ecology* **7**: 193–200.

Cousin I, Issa OM, Le Bissonnais Y. 2005. Microgeometrical characterisation and percolation threshold evolution of a soil crust under rainfall. *Catena* **62**: 173–188.

Cras A, Marc V, Travi Y. 2007. Hydrological behaviour of sub-Mediterranean alpine headwater streams in a badlands environment. *Journal of Hydrology* **339**: 130–144.

Cuartas LA, Tomasella J, Nobre AD, Hodnett MG, Waterloo MJ, Múnera JC. 2007. Interception water-partitioning dynamics for a pristine rainforest in Central Amazonia: marked differences between normal and dry years. *Agricultural and Forest Meteorology* **145**: 69–83.

Cutrim EMC, Martin DW, Butzow DG, Silva IM, Yulaeva E. 2000. Pilot analysis of hourly rainfall in central and eastern Amazonia. *Journal of Climate* **13**: 1326–1334.

Dai A. 2001. Global precipitation and thunderstorm frequencies. Part 1: seasonal and interannual variations. *Journal of Climate* **14**: 1092–1111.

Dai A, Lin X, Hsu K-L. 2007. The frequency, intensity, and diurnal cycle of precipitation in surface and satellite observations over low- and mid-latitudes. *Climate Dynamics* **29**: 727–744.

Dairaku K, Emori S, Oki T. 2004. Rainfall amount, intensity, duration and frequency relationships in the Mae Chaem watershed in southeast Asia. *Journal of Hydrometeorology* **5**: 458–470.

Deluca TH, Patterson WA, Freimund WA, Cole DN. 1998. Influence of llamas, horses, and hikers on soil erosion from established recreation trails in Western Montana, USA. *Environmental Management* **22**: 255–262.

De Luis M, Raventós J, Cortina J, González-Hidalgo JC, Sánchez JR. 2004. Fire and torrential rainfall: effects on the perennial grass *Brachypodium retusum*. *Plant Ecology* **173**: 225–232.

De Ploey J, Kirkby MJ, Ahnert F. 1991. Hillslope erosion by rainstorms—a magnitude-frequency analysis. *Earth Surface Processes and Landforms* **16**: 399–409.

Descheemaeker K, Nyssen J, Poesen J, Raes D, Haile M, Muys B, Deckers S. 2006. Runoff on slopes with restoring vegetation: a case study from the Tigray highlands, Ethiopia. *Journal of Hydrology* **331**: 219–241.

Dimoyiannis DG, Valmis S, Vyrlas P. 2001. A rainfall simulation study of erosion of some calcareous soils. *Global Nest* **3**: 179–183.

Ding Y, Reiter ER. 1982. A relationship between planetary waves and persistent rain- and thunderstorms in China. *Archives for Meteorology, Geophysics, and Bioclimatology* **B31**: 221–252.

Dougherty WJ, Fleming NK, Cox JW, Chittleborough DJ. 2004. Rainfall simulation underestimates runoff phosphorus concentrations from dairy pastures. *Proceedings of SuperSoil 2004 Conference (Third Australian New Zealand Soils Conference)*, 5–9 December 2004, Sydney, Australia.

Drufuc G, Zawadzki II. 1975. Statistics of raingauge data. *Journal of Applied Meteorology* **14**: 1419–1429.

Dubicki A, Malinowska-Malek J, Stronska K. 2005. Flood hazards in the upper and middle Odra River basin—a short review over the last century. *Limnologica* **35**: 1230131.

Dunkerley DL. 2008. Intra-storm evaporation as a component of canopy interception loss in dryland shrubs: observations from Fowlers Gap, Australia. *Hydrological Processes* **22**(12): 1985–1995 DOI: 10·1002/hyp.6783.

Dunkerley DL. Submitted. Identifying individual rain events from pluviograph records: a review with analysis of data from an Australian dryland site. Submitted. *Hydrological Processes*.

Dunkerley D, Martin N, Berg SS, Ferguson R. Submitted. Fire, Catchment Runoff and Erosion Processes, and Post-fire Rehabilitation Programs: Recent Australian Experience. Chapter to appear in two volume monograph on *Rehabilitation following forest fires*, edited by A Cerda and P Robichaud.

Dykes AP, Thornes JB. 2000. Hillslope hydrology in tropical rainforest steeplands in Brunei. *Hydrological Processes* **14**: 215–235.

Egodawatta P, Thomas E, Goonetilleke A. 2007. Mathematical interpretation of pollutant wash-off from urban road surfaces using simulated rainfall. *Water Research* **41**: 3025–3031.

Elliott AH, Tian YQ, Rutherford JC, Carlson WT. 2002. Effect of cattle treading on interrill erosion from hill pasture: modelling concepts and analysis of rainfall simulator data. *Australian Journal of Soil Research* **40**: 963–976.

Fink AH, Knippertz P. 2003. An extreme precipitation event in southern Morocco in spring 2002 and some hydrological implications. *Weather* **58**: 377–387.

Fleischbein K, Wilcke W, Goller R, Boy J, Valarezo C, Zech W, Knoblich K. 2005. Rainfall interception in a lower montane forest in Ecuador: effects of canopy properties. *Hydrological Processes* **19**: 1355–1371.

Foley JL, Silburn DM. 2002. Hydraulic properties of rain impact surface seals on three clay soils—influence of raindrop impact frequency and rainfall intensity during steady state. *Australian Journal of Soil Research* **40**: 1069–1083.

Fornis RL, Vermeulen HR, Niewenhuis JD. 2005. Kinetic energy-rainfall intensity relationship for central Cebu, Philippines for soil erosion studies. *Journal of Hydrology* **300**: 20–32.

Foster IDL, Fullen MA, Brandsma RT, Chapman AS. 2000. Drip-screen rainfall simulators for hydro- and pedo-geomorphological research: the Coventry experience. *Earth Surface Processes and Landforms* **25**: 691–707.

Franklin D, Truman C, Potter T, Bosch D, Strickland T, Bednarz C. 2007. Nitrogen and phosphorus runoff losses from variable and constant intensity rainfall simulations on loamy sand under conventional and strip tillage systems. *Journal of Environmental Quality* **36**: 846–854.

Fraser AI, Harrod TR, Haygarth PM. 1999. The effect of rainfall intensity on soil erosion and particulate phosphorus transfer from arable soils. *Water Science and Technology* **12**: 41–45.

Frauenfeld B, Truman C. 2004. Variable rainfall intensity effects on runoff and interrill erosion from two coastal plain ultisols in Georgia. *Soil Science* **169**: 143–154.

Gaume E, Livet M, Desbordes M, Villeneuve JP. 2004. Hydrological analysis of the river Aude, France, flash flood on 12 and 13 November 1999. *Journal of Hydrology* **286**: 135–154.

García-Bartual R, Schneider M. 2001. Estimating maximum expected short-duration rainfall intensities from extreme convective storms. *Physics and Chemistry of the Earth* **26**: 675–681.

Germer S, Elsenbeer H, Moraes JM. 2005. Throughfall and temporal trends of rainfall distribution in an open tropical rainforest, south-western Amazonia (Rondonia, Brazil). *Hydrology and Earth System Sciences Discussions* **2**: 2707–2738.

Gerold LA, Watkins DW. 2005. Short duration rainfall frequency analysis in Michigan using scale-invariance assumptions. *Journal of Hydrologic Engineering* **10**: 450–457.

Ghadiri H, Payne D. 1988. The formation and characteristics of splash following raindrop impact on soil. *Journal of Soil Science* **39**: 563–575.

Gochis DJ, Watts CJ, Garatuza-Payan J, Cesar-Rodriguez J. 2007. Spatial and temporal patterns of precipitation intensity as observed by the NAME event rain gauge network from 2002 to 2004. *Journal of Climate* **20**: 1734–1750.

Gómez JA, Nearing MA. 2005. Runoff and sediment losses from rough and smooth soil surfaces in a laboratory experiment. *Catena* **59**: 253–266.

Goudie AS. 2006. Global warming and fluvial geomorphology. *Geomorphology* **79**: 384–394.

Green VS, Stott DE, Norton LD, Graveel JG. 2000. Polyacrylamide molecular weight and charge effects on infiltration under simulated rainfall. *Soil Science of America Journal* **64**: 1786–1791.

Guo Y. 2006. Updating rainfall IDF relationships to maintain urban drainage design standards. *Journal of Hydrologic Engineering* **11**: 506–509.

Gutiérrez F, Gutiérrez M, Sancho C. 1998. Geomorphological and sedimentological analysis of a catastrophic flash flood in the Arás basin (central Pyrenees, Spain). *Geomorphology* **22**: 265–283.

Guy BT, Dickinson WT, Rudra RP. 1990. Hydraulics of sediment-laden sheetflow and the influence of simulated rainfall. *Earth Surface Processes and Landforms* **15**: 101–118.

Haga H, Matsumoto Y, Matsutani J, Fujita M, Nishida K, Sakamoto Y. 2005. Flow paths, rainfall properties, and antecedent soil moisture controlling lags to peak discharge in a granitic unchanneled catchment. *Water Resources Research* **41**: DOI: 10·1029/2005WR004236.

Hamed Y, Albergel J, Pépin Y, Asseline J, Nasri S, Zante P, Berndtsson R, El-Niazy M, Balah M. 2002. Comparison between rainfall simulator erosion and observed reservoir sedimentation in an erosion-sensitive semiarid catchment. *Catena* **50**: 1–16.

Hammad AHA, Børresen T, Haugen LE. 2006. Effects of rain characteristics and terracing on runoff and erosion under the Mediterranean. *Soil and Tillage Research* **87**: 39–47.

Hand WH, Fox NI, Collier CG. 2004. A study of twentieth-century extreme rainfall events in the United Kingdom with implications for forecasting. *Meteorological Applications* **11**: 15–31.

Hanke M, Perry D, Kung K-JS, Bubenzer G. 2004. A low-intensity, high-uniformity water application system. *Soil Science Society of America Journal* **68**: 1833–1837.

Harden CP, Mathews L. 2000. Rainfall response of degraded soil following reforestation in the Copper Basin, Tennessee, USA. *Environmental Management* **26**: 163–174.

Harden CP, Scruggs PD. 2003. Infiltration on mountain slopes: a comparison of three environments. *Geomorphology* **55**: 5–24.

Hastings BK, Breshears DD, Smith FM. 2005. Spatial variability in rainfall erosivity versus rainfall depth: implications for sediment yield. *Vadose Zone Journal* **4**: 500–504.

Hawke RM, Price AG, Bryan RB. 2006. The effect of initial soil water content and rainfall intensity on near-surface soil hydrologic conductivity: a laboratory investigation. *Catena* **65**: 237–246.

Heathwaite AL, Johnes PJ. 1996. Contribution of nitrogen species and phosphorus fractions to stream water quality in agricultural catchments. *Hydrological Processes* **10**: 971–983.

Heneker TM, Lambert MF, Kuczera G. 2001. A point rainfall model for risk-based design. *Journal of Hydrology* **247**: 54–71.

Herngren L, Goonetilleke A, Sukpum R, de Silva DY. 2005. Rainfall simulation as a tool for urban water quality research. *Environmental Engineering Science* **22**: 378–383.

Hignett CT, Gusli S, Cass A, Besz W. 1995. An automated laboratory rainfall simulation system with controlled rainfall intensity, raindrop energy and soilo drainage. *Soil Technology* **8**: 31–42.

Hjelmfelt AT. 1978. Amazon basin hydrometeorology. Proceedings of the American Society of Civil Engineers, *Journal of the Hydraulics Division* **104**: 887–897 (cited by Lloyd, 1990).

Holden J, Burt TP. 2002. Infiltration, runoff and sediment production in blanket peat catchments: implications of field rainfall simulation experiments. *Hydrological Processes* **16**: 2537–2557.

Hörmann G, Branding A, Clemen T, Herbst M, Hinrichs A, Thamm F. 1996. Calculation and simulation of wind controlled canopy interception of a beech forest in Northern Germany. *Agricultural and Forest Meteorology* **79**: 131–148.

Hosking JG, Stow CD. 1987. The arrival rate of raindrops at the ground. *Journal of Climate and Applied Meteorology* **26**: 433–442.

Houze RA. 1997. Stratiform precipitation in regions of convection: a meteorological paradox? *Bulletin of the American Meteorological Society* **78**: 2179–2196.

Huang C, Gascuel-Odoux C, Cros-Cayot S. 2001. Hillslope topographic and hydrologic effects on overland flow and erosion. *Catena* **46**: 177–188.

Huff FA. 1967. Time distribution of rainfall in heavy storms. *Water Resources Research* **3**: 1007–1019.

Hume M. 1992. A 1951-80 global land precipitation climatology for the evaluation of general circulation models. *Climate Dynamics* **7**: 57–72.

Hunsche M, Damerow L, Schmitz-Eiberger M, Noga G. 2007. Mancozeb wash-off from apple seedlings by simulated rainfall as affected by drying time of fungicide deposit and rain characteristics. *Crop Protection* **26**: 768–774.

Hussein MH, Othman AK. 1988. Soil and water losses in a low intensity rainfall region in Iraq. *Hydrological Sciences Journal* **33**: 257–267.

Iida S, Tanaka T, Sugita M. 2005. Change of interception process due to the succession from Japanese red pine to evergreen oak. *Journal of Hydrology* **315**: 154–166.

Issa OM, Cousin I, Le Bissonnais Y, Quétin P. 2004. Dynamic evolution of the unsaturated hydraulic conductivity of a developing crust. *Earth Surface Processes and Landforms* **29**: 1131–1142.

Jackson IJ. 1988. Daily rainfall over Northern Australia: deviations from the world pattern. *Journal of Climatology* **8**: 463–476.

James PA, Alexander RW. 1998. Soil erosion and runoff in improved pastures of the Clwydian Range, North Wales. *Journal of Agricultural Science, Cambridge* **130**: 473–488.

Jameson AR, Kostinski AB. 2002. When is rain steady? *Journal of Applied Meteorology* **41**: 83–90.

Joel A, Messing I. 2001. Infiltration rate and hydraulic conductivity measured with rain simulator and disc permeameter on sloping arid land. *Arid Land Research and Management* **15**: 371–384.

Jones DMA, Sims AL. 1978. Climatology of instantaneous rainfall rates. *Journal of Applied Meteorology* **17**: 1135–1140.

Keim RF, Skaugset AE, Weiler M. 2006. Storage of water on vegetation under simulated rainfall of varying intensity. *Advances in Water Resources* **29**: 974–986.

Kinnell PIA. 2005. Sediment transport by medium to large drops impacting flows at subterminal velocity. *Soil Science Society of America Journal* **69**: 902–905.

Kitchen B, Goonetilleke A. 2007. Rainfall simulation to evaluate pollutant build-up and wash-off from selected land uses within a port environment. American Society of Civil Engineers, *Ports 2007: 30 years of sharing ideas, 1977–2007.* Compiled byWade Watson, American Society of Civil Engineers, ISBN: 0784408343.

Konishi S, Tani M, Kosugi Y, Takanashi S, Sahat MM, Nik AR, Niiyama K, Okuda T. 2006. Characteristics of spatial distribution of throughfall in a lowland tropical rainforest, Peninsular Malaysia. *Forest Ecology and Management* **224**: 19–25.

Koutsoyiannis D, Kozonis D, Manetas A. 1998. A mathematical framework for studying rainfall intensity-duration-frequency relationships. *Journal of Hydrology* **206**: 118–135.

Kubota H, Shirooka R, Ushiyama T, Chuda T, Reddy KK, Katsumata Y, Oneyama K, Yasunaga K, Moteki Q, Sato N, Fujita M, Suginohara N. 2006. Diurnal variation of precipitation observed over Palau in the western Pacific. *Abstracts, 27th Conference on hurricanes and tropical meteorology*, Monterey, California: American Meteorological Society.

Kuhn NJ. 2007. Erodibility of soil and organic matters: independence of organic matter resistance to interrill erosion. *Earth Surface Processes and Landforms* **32**: 794–802.

Kumar K, Pant D, Panda YS, Satyal GS. 2002. Runoff and soil loss from steep slopes treated with low cost bioengineering measures. *The Environmentalist* **22**: 133–141.

Kume T, Kuraji K, Yoshifuji N, Morooka T, Sawano S, Chong L, Suzuki M. 2006. Estimation of canopy drying time after rainfall using sap flow measurements in an emergent tree in a lowland mixed-dipterocarp forest in Sarawak, Malaysia. *Hydrological Processes* **20**: 565–578.

Laakso L, Grönholm T, Rannik U, Kosmale M, Fiedler V, Vehkamäki H, Kulmala M. 2003. Ultrafine particle scavenging coefficients calculated from 6 years field measurements. *Atmospheric Environment* **37**: 3605–3613.

Le Bissonnais Y, Blavet D, De Noni G, Laurent JY, Asseline J, Chenu C. 2007. Erodibility of Mediterranean vineyard soils: relevant aggregate stability methods and significant soil variables. *European Journal of Soil Science* **58**: 188–195.

Legates DR, Willmott CJ. 1990. Mean seasonal and spatial variability in gauge-corrected, global precipitation. *International Journal of Climatology* **10**: 111–127.

Léonard J, Ancelin O, Ludwig B, Richard G. 2006. Analysis of the dynamics of soil infiltrability of agricultural soils from continuous rainfall-runoff measurements on small plots. *Journal of Hydrology* **326**: 122–134.

Li PW, Lai EST. 2004. Short-range quantitative precipitation forecasting in Hong Kong. *Journal of Hydrology* **288**: 189–209.

Li XY, Xie ZK, Yan XK. 2004. Runoff characteristics of artificial catchment materials for rainwater harvesting in the semiarid regions of China. *Agricultural Water Management* **65**: 211–224.

Lin ML, Jeng FS. 2000. Characteristics of hazards induced by extremely heavy rainfall in central Taiwan—Typhoon Herb. *Engineering Geology* **58**: 191–207.

Liu QQ, Chen L, Li JC, Singh VP. 2004. Two-dimensional kinematic wave model of overland flow. *Journal of Hydrology* **291**: 28–41.

Lloyd CR. 1990. The temporal distribution of Amazonian rainfall and its implications for forest interception. *Quarterly Journal of the Royal Meteorological Society* **116**: 1487–1494.

Loch RJ, Foley JL. 1992. Effects of plot size on size distributions of water-stable material at the soil surface under simulated rain. *Australian Journal of Soil Research* **30**: 113–118.

Lusby GC, Toy TJ. 1976. An evaluation of surface-mine spoils area restoration in Wyoming using rainfall simulation. *Earth Surface Processes* **1**: 375–386.

Mamedov AI, Shainberg I, Levy GJ. 2000. Irrigation with effluent water: effects of rainfall energy on soil infiltration. *Soil Science Society of America Journal* **64**: 732–737.

Mamedov AI, Shainberg I, Levy GJ. 2002. Wetting rate and sodicity effects on interrill erosion from semi-arid Israeli soils. *Soil and Tillage Research* **68**: 121–132.

Mandal UK, Rao KV, Mishra PK, Vittal KPR, Sharma KL, Narsimlu B, Venkanna K. 2004. Soil infiltration, runoff and sediment yield from a shallow soil with varied stone cover and intensity of rain. *European Journal of Soil Science* **56**: 435–443.

Marcos E, Tárrega R, Luis-Calabuig E. 2000. Comparative analysis of runoff and sediment yield with a rainfall simulator after experimental fire. *Arid Soil Research and Rehabilitation* **14**: 293–307.

Marques MJ, Bienes R, Jiménez L, Pérez-Rodríguez R. 2007. Effect of vegetal cover on runoff and soil erosion under light intensity events. Rainfall simulation over USLE plots. *Science of the Total Environment* **378**: 161–165.

Marques MJ, Bienes R, Pérez-Rodríguez R, Jiménez L. 2008. Soil degradation in central Spain due to sheet water erosion by low-intensity rainfall events. *Earth Surface Processes and Landforms* **33**: 414–423. DOI: 10·1002/esp.1564.

Martin DA, Moody JA. 2001. Comparison of soil infiltration rates in burned and unburned mountainous watersheds. *Hydrological Processes* **15**: 2893–2903.

Martínez-Casanovas JA, Ramos MC, Ribes-Dasi M. 2002. Soil erosion caused by extreme rainfall events: mapping and quantification in agricultural plots from very detailed digital elevation models. *Geoderma* **105**: 125–140.

Martz LW. 1992. The variation of soil erodibility with slope position in a cultivated Canadian Prairie landscape. *Earth Surface Processes and Landforms* **17**: 543–556.

Mathys N, Klotz S, Esteves M, Descroix L, Lapetite JM. 2005. Runoff and erosion in the Black Marls of the French Alps: observations and measurements at the plot scale. *Catena* **63**: 261–281.

McEwen LJ, Werritty A. 1988. The hydrology and long-term geomorphic significance of a flash flood in the Cairngorm Mountains Scotland. *Catena* **15**: 361–377.

Mermut AR, Luk SH, Römkens MJM, Poesen JWA. 1997. Soil loss by splash and wash during rainfall from two loess soils. *Geoderma* **75**: 203–214.

Mills AJ, Fey MV. 2004. Frequent fires intensify soil crusting: physicochemical feedback in the pedoderm of long-term burn experiments in South Africa. *Geoderma* **121**: 45–64.

Milly PCD, Wetherald RT, Dunne KA, Delworth TL. 2002. Increasing risk of great floods in a changing climate. *Nature* **415**: 514–517.

Mohr KI, Famiglietti JS, Zipser EJ. 1999. The contribution to tropical rainfall with respect to convective system type, size, and intensity estimated from 85-GHz ice-scattering signature. *Journal of Applied Meteorology* **38**: 596–606.

Moody JA, Martin DA. 2001. Post-fire, rainfall intensity-peak discharge relations for three mountainous watersheds in the western USA. *Hydrological Processes* **15**: 2981–2993.

Mueller EA. 1962. Raindrop distributions at Miami, Florida. Illinois State Water Survey, Champaign, Illinois (cited by Foley and Silburn, 2002).

Müller K, Trolove M, James TK, Rahman A. 2004. Herbicide loss in runoff: effects of herbicide properties, slopes, and rainfall intensity. *Australian Journal of Soil Research* **42**: 17–27.

Moszkowicz S. 2000. Small-scale structure of rain field—preliminary results basing on a digital gauge network and on MRL-5 Legionowo radar. *Physics and Chemistry of the Earth* **25**: 933–938.

Neave M, Rayburg S. 2007. Nonlinear biofluvial responses to vegetation change in a semiarid environment. *Geomorphology* **89**: 217–239.

Ndiaye B, Esteves M, Vandervaere JP, Lapetite JM, Vauclin M. 2005. Effect of rainfall and tillage direction on the evolution of surface crusts, soil hydraulic properties and runoff generation for a sandy loam soil. *Journal of Hydrology* **307**: 294–311.

Nearing MA, Pruski FF, O'Neal MR. 2004. Expected climate change impacts on soil erosion rates: a review. *Journal of Soil and Water Conservation* **59**: 43–50.

Nicholls N, Kariko A. 1993. East Australian rainfall events: interannual variations, trends, and relationships with the Southern Oscillation. *Journal of Climate* **6**: 1141–1152.

Nicolau JM. 2002. Runoff generation and routing on artificial slopes in a Mediterranean-continental environment: the Teruel coalfield, Spain. *Hydrological Processes* **16**: 631–647.

Nolan SC, van Vliet LJP, Goddard TW, Flesch TK. 1997. Estimating storm erosion with a rainfall simulator. *Canadian Journal of Soil Science* **77**: 669–676.

Nystuen JA, Amitai E. 2003. High temporal resolution of extreme rainfall rate variability and the acoustic classification of rainfall. *Journal of Geophysical Research* **108**: DOI 10·1029/2001JD0011481.

Oostwoud Wijdenes DJ, Ergenzinger P. 1998. Erosion and sediment transport on steep marly hillslopes, Draix, Haute-Provence, France: an experimental field study. *Catena* **33**: 179–200.

Palecki MA, Angel JR, Hollinger SE. 2005. Storm precipitation in the United States. Part 1: Meteorological characteristics. *Journal of Applied Meteorology* **44**: 933–946.

Palmer TN, Räisänen J. 2002. Quantifying the risk of extreme seasonal precipitation events in a changing climate. *Nature* **415**: 512–514.

Pan C, Shangguan Z. 2006. Runoff hydraulic characteristics and sediment generation in sloped grassplots under simulated rainfall conditions. *Journal of Hydrology* **331**: 178–185.

Parsons AJ, Stone PM. 2006. Effects of intra-storm variations in rainfall intensity on interrill runoff and erosion. *Catena* **67**: 68–78.

Pawlina AA. 2002. No rain intervals within rain events: some statistics based on Milano radar and rain-gauge data. *Proceedings of COST 280, First International Workshop on propagation impairment mitigation for millimetre wave radio systems*, 1–3 July 2002, Malvern, UK.

Peters O, Christensen K. 2002. Rain: relaxations in the sky. *Physical Review*. DOI: 10·1103/PhysRevE.66·036120.

Polyakov VO, Lal R. 2004. Soil erosion and carbon dynamics under simulated rainfall. *Soil Science* **169**: 590–599.

Potter TL, Truman CC, Strickland TC, Bosch DD, Webster TM, Franklin DH, Bednarz CW. 2006. Combined effects of constant versus variable intensity simulated rainfall and reduced tillage management on cotton preemergence herbicide runoff. *Journal of Environmental Quality* **35**: 1894–1902.

Poulendard J, Podwojewski P, Janeau JL, Collinet J. 2001. Runoff and soil erosion under rainfall simulation of Andisols from Ecuadorian *Páramo*: effect of tillage and burning. *Catena* **45**: 185–207.

Putuehna WM, Cordery I. 1996. Estimation of interception capacity of the forest floor. *Journal of Hydrology* **180**: 283–299.

Pypker TG, Bond BJ, Link TE, Marks D, Unsworth MH. 2005. The importance of canopy structure in controlling the interception loss of rainfall: examples from a young and an old-growth Douglas-fir forest. *Agricultural and Forest Meteorology* **130**: 113–129.

Ramos MC, Matínez-Casanovas JA. 2004. Nutrient losses from a vineyard in northeastern Spain caused by an extraordinary rainfall event. *Catena* **55**: 79–90.

Ramos MC, Matínez-Casanovas JA. 2006. Erosion rates and nutrient losses affected by composted cattle manure application in vineyard soils of NE Spain. *Catena* **68**: 177–185.

Rauch W, De Toffol S. 2006. On the issue of trend and noise in the estimation of extreme rainfall properties. *Water Science and Technology* **54**: 17–24.

Rebora N, Ferraris L. 2006. The structure of convective rain cells at mid-latitudes. *Advances in Geosciences* **7**: 3–35.

Riezebos HT, Seyhan E. 1977. Essential conditions of rainfall simulation for laboratory water erosion experiments. *Earth Surface Processes* **2**: 185–190.

Römkens MJM, Prasad SN. 2006. Rain infiltration into swelling/shrinking/cracking soils. *Agricultural Water Management* **86**: 196–205.

Roth CH, Helming K. 1992. Dynamics of surface sealing, runoff formation and interrill erosion loss as related to rainfall intensity, microrelief and slope. *Zeitschrift für Pflanzenernährung und Bodenkunde* **155**: 209–216.

Rulli MC, Bozzi S, Spada M, Bocchiola D, Rosso R. 2006. Rainfall simulations on a fire disturbed Mediterranean area. *Journal of Hydrology* **327**: 323–338.

Sadras VO. 2003. Influence of size of rainfall events on water-driven processes I. Water budget of wheat crops in south-eastern Australia. *Australian Journal of Agricultural Research* **54**: 341–351.

Scarbrough DA, Coblentz WK, Humphry JB, Coffey KP, Sauer TJ, Jennings JA, Daniel TC, Turner JE, Kellog DW. 2004. Estimating losses of dry matter from simulated rainfall on Bermudagrass and Orchardgrass forages using cell well components as markers. *Agronomy Journal* **96**: 1680–1787.

Scatena FN. 1990. Watershed scale rainfall interception on two forested watersheds in the Liquillo Mountains of Puerto Rico. *Journal of Hydrology* **113**: 89–102.

Sato Y, Kumagai T, Kume A, Otsuki K, Ogawa S. 2004. Experimental analysis of moisture dynamics of litter layers—the effects of rainfall conditions and leaf shapes. *Hydrological Processes* **18**: 3007–3018.

Schellekens J, Bruijnzeel LA, Scatena FN, Bink NJ, Holwerda F. 2000. Evaporation from a tropical rain forest, Luquillo Experimental Forest, eastern Puerto Rico. *Agricultural and Forest Meteorology* **36**: 2183–2196.

Scherrer S, Naef F, Faeh AO, Cordery I. 2007. Formation of runoff at the hillslope scale during intense precipitation. *Hydrology and Earth System Sciences* **11**: 907–922.

Schiettecatte W, Jin K, Yao Y, Cornelis WM, Lu J, Wu H, Verbist K, Cai D, Gabriels D, Hartmann R. 2005a. Influence of simulated rainfall on physical properties of a conventionally tilled loess soil. *Catena* **64**: 209–221.

Schiettecatte W, Ouessar M, Gabriels D, Tanghe S, Heirman S, Abdelli F. 2005b. Impact of water harvesting techniques on soil and water conservation: a case study on a micro catchment in southeastern Tunisia. *Journal of Arid Environments* **61**: 297–313.

Schlesinger WH, Abrahams AD, Parsons AJ, Wainwright J. 1999. Nutrient losses in runoff from grassland and shrubland habitats in southern New Mexico: I. rainfall simulation experiments. *Biogeochemistry* **45**: 21–34.

Schumacher C, Houze RA. 2006. Straiform precipitation production over sub-Saharan Africa and the tropical East Atlantic as observed by TRMM. *Quarterly Journal of the Royal Meteorological Society* **132**: 2235–2255.

Seager R, Ting M, Held I, Kushnir Y, Lu J, Vecchi G, Huang HP, Harnik N, Leetmaa A, Lau NC, Li C, Velez J, Naik N. 2007. Model projections of an imminent transition to a more arid climate in southwestern North America. *Science* **316**: 1181–1184.

Seeger M. 2007. Uncertainty of factors determining runoff and erosion processes as quantified by rainfall simulations. *Catena* **71**: 56–67.

Shainberg I, Mamedov AI, Levy GJ. 2003. Role of wetting rate and rain energy in seal formation and erosion. *Soil Science* **168**: 54–62.

Sheridan GJ, Lane PNJ, Noske PJ. 2007. Quantification of hillslope runoff and erosion processes before and after wildfire in a wet *Eucalyptus* forest. *Journal of Hydrology* **343**: 12–28.

Shin DB, Chiu LS, Kafatos M. 2001. Comparison of the monthly precipitation derived from the TRMM satellite. *Geophysical Research Letters* **28**: 795–798.

Smith DR, Owens PR, Leytem AB, Warnemuende EA. 2007. Nutrient losses from manure and fertilizer applications as impacted by time to first runoff event. *Environmental Pollution* **147**: 131–137.

Smith JA, Krajewski WF. 1993. A modelling study of rainfall rate-reflectivity relationships. *Water Resources Research* **29**: 2505–2514.

Smith JA, De Veaux RD. 1994. A stochastic model relating rainfall intensity to raindrop processes. *Water Resources Research* **30**: 651–664.

Smith JA, Baeck ML, Zhang Y. 2001. Extreme rainfall and flooding from supercell thunderstorms. *Journal of Hydrometeorology* **2**: 469–489.

Smith JA, Miller AJ, Baeck ML, Nelson PA, Fisher GT, Meierdiercks KL. 2005. Extraordinary flood response of a small urban watershed to short-duration convective rainfall. *Journal of Hydrometeorology* **6**: 599–617.

Struthers I, Sivapalan M, Hinz C. 2007. Conceptual examination of climate-soil controls upon rainfall partitioning in an open-fractured soil: 1. Single storm response. *Advances in Water Resources* **30**: 505–517.

Tejada M, Gonzalez JL. 2007. Influence of organic amendments on soil structure and soil loss under simulated rain. *Soil and Tillage Research* **93**: 197–205.

Telesca L, Lapenna V, Scalcione E, Summa D. 2007. Searching for time-scaling features in rainfall sequences. *Chaos, Solitons and Fractals* **32**: 35–41.

Terry JP. 1998. A rainsplash component analysis to define mechanisms of soil detachment and transportation. *Australian Journal of Soil Research* **36**: 525–542.

Thurow TL, Blackburn WH, Taylor CA. 1988. Infiltration and interrill erosion responses to selected livestock grazing strategies, Edwards Plateau, Texas. *Journal of Range Management* **41**: 296–302.

Toba T, Ohta T. 2005. An observational study of the factors that influence interception loss in boreal and temperate forests. *Journal of Hydrology* **313**: 208–220.

Tobón Marin C, Bouten W, Sevink J. 2000. Gross rainfall and its partitioning into throughfall, stemflow and evaporation of intercepted water in four forest ecosystems in western Amazonia. *Journal of Hydrology* **237**: 40–57.

Tokay A, Short DA. 1996. Evidence from tropical raindrop spectra of the origin of rain from stratiform versus convective clouds. *Journal of Applied Meteorology* **35**: 355–371.

Tokay A, Wolff DB, Wolff KR, Bashor P. 2003. Rain gauge and disdrometer measurements during the Keys Area Microphysics Project (KAMP). *Journal of Atmospheric and Oceanic Technology* **20**: 1460–1477.

Trefry CM, Watkins DW, Johnson D. 2005. Regional rainfall frequency analysis for the State of Michigan. *Journal of Hydrologic Engineering* **10**: 437–449.

Trenberth KE. 1998. Atmospheric moisture residence times and cycling: implications for rainfall rates and climate change. *Climatic Change* **39**: 667–694.

Truman CC, Strickland TC, Potter TL, Franklin DH, Bosch DD, Bednarz CW. 2007. Variable rainfall intensity and tillage effects on runoff, sediment, and carbon losses from a loamy sand under simulated rainfall. *Journal of Environmental Quality* **36**: 1495–1502.

Uijlenhoet R, Sempere Torres D. 2006. Measurement and parameterization of rainfall microstructure. *Journal of Hydrology* **328**: 1–7.

Valmis S, Dimoyiannis D, Danalatos NG. 2005. Assessing interrill erosion rate from soil aggregate instability index, rainfall intensity and slope angle on cultivated soils in central Greece. *Soil and Tillage Research* **80**: 139–147.

Van Dijk AIJM, Bruijnzeel LA. 2001. Modelling rainfall interception by vegetation of variable density using an adapted analytical model. Part 2. Model validation for a tropical upland mixed cropping system. *Journal of Hydrology* **247**: 239–262.

Van Dijk AIJM, Bruijnzeel LA, Rosewell CJ. 2002. Rainfall intensity-kinetic energy relationships: a critical literature appraisal. *Journal of Hydrology* **261**: 1–23.

Victoria C, Kacevas A, Fiori H. 1997. Soil vulnerability in Uruguay: potential effects of an increase in erosive rainfall on soil loss. *Climate Research* **9**: 41–46.

Vigiak O, Romanowicz RJ, van Loon EE, Sterk G, Beven KJ. 2006. A disaggregating approach to describe overland flow occurrence within a catchment. *Journal of Hydrology* **323**: 22–40.

Vilar E, Burgueño A. 1995. Statistical properties of 49 years of rainfall rate events. *Theoretical and Applied Climatology* **50**: 213–225.

Volken M, Schumann T. 1993. A critical review of below-cloud aerosol scavenging results on Mt Rigi. *Water, Air, and Soil Pollution* **68**: 15–28.

Wallbrink PJ, Croke J. 2002. A combined rainfall simulator and tracer approach to assess the role of Best Management Practices in minimising sediment redistribution and loss in forests after harvesting. *Forest Ecology and Management* **170**: 217–232.

Wan Y, El-Swaify SA, Sutherland RA. 1996. Partitioning interrill splash and wash dynamics: a novel laboratory approach. *Soil Technology* **9**: 55–69.

Watung RL, Sutherland RA, El-Swaify SA. 1996. Influence of rainfall energy flux density and antecedent soil moisture content on splash transport and aggregate enrichment ratios for a Hawaiian Oxisol. *Soil Technology* **9**: 251–272.

Wei W, Chen L, Fu B, Huang Z, Wu D, Gui L. 2007. The effect of land uses and rainfall regimes on runoff and soil erosion in the semi-arid loess hilly area, China. *Journal of Hydrology* **335**: 247–258.

Wilson JP, Seney JP. 1994. Erosional impact of hikers, horses, motorcycles, and off-road bicycles on mountain trails in Montana. *Mountain Research and Development* **14**: 77–88.

Winston WE, Criss RE. 2002. Geochemical variations during flash flooding, Meramec River basin, May 2000. *Journal of Hydrology* **265**: 149–163.

Wischmeier WH, Smith DD. 1978. Predicting rainfall erosion losses—a guide to conservation planning. U.S. Department of Agriculture, Agriculture Handbook 537.

Woo MK, Huang L, Zhang S, Li Y. 1997. Rainfall in Guangdong province, South China. *Catena* **29**: 115–129.

Yang M, Yao T, Gou X, Wang H, Hao L. 2007. Comparison analysis of the summer monsoon precipitation between northern and southern slopes of Tanggula Mountains, Qinghai-Xizang (Tibetan) Plateau: a case study in summer 1998. *Hydrological Processes* **21**: 1841–1847.

Zehetner F, Miller WP. 2006. Erodibility and runoff-infiltration characteristics of volcanic ash soils along an altitudinal chronosequence in the Ecuadorian Andes. *Catena* **65**: 201–213.

Zeng N, Shuttleworth JW, Gash JHC. 2000. Influence of temporal variability of rainfall on interception loss. Part I. Point analysis. *Journal of Hydrology* **228**: 228–241.

Zhang J, Jiao JJ, Yang J. 2000. In situ infiltration studies at a hillside in Hubei Province, China. *Engineering Geology* **57**: 31–38.

Zhang G, Zeng GM, Jiang YM, Huang GH, Li JB, Yao JM, Tan W, Xaing R, Zhang XL. 2006. Modelling and measurement of two-layer canopy interception losses in a subtropical evergreen forest of central-south China. *Hydrology and Earth System Sciences* **10**: 65–77.

Zhou M, Li YC, Ndeki-Kizza P, O'Hair SK. 2003. Endosulfan losses through runoff and leaching from calcareous gravelly or marl soils. *Vadose Zone Journal* **2**: 231–238.

Zhou ZC, Shangguan ZP. 2007. The effects of ryegrass roots and shoots on loess erosion under simulated rainfall. *Catena* **70**: 350–355.

Ziegler AD, Sutherland RA. 1998. Reduction in interrill sediment transport by rolled erosion control systems. *Soil and Tillage Research* **45**: 265–278.

Ziegler AD, Negishi JN, Sidle RC, Noguchi S, Nik AR. 2006. Impacts of logging disturbance on hillslope saturated hydraulic conductivity in a tropical forest in Peninsular Malaysia. *Catena* **67**: 89–104.