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Impact of multi-day rainfall events on surface roughness and physical crusting of very fine soils

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ABSTRACT

Soil surface roughness (SSR), a description of the micro-relief of soils, affects the surface storage capacity of soils, influences the threshold flow for wind and water erosion and determines interactions and feedback processes between the terrestrial and atmospheric systems at a range of scales. Rainfall is an important determinant of SSR as it can cause the dislocation, reorientation and packing of soil particles and may result in the formation of physical soil crusts which can, in turn, affect the roughness and hydrological properties of soils. This paper describes an experiment to investigate the impact of a multi-day rainfall event on the SSR and physical crusting of very fine soils with low organic matter content, typical of a semi-arid environment. Changes in SSR are quantified using geostatistically-derived indicators calculated from semivariogram analysis of high resolution laser scans of the soil surface captured at a horizontal resolution of 78 µm (0.078 mm) and a vertical resolution of 12 µm (0.012 mm). Application of 2 mm, 5 mm and 2 mm of rainfall each separated by a 24 h drying period resulted in soils developing a structural two-layered 'sieving' crust characterised by a sandy micro-layer at the surface overlying a thin seal of finer particles. Analysis of the geostatistics and soil characteristics (e.g. texture, surface resistance, infiltration rate) suggests that at this scale of enquiry, and for low rainfall amounts, both the vertical and horizontal components of SSR are determined by raindrop impact rather than aggregate breakdown. This is likely due to the very fine nature of the soils and the low rainfall amounts applied.

1. Introduction

Soil surface roughness (SSR) describes the micro-relief of soils at the centimetre to decimetre scale (Römkens and Wang, 1986). This micro-relief affects the susceptibility of soils to erosion by both water (Kirkby, 2002) and wind (Zobeck and Popham, 1997) through its influence on infiltration (Vidal Vázquez et al., 2006), runoff (Dunkerley, 2004; Helming et al., 1998), overland flow (Darboux et al., 2001; Smith et al., 2011), drainage network evolution (Römkens et al., 2001), evaporation (Allmaras et al., 1977), threshold wind erosion (Chappell et al., 2006) and the surface storage capacity of both water and loose erodible material (Kamphorst et al., 2000; Onstad, 1984). Soil surface roughness also plays a role in modifying exchanges between the terrestrial and atmospheric systems (Rodríguez-Caballero et al., 2012) and affects interactions and feedback processes at a range of scales (Cammeraat, 2002; Smith, 2014).

SSR is controlled primarily by the soil's physical and chemical properties and changes over time in response to both natural and anthropogenically-enhanced physical (erosion/deposition) and biological processes. Römkens and Wang (1986) identified five scales of surface roughness the smallest of which is determined by primary soil particles (≤ mm). The next two scales are driven by the size and organisation of soil aggregates (mm) and by micro-topography (cm) caused by clods and surface cracking. The two largest scales identified are oriented roughness (dm) caused by agricultural activities and topographic roughness (≥ dm) caused by local topography and slope. For those scales where SSR is controlled by soil factors (up to cm scale), the surface roughness of soils typically reduces through time in response to rainfall as the original soil structural units are broken down from macroaggregates (> 250 µm) to microaggregates (20–250 µm) to primary soil particles by slaking, differential swelling, raindrop impacts and physico-chemical dispersion (Emerson and Greenland, 1990; Le Bissonais, 1996a, 1996b). The breakdown of aggregates can result in the formation of physical soil crusts. These crusts typically comprise a thin layer only a few millimetres thick that is more dense and with lower porosity than the underlying soil (Assouline, 2004). Known as

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'seals' when they are wet and crusts when dry, the presence of clays at the surface acts to bind particles together (Shainberg, 1992). Physical crust formation can therefore reduce rates of infiltration and splash erosion, but increase surface runoff and cumulative sediment yield (Agassi et al., 1985; Bradford and Huang, 1993; Kidron, 2007; Morin et al., 1989).

Physical soil crusts can be divided in to those formed by raindrop impact which causes in situ breakdown of aggregates into fine particles (structural crusts) and those formed by the translocation of fine particles into (depositional crusts) or away from (erosional crusts) an area (Shainberg and Letey, 1984; Valentin and Bresson, 1992). Crust formation, and subsequent resistance to erosion, is strongly controlled by soil texture and structure (Bedaiwy, 2007; Hu et al., 2012). For example, Farres (1978) found that soils with large numbers of small aggregates had a greater tendency to structural crust formation than those with fewer but larger aggregates. Crust formation is also affected by the amount, temporal distribution and intensity of rainfall (Fan et al., 2008; Nciizah and Wakindiki, 2014, 2015; Truman et al., 2007). Physical crust formation typically reduces SSR due to the breakdown of aggregates and although the vertical surface change may only be of the order of 1–2 mm, these microrelief dynamics can affect soil erosion processes (Croft et al., 2013; Vermang et al., 2013).

The aim of this research is to determine the impact of rainfall on the surface roughness characteristics and physical crusting of very fine soils. It differs from previous research by focusing only on small aggregates (< 1.4 mm) and therefore only considers the three smallest scales of SSR as identified by Römkens and Wang (1986). The study focuses on dryland soils with very low organic content and on how SSR changes during a multi-day rainfall event typical of a semi-arid region. The specific objectives are to quantify changes in SSR and physical crust strength in response to a multi-day rainfall event, and to investigate whether soil characteristics, such as soil texture or propensity to disaggregation, can explain these changes.

2. Materials and methods

2.1. Sample characteristics

All the soils used in this study were collected from eastern Australia and comprised individual particles and fine aggregates < 1.4 mm diameter (Table 1). Particle-size analysis of the soils was undertaken using a Beckman Coulter LS280 laser sizer in the range 0.375–1000 μm with 85 class intervals. This method of quantifying the particle-size distribution of soils can also lead to the breakdown of soil aggregates. Previous studies have exploited this and used laser sizing to quantify aggregate stability. Following the protocols of Mason et al. (2003, 2011) hydrodynamic disaggregation of the soils was quantified by circulating the samples in a water suspension for 180 min and analysing the particle-size distribution every 1 min for the first 10 min, every 5 min for the next 30 min and every 10 min thereafter. No sonication was used before or during the analysis. In a laser sizer, it is not possible to calculate zero dispersion i.e. no hydrodynamic force applied, and therefore minimal dispersion (MD), the closest measurement to that of the dry soil, is defined as the particle-size distribution measured following 1 min of circulation. The particle-size distribution measured after 180 min of circulation is considered to be mechanically fully-disaggregated and is referred to as the measure of intermediate dispersion (ID). Full disaggregation (FD) which includes sonication and chemical dispersion was not measured in this paper.

For the analyses in this paper, in addition to using the differences in particle-size distribution (clay, silt sand fractions) with minimal and intermediate dispersion as an indicator of aggregate stability, a single indicator was adopted for the purposes of statistical analysis. A number of techniques exist for summarising the aggregate stability of soils using sieve or pipette analyses (e.g. Amezketa, 1999; Le Bissonais, 1996a) but the indicator used here is that known as 'disaggregation reduction' (DR) proposed by Rawlins et al. (2013) who specifically developed it for use where particle size distribution has been determined using laser instruments. Using Rawlins et al. (2013), mean weight diameter (MWD; μm) between continuous size distributions is calculated by:

$$MWD = \sum_{i=1}^{n} \bar{x}_i w_i \tag{1}$$

where $\bar{x}_i$ is the mean diameter of each size fraction (μm), and w_i is the volume proportion (expressed as a decimal proportion) of the sample corresponding to that size fraction. Rawlins et al. (2013) focus on comparing the particle size distribution following complete disaggregation (FD) to the particle-size distribution of water stable aggregates (ID) and therefore they interpret larger values of disaggregation reduction to indicate greater hydrodynamic stability. For this paper, we use the principle of disaggregation reduction to compare MD

Table 1
Locations and characteristics of soil samples. LOI = loss on ignition; Disp. = degree of dispersion; MD = minimal dispersion; ID = intermediate dispersion; DR = disaggregation reduction. See text for details. Crusting Index (CI) is calculated for minimally-dispersed soils.

Sample	Latitude	Longitude	Salinity (μS)	LOI %	Soil type	Disp.	% Clay	% Silt	% Sand	DR (μm)	CI
Diamantina Lakes [c]	− 23.7622	140.9948	101.2	1.06	Sandy loam	MD	6.38	45.19	48.43		1.09
						ID	12.21	39.32	48.48	7.75	
Diamantina Lakes [s]	− 23.7676	140.9951	117.2	0.91	Loamy fine sand	MD	3.85	19.75	76.41		0.87
						ID	12.63	28.96	58.42	12.48	
Spoilbank	− 23.5977	143.2081	38.3	1.66	Sandy loam	MD	3.82	31.33	64.85		0.61
						ID	24.50	45.38	30.12	34.66	
Pimpara Lakes	− 30.4552	141.7019	117.3	1.04	Loamy fine sand	MD	3.42	18.33	78.25		0.76
						ID	14.68	31.07	54.25	61.90	
Eulo	− 28.1422	145.0273	24.5	1.11	Sandy loam	MD	5.18	36.75	58.08		1.17
						ID	11.47	41.85	46.68	10.43	
Thargomindah	− 27.9934	143.8204	23.38	1.01	Silt loam	MD	4.94	50.87	44.19		1.28
						ID	12.22	44.01	43.77	10.24	
Waanaaring	− 29.7006	144.1450	30.1	1.13	Sandy loam	MD	5.43	30.11	64.46		0.82
						ID	12.81	31.83	55.35	10.11	
Mallee Cliffs	− 34.4763	142.4185	139.3	3.16	Sandy loam	MD	7.06	28.65	64.29		0.51
						ID	11.60	36.39	52.01	24.89	
Tapio Station	− 34.0074	142.1507	42.3	1.66	Sandy loam	MD	7.52	26.45	66.03		0.57
						ID	12.25	27.53	60.22	2.87	
Tibooburra	− 29.4291	142.0106	60.7	1.67	Sandy loam	MD	2.91	28.35	68.74		0.68
						ID	22.64	46.89	30.47	26.62	
Lake Millyera	− 31.0363	139.9842	46,700	1.98	Loam	MD	15.71	46.24	38.05		0.21
						ID	58.59	39.83	1.58	48.76	