Aust. J. Soil Res., 1983, **21**, 47–58

Field Rainfall Simulator Studies on two Clay Soils of the Darling Downs, Queensland. II* Aggregate Breakdown, Sediment Properties and Soil Erodibility

R. J. Loch[A] *and T. E. Donnollan*[A,B]

[A] Queensland Department of Primary Industries, Queensland Wheat Research Institute, Toowoomba, Qld 4350.
[B] Present address: Queensland Department of Primary Industries, P.O. Box 591, Ayr, Qld 4807.

Abstract

Size distributions of the solids in runoff water were measured for two clay soils subjected to simulated rain under a range of plot lengths and two tillage orientations. Selective transport did not appear to have affected the sediment size distributions. Therefore, these could be used as a measure of soil structure and aggregate breakdown by rainfall and runoff.

There was little dispersed clay, most of the sediment remaining aggregated. For each soil, sediment size distributions were bimodal, peaks in sediment size being related to orders of aggregation in each soil. Concentrations of dispersed clay provide evidence that stresses on aggregates moved by rain impact on flowing water were greater than on those moved in rills by flowing water alone. Consistent with this, sediment size distributions showed much less breakdown to sizes $<0 \cdot 125$ mm in rills.

Suspended load (sediment <20 μm) showed little temporal fluctuation, and little or no decrease with time, suggesting that for these soils, aggregate disruption by raindrops and overland flow provides a continuous source of suspendable material. Bed-load was more variable and saltating and contact load appeared to be complementary to some extent.

Large differences between the two soils in measured sediment concentrations could not be explained by slight differences in sediment size. However, large differences between the soils in the water content and density of saturated aggregates were found. Transport equations for bed-load sediment suggest that the measured difference in aggregate density is sufficient to explain the difference between the soils in rates of sediment transport.

Introduction

In a study of soil erosion over a range of plot lengths, Loch and Donnollan (1983) showed that:

(i) sediment was transported by two distinct erosion processes (rain-flow and rilling);

(ii) for both processes, concentrations of bed-load sediment (and therefore rates of bed-load transport) were limited by the transport capacity of overland flow rather than the supply of detached material for transport.

Most of the sediment was bed-load, so it is likely that rates of sediment transport were largely controlled by sediment size and density.

The importance of sediment properties is acknowledged by equations developed to predict soil erodibility (e.g. Wischmeier *et al.* 1971; El-Swaify and Dangler 1976; Romkens *et al.* 1977). These equations all include parameters correlated with aggre-

* Part I, *Aust. J. Soil Res.*, 1983, **21**, 33.

0004–9573/83/010047$2.00

gate stability. Although important for surface stability and infiltration, these measures of stability, together with a textural parameter, also take account of sediment size.

Despite this, there are few data on the properties of sediment eroded from agricultural soils (Meyer *et al.* 1980; Young 1980). Barnett *et al.* (1978) found that size distributions of sediment helped explain deviations in measured K-factors from those predicted by nomograph. Studies of sediment properties may therefore help identify soil properties that are correlated with soil erodibility.

Most of the arable soils of Queensland and northern New South Wales have higher clay contents and more active clay minerals than those for which equations to predict soil erodibility have been developed. Hence erodibility cannot be predicted confidently from soil properties for these soils by existing empirical methods, and there is a need to identify soil properties that determine rates of sediment transport by raindrops and overland flow on these clay soils. This paper, based on the field experiments reported in Part I:

(i) discusses aggregate breakdown by raindrop impact and overland flow;

(ii) considers effects of erosion processes on the sizes of sediment eroded from two clay soils;

(iii) examines the relationship between sediment properties and rates of sediment transport.

Materials and Methods

Properties of the Middle Ridge clay loam and Irving clay soils used, together with details of the rainulator and field experimentation, were given in Part I (Loch and Donnollan 1983).

Sediment samples were wet sieved for 15 min, using sieve appertures of 5, 2, 1, 0·5, 0·25 and 0·125 mm, in a modified Yoder wet sieving apparatus (Coughlan *et al.* 1973). Sediment $<20\ \mu m$ and $<2\ \mu m$ was measured by pipette sampling. Material 0·125–0·02 mm was estimated by difference. Wet sieving was carried out within 24 h of sampling for the Irving clay, as the finer particles tended to bond the rest of the sediment together if allowed to settle.

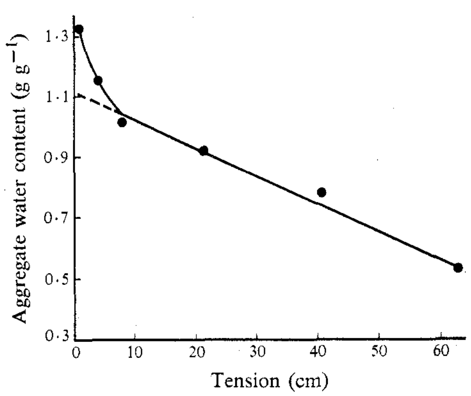

Fig. 1. Effect of tension on water content of 1–0·5 mm water-stable aggregates of an Irving clay.

The measurement of sediment density on the basis of settling velocity presented problems both in handling saturated aggregates and in the accurate definition of aggregate diameter. Instead, sediment densities were calculated from the water contents of saturated aggregates, assuming zero air content at saturation and bulk densities for soil solids and water of 2·66 and $1 \cdot 0\ g\ cm^{-3}$ respectively.

Aggregate water content was found to increase linearly with decreasing tension until, at very low tensions, the retention of significant quantities of water in pores between aggregates caused a rapid increase in water content, e.g. at tensions <8 cm (Fig. 1). To correct for this additional water between aggregates, it was assumed that aggrgate water content continued to increase linearly with decreasing tension. The extrapolation of this linear response is also shown in Fig. 1. Water-stable aggregate size ranges used for water content determinations were 2–1 mm for the Middle Ridge clay loam and 1–0·5 mm for the Irving clay.

Results and Discussion

In Part I, mean sediment concentrations were reported for a 5 min period taken immediately after runoff rate had begun to stabilize. Mean sediment size distributions over the same 5-min periods, for most of the plots reported in Part I, are given in Tables 1 and 2, for Middle Ridge clay loam and Irving clay respectively. Data for the simulated 225 m long plots are also presented, although in this context they could be considered misleading, as the actual distance of travel was only 22·5 m. Greater comminution of sediment may have occurred if sediment had actually travelled 225 m, though as noted later, breakdown during rill transport was not as great as in rain-flow. For the purposes of this paper, these plots would be better considered similar to 22·5 m long plots tilled across the slope.

Table 1. Mean sediment size distributions for a Middle Ridge clay loam, for a range of plot lengths and two tillage methods

Data taken from the 5 min period used for sediment concentration measurements

Plot length (m) and tillage orientation	Dominant erosion process	Plot discharge (l s^{-1})	Sediment concn (%)	% of sediment in each size range. Size range (mm): >5	5–2	2–1	1–0·5	0·5–0·25	0·25–0·125	0·125–0·02	0·02–0·002	<0·002
3^{A}	rain-flow	0·26	1·0	2	4	12	9	7	5	17	40	4
3^{B}	rain-flow	0·26	1·1	2	8	13	9	5	4	23	33	3
9^{A}	rain-flow	0·27	1·4	1	7	20	15	10	5	17	21	4
9^{B}	rain-flow	0·30	1·15	1	5	16	17	10	6	19	21	6
16^{A}	rill	1·10	4·1	1	8	14	13	8	6	29	19	1
16^{B}	rill	1·23	6·6	1	9	16	19	15	8	16	14	1
22·5^{A}	rill	1·30	5·4	1	9	24	22	12	5	14	12	1
22·5^{B}	rill	1·48	3·7	2	10	15	15	11	10	18	15	1
225^{A,C}	rill	12·3	4·5	2	16	19	13	7	5	28	9	1
225^{A,C}	rill	8·3	4·2	0	17	22	19	10	4	19	8	1

^{A} Tilled up and down the slope.
^{B} Tilled across slope.
^{C} Simulated plot length in terms of discharge. Actual plot length, 22·5 m.

Differences in sediment concentration associated with different erosion processes were discussed in Part I. Sediment concentrations, plot discharges, and the observed dominant erosion process for each plot are also given in Tables 1 and 2.

The lack of evidence for selective transport

Sediment size distributions for material leaving a site could be a function of either aggregate size, selective transport of various size fractions, or both. If selective transport occurs, initially high rates of transport of easily transported material could be expected to decline as other size fractions become preferentially concentrated on the plot surface and begin to dominate the soil surface–overland flow interface. However, for plots tilled up and down the slope there was little or no overall decline in any size fraction >20 μm relative to the others (Fig. 2). For plots tilled across slope, as Figs 3 and 4 show, the initial surge of runoff from depression storage caused large fluctuations in sediment concentrations. However, Fig. 3 shows no evidence of an initial flush of fine material, which is consistent with limited data from the initial surge of runoff from other plots. Once runoff stabilized, usually after 2–3 min, there

Table 2. Mean sediment size distributions for an Irving clay, for a range of plot lengths and two tillage methods

Data taken from the 5-min periods used for sediment concentration measurements

Plot length (m) and tillage orientation	Dominant erosion process	Plot discharge (l s^{-1})	Sediment concn (%)	% of sediment in each size range Size range (mm): >5	5–2	2–1	1–0·5	0·5–0·25	0·25–0·125	0·125–0·02	0·02–0·002	<0·002
1·5^{A}	rain-flow	0·09	2·49	0	1	6	22	28	10	4	22	7
1·5^{B}	rain-flow	0·10	1·93	0	2	5	12	19	10	19	22	11
3^{A}	rain-flow	0·25	1·57	0	5	13	13	6	3	8	32	20
3^{B}	rain-flow	0·25	2·5	0	3	8	13	10	4	27	20	15
6^{A}	rain-flow	0·45	2·5	0	1	8	27	17	3	14	20	10
6^{B}	rill	0·49	8·14	1	4	12	25	25	18	16	11	7
9^{A}	rill	0·68	4·58	0	0	3	33	25	10	7	14	9
9^{B}	rill	0·68	5·63	0	1	8	32	20	9	15	9	6
22·5^{A}	rill	1·35	8·06	0	1	9	40	17	6	12	9	6
22·5^{B}	rill	1·25	8·67	3	10	15	19	12	7	18	10	6
225^{A,C}	rill	17·9	8·36	1	4	9	16	13	4	33	11	9
225^{A,C}	rill	17·5	8·08	2	7	20	27	12	6	8	12	6

^{A} Tilled up and down the slope.
^{B} Tilled across the slope.
^{C} Simulated plot length—see Table 1.

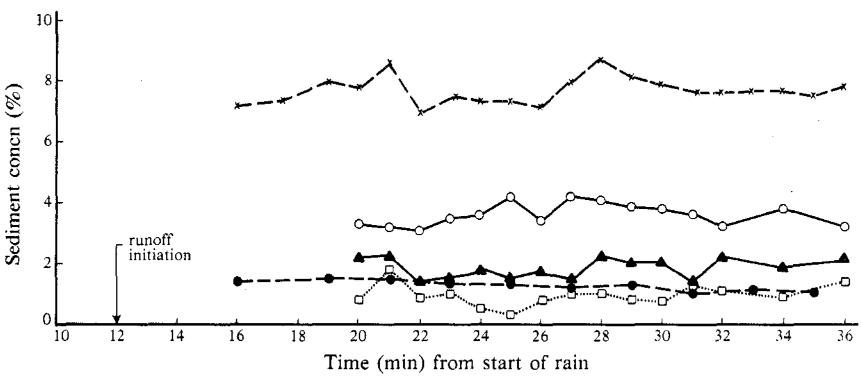

Fig. 2. Changes with time in concentrations of total sediment and several sediment size fractions from a 22·5 m long plot, tilled up and down the slope on an Irving clay. ×- - - -× Total sediment; ○——○ sediment >0·5 mm; ▲——▲ sediment 0·5–0·125 mm; □······□ sediment 0·125–0·02 mm; ●- - - -● sediment <0·02 mm.

was little apparent decline in any size fraction >20 μm relative to the others, though short-term fluctuations were common. This indicates that selective transport had little influence on the sediment size distributions shown in Tables 1 and 2. These can therefore be considered to reflect soil structure and aggregate breakdown associated with detachment and transport.

Sediment size

Relationship with Soil Properties

Despite clay contents of 53% and 73% for Middle Ridge clay loam and Irving clay respectively, there was little sediment <2 μm. Only 1–6% of the total sediment

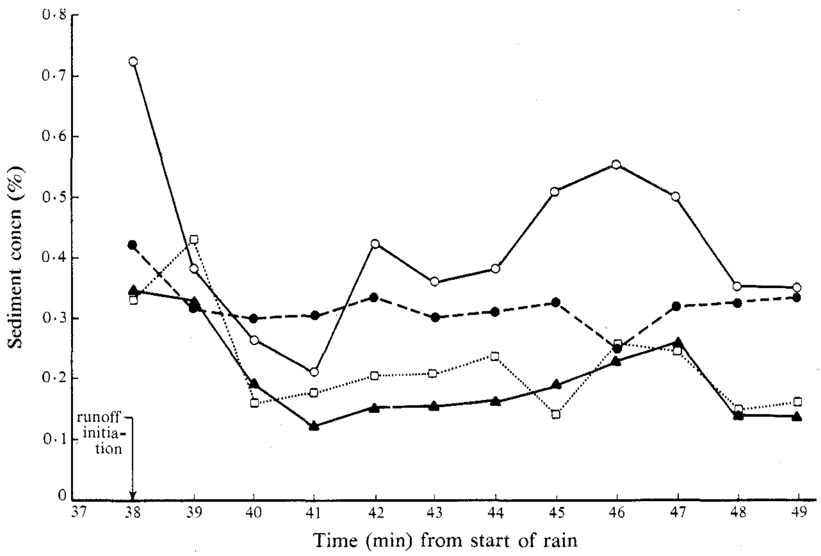

Fig. 3. Changes with time in concentrations of several sediment size fractions from a 9 m long plot, tilled across slope on a Middle Ridge clay loam. ○——○ Sedimen >0·5 mm; ▲——▲ sediment 0·5–0·125 mm; □······□ sediment 0·125–0·02 mm; ●– – – –● sediment <0·02 mm.

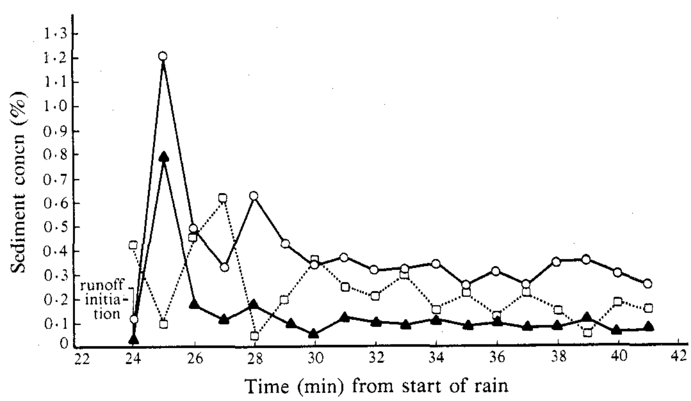

Fig. 4. Changes with time in concentrations of several sediment size fractions from a 3 m long plot, tilled across slope, on a Middle Ridge clay loam. ○——○ sediment >0·5 mm, ▲——▲ sediment 0·5–0·125 mm, □······□ sediment 0·125–0·02 mm.

from the Middle Ridge clay loam and 6–20% (generally 6–10%) of the total sediment from the Irving clay was <2 μm, indicating that relatively little clay dispersion occurred and that most of the sediment was in the form of aggregates. This is consistent with field observations of deposited sediment from similar soils.

For each soil, sediment size distributions as shown appear bimodal, e.g. with peaks at 2–0·5 mm and 0·02–0·002 mm for rain-flow eroded plots (3 m and 9 m long) on the Middle Ridge clay loam (Table 1) and a minimum at 0·25–0·125 mm. For the Middle Ridge clay loam, and to a lesser extent the Irving clay, the peak in the smaller size-range moved from 0·02–0·002 mm to 0·125–0·02 mm as rilling developed, and the proportion of sediment <0·002 mm decreased. This can be related to changes in transport process and resultant severity of breakdown which are discussed later.

For each soil the peak at the larger size occurred over a consistent size range. For the Middle Ridge clay loam this was 2–0·5 mm, while for the Irving clay it was 1–0·5 mm with some apparant overlap into the 0·5–0·25 mm size range. These sediment size distributions are interpreted as showing, for each soil, one peak at a preferred or 'natural' aggregate size and another at a smaller size representing material from aggregates disrupted by drop impact, transport by overland flow, or tillage. However, tillage was not a major factor in breakdown to <0·125 mm. Dry aggregate size distributions for both soils prior to rainfall typically contained <5% of dry aggregates <0·5 mm.

For both soils, the apparent minimum in sediment at an intermediate size range, usually 0·25–0·125 mm, suggests two levels of aggregation, with the larger peak sediment size for each soil representing a relatively stable agglomerate of aggregates of the smaller peak size. It could be expected that agglomerates would break down to their constituent aggregates, and produce bimodal distributions of sediment sizes, since bonds between aggregates in an agglomerate would be weaker than those within the aggregates. Also, stresses imposed upon an agglomerate will be concentrated in cracks and pores (Lowrison 1974), tending to split an agglomerate into its component aggregates.

Table 3. Percentage by weight of dry aggregate size fractions of an Irving clay that was quartz >20 μm

Size fraction (mm)	10–5	5–2	2–1	1–0·5	0·5–0·25	whole soil
% quartz >20 μm	14·1	14·2	12·7	14·6	25·3	11

Other evidence of different levels of aggregation in similar soils agrees well with the peak size ranges shown in Tables 1 and 2. For a swelling clay soil similar to the Irving clay, Coughlan (1974) identified natural aggregates 0·5–0·2 mm, stable to field wetting and drying, which were agglomerates of aggregates of approximately 0·02 mm diameter. Coughlan *et al.* (1978*a*) used several measurements to estimate the size of natural water-stable agglomerates in cracking clay soils, including the percentage by weight of each size fraction that was quartz >0·02 mm. Table 3 shows an increase in quartz >0·02 mm in the 1–0·05 mm dry aggregate size range for the Irving clay, which is evidence of a natural aggregate size of ⩽1 mm. This agrees well with the observed peak in the 1–0·5 mm size range for sediment from the Irving clay.

For three krasnozem soils, similar to the Middle Ridge clay loam, Coughlan *et al.* (1973) identified a 0·5–0·1 mm size fraction of dry aggregates which they suggested represented a basic level of water stable aggregation. They also suggested that aggregates in the size range 5–0·5 mm were stabilized by organic matter, and hence it is possible that the 2–0·5 mm aggregates observed for the Middle Ridge clay loam were agglomerates of 0·125–0·02 mm aggregates, held together by organic bonds.

Meyer *et al.* (1980) reported size distributions of sediment from 10 Mississippi soils, of which several (with higher clay contents and consequently stronger aggregates (Coughlan *et al.* (1978*b*)) also produced bimodal sediment size distributions. Generally, however, in these soils the smaller peak size represented primary particles, not aggregates. Young (1980) also noted that many clay soils are well aggregated and tend to erode as aggregates, suggesting that the sediment size distributions measured in this study are typical of many clay soils.

Clay Dispersion and Stresses on Aggregates during Transport

For soils such as these that do not disperse spontaneously in water, clay dispersion requires external work to be performed. Consequently concentrations of dispersed clay provide an indication of the forces that acted on aggregates during detachment and transport by rain-flow and rilling.

For the Middle Ridge clay loam the concentration of sediment $<2\ \mu$m (calculated from total sediment concentrations and the percentage sediment $<2\ \mu$m for each plot (Table 1)) was the same for plots whether rain-flow or rill transport was dominant —an average of 0·05% in each case. This means the concentration of dispersed clay in overland flow reaching rills did not increase during subsequent transport in the rill. This implies that clay dispersion did not occur in the rills, but only during rain-flow transport to them. This is evidence that stresses on aggregates are greater during rain-flow than rill transport.

The greater stresses on aggregates carried by rain-flow can be attributed largely to raindrop impact, occurring either on bare soil or soil covered by shallow overland flow. Sediment may also have been splashed into the air a number of times before reaching areas of flow and being transported off the plots. In contrast, flow in rills was commonly 20–30 mm deep, and much of the impact of the drops (2·1 mm median diameter) would have been absorbed by the water. Therefore collisions between entrained aggregates and with the rill bed would have been the major source of stress on aggregates transported in rills. From comparison of raindrop velocities of approximately 5 m s^{-1} (Meyer 1958) and velocities of aggregates in rill flow ($<0\cdot3$ m s^{-1}) it appears that such stresses would be small relative to raindrop impacts. The two processes also differ in their sources of sediment. Rain-flow transport was confined to the soil surface, while rills detached soil from depths of up to 10 cm which had not been exposed to raindrop impact.

Unlike the Middle Ridge clay loam, the Irving clay showed a significant increase ($P < 0\cdot01$) in dispersed clay during rill transport. Average concentrations of sediment $<2\ \mu$m were 0·26 and 0·51% for plots dominated by rain-flow or rill transport respectively. The small stresses on aggregates during rill transport were still sufficient to cause dispersion of this soil, suggesting that the critical stress required to disperse clay is lower for the Irving clay than for the Middle Ridge clay loam. This may be related to differences in clay mineralogy, with the swelling Irving clay having a higher water content on saturation, and consequently lower cohesion. The presence of large quantities of sesquioxides may also have contributed to the greater clay stability of the Middle Ridge clay loam.

Aggregate Breakdown—Interactions with Erosion Processes

For both soils the size distributions of sediment detached and transported by rills were estimated by comparing sediment size data for 3 and 22·5 m long plots tilled across the slope. For the 22·5 m long plots:

$$\text{total sed. concn} = \text{rain-flow sed. concn} + \text{rill sed. concn.}$$

The sediment concentrations from 3 m long plots were used as an estimate of rain-flow inputs to rills on the 22·5 m long plots. This was considered realistic as all plots were 4 m wide, and the rills were well off-centre on the 22·5 m long plots so that rain-flow transport to rills usually required movement over approximately 3 m distance. Therefore, for each sediment size fraction:

$$\text{rill concn} = \text{total concn} - \text{rain-flow (3 m plot) concn.}$$

Table 4 shows the sediment in each size fraction estimated to have been detached by rilling as a percentage of the total sediment in that size fraction eroded from the 22·5 m long plots tilled across slope. Estimated rill contributions to total sediment loads from the 22·5 m long plots were 70 and 71% respectively for Middle Ridge clay loam and Irving clay. Table 4 shows that, for both soils, rilling contributed more

Table 4. Estimated rill contributions to each sediment size range for a Middle Ridge clay loam and an Irving clay

Derived from data for 3 m and 22·5 m long plots ploughed across slope

Soil	Estimated % of sediment in each size fraction that was detached and transported by rills only								
	Size fraction (mm): >5	5–2	2–1	1–0·5	0·5–0·25	0·25–0·125	0·125–0·02	0·02–0·002	<0·002
Middle Ridge clay loam	78	76	74	82	88	88	63	34	0
Irving clay	100	91	85	80	76	84	57	45	29

than 70% of the sediment in size fractions >0·125 mm, and increasingly less than 70% for smaller size fractions. This means that there was much less aggregate breakdown to <0·125 mm in rills than in rain-flow, which explains the shift in the smaller peak size from 0·02–0·002 mm to 0·125–0·02 mm for both soils as the dominant erosion process changes from rain-flow to rilling. It is also consistent with the conclusion from clay dispersion data that stresses on aggregates were greater during transport by rain-flow than during transport in rills.

Transport mechanisms

As noted by Moss *et al.* (1979), sediment transport can be divided into suspended, saltating and contact (rolling) loads, each normally being broadly associated with particular sediment size ranges. By examining short-term changes in sediment size distribution and identifying size ranges moved by 'fast' and 'slow' transport mechanisms, it is possible to define the approximate size fractions moved as suspended, saltating, or contact loads. Cummings (1981) suggested the following:

<31 μm, suspended load; 31–211 μm, saltating bed load; >211 μm, contact bed load.

Our size analysis did not partition at 31 μm, so 20 μm was adopted as an approximate boundary between suspended and bed load. A transition from saltating to contact load in the size range 250–125 μm is indicated by the limited data available from this study. Sediment was generally not sampled early in the runs, but for a 3 m long plot on the Middle Ridge clay loam, the initial breakthrough of surface detention, associated with runoff initiation, provided an event where large quantities of sediment began moving at the same time. Fig. 4 shows that peaks of sediment 5–0·5 and 0·5–0·125 mm reached the collection guttering at 25 min, while a peak of sediment 0·125–0·02 mm took a further 2 min to reach the guttering. The slower movement of sediment 0·125–0·02 mm (shown by a dotted line in Figs 2–4) is consistent with movement by saltation (Moss *et al.* 1979).

On longer plots with higher discharges, or possibly on the more erodible Irving clay, the transition from saltating to contact load may have occurred at a sediment size larger than 0·125 mm. However, for the Middle Ridge clay loam there is relatively little sediment in the 0·5–0·125 mm size range (Table 1). For the Irving clay there is relatively little sediment 0·25–0·125 mm, with the 0·5–0·25 mm size fraction becoming less important as plot size increases (Table 2). This means that errors in defining the transition from saltating to contact load have little effect on the estimates of the relative amounts of sediment in each.

Bed-load

Concentrations of individual bed-load size fractions often showed short-term fluctuations (Figs 2–4). This can be attributed to both the intermittent nature of some sediment inputs (e.g. rill bank collapse) and the intermittent nature of the movement involved. Moss *et al.* (1979) also found considerable temporal variation in bed-load transport rates. In Figs 2–4 contact and saltating load appear to be complementary to some extent. This may reflect fluctuations in availability of different size fractions, or may be simply due to the occasional large inputs of sediment, of which different size fractions are transported at different speeds. Moss *et al.* (1979) also noted that generally the sum of saltating and contact loads varied less than either individually.

Suspended Load

Sediment <20 μm shows less temporal fluctuation than do larger, bed-load size fractions (Figs 2 and 3). This is consistent with most of the sediment <20 μm being produced by raindrop impact, which is constant through time. While Moss *et al.* (1979) reported a rapid decline in suspended load from an alluvial sand due to rapid transport and depletion of suspendable material, Figs 2 and 3 show little or no decrease in suspended load with time. This is evidence that, on these clay soils, aggregate breakdown by drop impact and overland flow represents a continuous source of suspendable material, with transport of suspended laod, in direct contrast to bed-load, being *limited by the rate at which it is being supplied.*

On more weakly aggregated soils a decline in suspended load with time could be expected. Disruption of surface aggregates by rainfall, and subsequent rapid removal of clay leaves a surface visibility enriched by less-transportable sand grains. Thus, with time, the soil–overland flow interface can become dominated by slowly moved sand grains, reducing both suspended and total loads. Some of the soils studied by

Cummings (1981) show a marked decline in suspended load with time. Other evidence comes from a rainfall simulator study of Young and Onstad (1978), who reported clay contents of interrill sediment that were lower than in the original soils. As some of their runs were quite long (up to 165 min) it is quite likely that their sampling at 'equilibrium condition' occurred well after the initial removal of clay from surface aggregates.

Suspended load may provide a useful estimate of delivery ratios in the field, i.e. a measure of the diminution of eroded sediments by deposition as they move from the point of erosion to any designated downstream location. Land slope (up to 8%) is greater than the gradient of contour bank channels (0·3%). Consequently, most of the sediment reaching contour banks is deposited, and only a small proportion of the sediment is discharged from the contour bank channels into waterways or gullies. For contour bays on an Irving clay of 6–7% slope described by Freebairn and Boughton (1981), an average of 14% of the sediment reaching the contour banks was discharged into a waterway (Freebairn, personal communication). This agrees well with estimates of suspended load for the Irving clay (Table 2) as 15–23% of total sediment from rill-eroded plots.

Soil erodibility (transportability) and sediment properties

Sediment concentrations (Tables 1 and 2) show a large difference between the two soils, mean concentrations produced by rilling being 4·8 and 8·3% for Middle Ridge clay loam and Irving clay respectively.

Table 5. Mean concentrations in percentage of suspended load and bed-load[A] for plots eroded by rain-flow or by rain-flow plus rilling on a Middle Ridge clay loam and an Irving clay

Erosion process and size fractions	Middle Ridge clay loam	Irving clay
Rain-flow only, suspended load	0·37	0·76
Rain-flow only, bed-load	0·79	1·44
Rain-flow plus rilling, suspended load	0·66	1·31
Rain-flow plus rilling, bed-load	3·79	6·82

[A] Atypical plots excluded from bed-load estimations as detailed in Part I.

The two soils differ in the actual concentration of suspended load (Table 5), with the less stable Irving producing more sediment $<20\ \mu m$. However, suspended load explains only a small proportion of the difference between these soils, which is largely due to differences in rates of bed-load transport. For both rain-flow and rilling, the Irving clay produces bed-load concentrations approximately 1·8 times those of the Middle Ridge clay loam. There is good evidence (discussed in Part I) that bed-load transport was controlled by transport capacity, not detachment. Therefore, differences between the soils in rates of bed-load transport are likely to reflect differences in the 'transportability' of the sediment produced.

Bed-load sediment from the Middle Ridge clay loam was generally slightly coarser than that from the Irving clay, which is consistent with the differences in bed-load sediment concentration shown in Table 5. However, it is possible to find similar plots, e.g. the 22·5 m plots tilled across slope on both Middle Ridge clay loam and

Irving clay, where sediment size distributions are virtually identical, yet sediment concentrations show the same differences. Therefore, it seems unlikely that slight differences in sediment size can explain the large differences in sediment concentrations between these soils.

As bed-load transport is influenced by both sediment size and density, densities of saturated aggregates were calculated for the larger peak size fraction for each soil (Table 6). These differ considerably between the non-swelling Middle Ridge clay loam and the swelling Irving clay. Graf (1971) described bed-load transport equations proposed by Schoklitsh in 1914 and Shields in 1936. For similar size particles, with the densities shown in Table 6, these two equations predict that bed-load transport rates of the Irving clay would be respectively 1·8 and 3·2 times those of the Middle Ridge clay loam. The relatively poor agreement between these bed-load transport equations is not unusual. However, both agree reasonably well with measured differences in bed-load transport between the two soils (Table 5). This adds further credence to the suggestion that differences between the two soils in rates of bed-load transport are largely due to differences in saturated aggregated (sediment) density.

Table 6. Gravimetric water contents and wet densities of saturated aggregates from a Middle Ridge clay loam and an Irving clay

Soil	Gravimetric water content (%)	Wet density ($g\ cm^{-3}$)
Middle Ridge clay loam	45	1·76
Irving clay	110	1·42

More rigorous application of sediment transport equations does not seem warranted until data are available for a wider range of soils.

Conclusions

One important conclusion from this paper is that sediment size distributions, in at least some instances, can be used as a measure of aggregate breakdown under rainfall. In fact this may be a more valid measure of such breakdown than most traditional tests of aggregate stability. The level of breakdown was concluded to be a function of the magnitude of the stresses acting on aggregates, and raindrop impact appeared to be the major source of large stresses.

Total soil loss is the sum of suspended, saltating and contact loads. The data show that each of these loads is supplied (detached) and transported at different rates and by different mechanisms. Consequently total soil loss has little physical meaning, and the partition of soil loss into these more meaningful components appears to be essential both for initial data interpretation and for subsequent use of such data for soil loss prediction.

Sediment density is often considered to have little effect on soil erodibility (Meyer and Harmon 1979). However, it appears that, at least for swelling clays, saturated aggregate (sediment) density may be a major factor governing soil erodibility. Similar large differences in density (and therefore, in erodibility) may be found between soils which erode as either aggregates or ultimate particles.

Acknowledgments

We would like to thank Dr A. J. Moss and Dr K. J. Coughlan for their advice during the preparation of this paper.

References

Barnett, A. P., Dooley, A. E., and Smith, G. A. (1978). Soil erosion and sediment movement under sugarcane culture on the flat lands of southern Louisiana. *Trans. Am. Soc. Agric. Eng.* **21**, 1144–50.

Coughlan, K. J. (1974). A study of the aggregation process in a swelling clay. Ph.D. Thesis, Univ. Qld.

Coughlan, K. J., Fox, W. E., and Hughes, J. D. (1973). A study of the mechanisms of aggregation in a krasnozem soil. *Aust. J. Soil Res.* **11**, 65–73.

Coughlan, K. J., Smith, G. D., and Fox, W. E. (1978*a*). Measurement of the size of natural aggregates. In 'Modification of Soil Structure', ed. W. E. Emerson, R. D. Bond, and A. R. Dexter. pp. 205–10. (John Wiley and Sons: New York.)

Coughlan, K. J., Loch, R. J., and Fox, W. E. (1978*b*). Binary packing theory and the physical properties of aggregates. *Aust. J. Soil Res.* **16**, 283–9.

Cummings, D. (1981). Soil erodibility determinations from laboratory rainfall simulation. M. Agric. Sci. Thesis, Univ. Melb.

El-Swaify, S. A., and Dangler, E. W. (1976). Erodibilities of selected tropical soils in relation to structural and hydrological parameters. *In* 'Soil Erosion: Prediction and Control'. pp. 105–14. (Soil Cons. Soc. Am.: Ankeny, Iowa.)

Freebairn, D. M., and Boughton, W. C. (1981). Surface runoff experiments on the eastern Darling Downs. *Aust. J. Soil Res.* **19**, 133–46.

Graf, W. H. (1971). 'Hydraulics of Sediment Transport.' (McGraw–Hill: New York.)

Loch, R. J., and Donnollan, T. E. (1983). Field rainfall simulator studies on two clay soils of the Darling Downs, Queensland. I. The effects of plot length and tillage orientation on erosion processes and runoff and erosion rates. *Aust. J. Soil Res.* **21**, 33–46.

Lowrison, G. C. (1974). 'Crushing and Grinding. The Size Reduction of Solid Materials.' (Butterworths: London.)

Meyer, L. D. (1958). An investigation of methods for simulating rainfall on standard runoff plots and a study of the drop size, velocity adn kinetic energy of selected spray nozzles. Spec. Rep. No. 81, Soil Water Conserv. Res. Div., Agric. Res. Serv., U.S. Dep. Agric.

Meyer, L. D., and Harmon, W. C. (1979). Multiple-intensity rainfall simulator for erosion research on row sideslopes. *Trans. Am. Soc. Agric. Eng.* **22**, 100–3.

Meyer, L. D., Harmon, W. C., and McDowell, L. L. (1980). Sediment sizes eroded from crop row sideslopes. *Trans. Am. Soc. Agric. Eng.* **23**, 891–8.

Moss, A. J., Walker, P. H., and Hutka, J. (1979). Raindrop-stimulated transportation in shallow water flows: an experimental study. *Sediment Geol.* **22**, 165–84.

Romkens, M. J. M., Roth, C. B., and Nelson, D. W. (1977). Erodibility of selected clay subsoils in relation to physical and chemical properties. *Soil Sci. Soc. Am. J.* **41**, 954–60.

Wischmeier, W. H., Johnson, C. B., and Cross, B. V. (1971). A soil erodibility monograph for farmland and construction sites. *J. Soil Water Conserv.* **26**, 189–93.

Young, R. A. (1980). Characteristics of eroded sediment. *Trans. Am. Soc. Agric. Eng.* **23**, 1139–42.

Young, R. A., and Onstad, C. A. (1978). Characterization of rill and interrill eroded soil. *Trans. Am. Soc. Agric. Eng.* **21**, 1126–30.

Manuscript received 26 July 1982, accepted 9 September 1982